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CONCENTRATION OF HYDROGEN IN THE UPPER ATMOSPHERE OF THE
EARTH IN THE 300-600 KM ALTITUDE RANGE ACCORDING
TO IONOSPHERIC DATA

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by M. N. Fatkullin

SUMMARY

Utilizing the results of rocket and satellite investigations of the ionic composition of the outer part of the F2-region of the ionosphere and of measurements of atomic oxygen concentration at corresponding heights, estimates are presented of hydrogen concentration in the 300 - 600 km altitude range during solar activity minimum. Comparison is also made between the estimated values of hydrogen concentration and the results of direct measurements.

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* *

According to data of ionosphere investigations, the altitude distribution of neutral hydrogen concentration $n(h, H)$ may be estimated from measurements of the ionic composition of the medium, in particular of hydrogen ion H^+ concentrations, the latter forming and vanishing in the course of reactions of asymmetrical resonance transfer of the charge [1]



and of atomic oxygen O^+ . These estimates are based upon the equation

$$\frac{\partial n(H^+)}{\partial t} = S(H^+) - L(H^+) - \frac{\partial F(H^+)}{\partial h}, \quad (2)$$

determining the altitude-time distribution $n(H^+)$. In equation (2)

$$\begin{aligned} S(H^+) &= k(H, O^+)n(H)n(O^+), \\ L(H^+) &= k(H^+, O)n(H^+)n(O), \\ F(H^+) &= n(H^+)u(H^+) \simeq -D(H^+) \left(\frac{\partial n(H^+)}{\partial h} + \frac{n(H^+)}{H(H^+)} \right) \end{aligned} \quad (3)$$

represent respectively the source, the losses and flux of ions H^+ : k are the coefficients of reactions rates (1), $u(H^+)$ is the vertical velocity component of the directed motion of protons, $D(H^+)$ is the diffusion coefficient, H is the scale height.

In the outer part of the F_2 -region of the ionosphere ($h \approx 300 - 600$ km) the settling time of chemically equilibrium distribution of ion H^+ concentration ($S = L$), determined as $\tau_1 = 1/k(H^+, O)n(O)$, constitutes a few tens or less minutes, while the settling time of diffusive equilibrium distribution [$u(H^+) = 0$] is

$$\tau_2 \approx H^2(H^+) / D(H^+) \approx 10^3 \div 10^4 \text{ min.}$$

More detailed data on the orders of magnitudes of τ_1 and τ_2 in daytime (about 1600 hours LT) at various heights are compiled in Table 1(*):

T A B L E 1

characteristics	altitude, km				ref
	300	400	500	600	
$n(H^+)/n(O^+)$	$6,5 \cdot 10^{-3}$	$1,1 \cdot 10^{-2}$	$2,5 \cdot 10^{-2}$	$5,2 \cdot 10^{-2}$	[5]
$n(O)$, cm^{-3}	$1,0 \cdot 10^8$	$1,6 \cdot 10^7$	$2,4 \cdot 10^6$	$4,0 \cdot 10^5$	[8]
T_i , $^\circ\text{K}$	970	1100	1370	1670	[11]
T , $^\circ\text{K}$	880	880	880	880	
τ_1, min	$4,2 \cdot 10^{-1}$	$2,6 \cdot 10^0$	$1,7 \cdot 10^1$	$1,0 \cdot 10^2$	
τ_2, min	$1,7 \cdot 10^4$	$9,3 \cdot 10^3$	$4,0 \cdot 10^3$	$2,5 \cdot 10^3$	
τ_2, min	$2,5 \cdot 10^{-5}$	$6,4 \cdot 10^{-5}$	$4,2 \cdot 10^{-3}$	$3,5 \cdot 10^{-2}$	
$n(H)$, cm^{-3}	$5,8 \cdot 10^5$	$1,6 \cdot 10^5$	$5,6 \cdot 10^4$	$1,8 \cdot 10^4$	
$(T = T_i)$					
$n(H)$, cm^{-3}	$5,5 \cdot 10^5$	$1,4 \cdot 10^5$	$4,5 \cdot 10^4$	$1,3 \cdot 10^4$	
$(T \neq T_i)$					
$n(H)$, cm^{-3}	$7,60 \cdot 10^4$	$6,50 \cdot 10^4$	$5,60 \cdot 10^4$	$4,98 \cdot 10^4$	[12]
$(T = 860^\circ\text{K})$					
$n(H)$, cm^{-3}	$3,20 \cdot 10^4$	$2,82 \cdot 10^4$	$2,49 \cdot 10^4$	$2,21 \cdot 10^4$	[13]
$(T = 838^\circ\text{K})$					
$n(H)$, cm^{-3}	$2,30 \cdot 10^4$	$2,06 \cdot 10^4$	$1,86 \cdot 10^4$	$1,67 \cdot 10^4$	[13]
$(T = 990^\circ\text{K})$					
$n(H)$, cm^{-3}	—	—	$2,70 \cdot 10^4$	—	[14]
$(T = 1000^\circ\text{K})$					
$n(H)$, cm^{-3}	—	—	$7,5 \cdot 10^4$	—	[15]
$(T = 1000^\circ\text{K})$					

On the other hand, at considered heights, the term of motion in Eq.(2) is much smaller by comparison with S and L , since (see Table 1)

$$\delta_1 = \left| \frac{\partial F(H^+)/\partial h}{S(H^+)} \right| \approx \frac{D(H^+)n(H^+)}{k(H, O^+)n(O^+)n(H)H^2(H^+)} \ll 1,$$

$$\delta_2 = \left| \frac{\partial F(H^+)/\partial h}{L(H^+)} \right| \approx \frac{D(H^+)}{k(H^+, O)n(O)H^2(H^+)} \ll 1$$

(*) see this remark at the end of the paper.

Therefore, it is clear that in Eq.(2)

$$S(H^+), L(H^+) \gg \left| \frac{\partial n(H^+)}{\partial t} \right|, \left| \frac{\partial F(H^+)}{\partial h} \right|.$$

Under these conditions Eq.(2) has the following simple solutions:

$$\frac{n(H^+)}{n(O^+)} = \frac{9}{8} \frac{n(H)}{n(O)} \quad (4)$$

at equality of ion temperatures $T_i = T_i(H^+) = T_i(O^+)$, of neutral particles' T [6] and

$$\frac{n(H^+)}{n(O^+)} = \frac{9}{8} \frac{n(H)}{n(O)} \left\langle \frac{T_i(O^+)/16 + T}{T_i(H^+) + T/16} \right\rangle^{1/2} \quad (5)$$

for $T_i \neq T$ [7]. In the presence of information on the values of $n(O)$, T_i , T , $n(H^+)/n(O^+)$, we may estimate from (4) and (5) the value of $n(h, H)$. For this estimate we utilize the results of measurements of the required quantities with reference to a period close to solar activity minimum.

The measurements of $n(h, O)$ were conducted with the help of mass-spectrometers on AES "EXPLORER-17" during April and May 1963 [8]. The processing of the data related to 1100 - 2000 hours LT and the altitude range 260 - 600 km by the method of least squares allowed us to obtain for the distribution $n(h, O)$ the following expression:

$$n(h; O) = 2,44 \cdot 10^{10} \cdot \exp\left(-\frac{h}{54,4} \text{ km}\right) [\text{cm}^{-3}], \quad (6)$$

whence it is easy to find the estimate $T \approx 880^\circ\text{K}$.

Several measurements of ion composition of the outer part of the F_2 -region of the ionosphere were performed by various methods, utilizing radio-frequency mass-spectrometers on AES "ELEKTRON-2" [5] (10 to 16 February 1964, $h = 440 - 1800$ km, $t = 1400 - 1900$ hours LT), of incoherent radiowave scattering [9] (December 1964 and July 1965, $h = 450 - 1200$ km, at middle latitudes and in magnetoquiet days) and of magnetic mass-spectrometer on a rocket [10] (17 January 1964, $h = 200 - 926$ km, $t = 00$ hours LT, at middle latitudes).

The results of calculations of two variants of $n(h, H)$ distribution at about 1600 hours LT, related respectively to conditions $T = T_i$ and $T \neq T_i$, are compiled in Table 1. The indicated distribution of T_i was borrowed from [11]. The taking into account of ion and neutral temperatures affects little the distribution $n(h, H)$; for example, on $h \approx 500$ km, two distribution variants differ only by a factor of ~ 1.2 .

Comparison of ionospheric estimates of $n(h, H)$ for daytime conditions with model calculations of [12-14] shows a good agreement with those of [12]

for $h \approx 500$ km [12] (see Table 1). However, the value $n(500, H) = 7.5 \cdot 10^4$ cm^{-3} at $T = 1000^\circ\text{K}$ [15], is apparently somewhat overrated.

Let us now consider the night conditions. The data processing of [8], related to the distribution of $n(h, 0)$ in night hours (2200-0400 hours LT) in the altitude range 300 - 680 km, also applying the method of least squares, resulted in the expression

$$n(h; 0) = 5.55 \cdot 10^{10} \cdot \exp\left(-\frac{h}{47.6}\right) [\text{cm}^{-3}] \quad (7)$$

$[H(0) \approx 47.6$ km corresponds to thermosphere temperature $T \approx 760^\circ\text{K}]$.

In the near-midnight hours $n(450, 0) = 4.86 \cdot 10^6$ cm^{-3} . According to [9] at the same height $n(H^+) / n(O^+) \approx 0.6$. Under these conditions, at midnight $n(450, H) \approx 2.6 \cdot 10^6$ cm^{-3} .

The results of estimates of $n(h, H)$ at midnight according to data of [10] and the expression (7) are compiled in Table 2., where we find also the estimates of $n(h, H)$ based upon the data of [10] and other model distributions of $n(h, 0)$ [12, 13] (*) It should be noted that the distribution of $n(h, 0)$, pre-assigned by expression (7), agrees well with the models of [12 & 13] on the heights considered (the divergences do not exceed a factor of ~ 2). The estimates of $n(h, H)$ in the near-midnight hours according to the distribution of $n(H^+) / n(O^+)$ [10] lead to values which are at least by two orders greater than the model ones [12-15].

T A B L E 2

characteristics	altitude. km			ref.
	400	450	500	
$n(H^+)/n(O^+)$	1,0	3,3	10,0	[10]
$n(O), \text{cm}^{-3}$	$1,25 \cdot 10^7$	$4,85 \cdot 10^6$	$1,52 \cdot 10^6$	[8]
$n(H), \text{cm}^{-3}$	$1,11 \cdot 10^7$	$1,40 \cdot 10^7$	$1,35 \cdot 10^7$	
$(T = T_i)$				
$n(O), \text{cm}^{-3}$	$1,91 \cdot 10^7$	$5,57 \cdot 10^6$	$1,66 \cdot 10^6$	[12]
$n(H), \text{cm}^{-3}$	$1,70 \cdot 10^7$	$1,63 \cdot 10^7$	$1,48 \cdot 10^7$	
$n(H), \text{cm}^{-3}$	$4,10 \cdot 10^5$	$3,75 \cdot 10^5$	$3,50 \cdot 10^5$	[12]
$(T = 670^\circ\text{K})$				
$n(O), \text{cm}^{-3}$	$2,64 \cdot 10^7$	—	$3,71 \cdot 10^6$	[13]
$n(H), \text{cm}^{-3}$	$2,35 \cdot 10^7$	—	$3,30 \cdot 10^7$	
$n(H), \text{cm}^{-3}$	$2,82 \cdot 10^4$	—	$2,49 \cdot 10^4$	[13]
$(T = 840^\circ\text{K})$				
$n(H), \text{cm}^{-3}$	—	—	$2,08 \cdot 10^5$	[14]
$(T = 750^\circ\text{K})$				
$n(H), \text{cm}^{-3}$	—	—	$\sim 1,5 \cdot 10^5$	[15]
$(T = 700^\circ\text{K})$				

The experimental data on hydrogen concentration on the height considered or close to them are obtained from the observations of L_α -emission and measurements of atmosphere density. A review of data available on the distribution of $n(h, H)$ can be found in the papers [15, 16].

(*) and the results of model calculations [12-15]

The determination of hydrogen concentration by atmosphere density leads to clearly overrated values. Thus, for example, it was obtained according to the density measured by sensors installed onboard "Explorer-17" [17] that in April-June 1963 at 900 km altitude, $n(H) \approx 10^7 \text{ cm}^{-3}$.

Mass-spectrometric observations of atmosphere composition were lately conducted on "Explorer-32" in the $h \approx 290-1000 \text{ km}$ altitude range [18]. According to this report the typical value $n(H) = 2 \cdot 10^6 \text{ cm}^{-3}$ was registered at 0900 hours local time at an altitude $h \approx 300 \text{ km}$. If we combined the data of [19] with the results of our own estimates, we would obtain for $h \approx 300 \text{ km}$ a daily course for $n(H)$ ($t = 2200-0400 \text{ h.LT. } n(H) \approx 10^7 \text{ cm}^{-3}$; $t = 0900 \text{ h, } n(H) = 2 \cdot 10^6 \text{ cm}^{-3}$; $t = 1100-2000 \text{ h. LT, } n(H) \approx 5 \cdot 10^5 \text{ cm}^{-3}$), qualitatively agreeing with the daily variations of the temperature of the thermosphere.

***** T H E E N D *****

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(*) NOTE from page 3 (intrapaginal)

When estimating the values of τ_1, τ_2, δ_2 the diffusion coefficients of H^+ ions in a gas of ions O^+ and through a neutral oxygenous atmosphere was respectively determined as [2] $D(H^+, O^+) \approx 9.3 \cdot 10^7 T_i^{3/2} / n(O^+)$ and $D(H^+, O) \approx 2.1 \cdot 10^{18} T_i^{1/2} / n(O)$, whereupon $D(H^+) = D(H^+, O^+) D(H^+, O) / [D(H^+, O^+) + D(H^+, O)]$. According to ionospheric estimates $k(H^+, O) \approx 4.0 \cdot 10^{-10} \text{ cm}^3 \cdot \text{sec}^{-1}$ [3]. According to data of laboratory measurements [4] at $T_i = T_i(O^+) = T_i(H^+) \approx 1700^\circ \text{K}$, $k(H^+, O) \approx 7 \cdot 10^{-10} \text{ cm}^3 \cdot \text{sec}^{-1}$. The first value of $k(H^+, O)$ was admitted. The distribution of $n(h, O^+)$ was computed by the relative concentration of ions O^+ [5] and $N(h)$ -profile at middle latitudes for 10 May 1963 at 1632 h. LT, obtained during external sounding of the ionosphere from AES "Alouette". Expression (6) was taken for the distribution of $n(h, O)$.

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