cles were collected above the canopy by filtration and with cascade impactors. We extracted and analyzed for the water- and dilute acid-soluble (0.1*N* Ultrex HNO_3) fraction of each metal by using flameless atomic absorption spectroscopy; insoluble material is not readily available for uptake by plants (*7*).
4. S. B. McLaughlin, D. S. Shriner, R. K. McConathy, L. K. Mann, *Environ. Exp. Bot.* **19**, 179 (1979).
5. S. E. Lindberg and R. C. Harriss, *Water Air Soil Pollut.* **16**, 13 (1981).
6. These values were estimated from the measured concentrations of distilled water–soluble metals that had accumulated on upper-canopy surfaces during the 2.4-day dry period preceding the rain event on 18 May. The metal concentrations result from the dissolution of this material in the rain, the initial composition of which was measured; the solution was then concentrated to 1/10 of its earlier volume by evaporation. Factors influencing these calculations are as follows: (i) rainfall *p*H; as *p*H decreases, dry deposition solubility increases; (ii) moisture contact time with the leaf; although heavier rains result in runoff, brief drizzle and captured fog droplets may remain on the leaf surface for 2 to 5 hours prior to evaporation; (iii) degree of concentration by evaporation; any value could be used because ultimately the solution volume reaches zero; and (iv) deposited-particle distribution on the leaf surface; as suggested in Fig. 1, considerably higher (or lower) concentrations than the surface average value determined here can occur on smaller areas of the leaf.
7. G. H. M. Krause and H. Kaiser, *Environ. Pollut.* **12**, 63 (1977).
8. R. Guderian, in *Air Pollutant Effects on the Terrestrial Ecosystem*, S. V. Krupa and A. H. Legge, Eds. (Wiley, New York, in press).
9. A. C. Chamberlain and R. C. Chadwick, *Nucleonics* **8**, 22 (1953).
10. We sampled suspended particles above the forest canopy (*3*) during four periods simultaneously with the exposure of flat plates in the canopy (Table 1). The wind speeds during each period were as follows (given as the frequency of occurrence of speeds in each of three classes, ≤ 0.8, 0.8 to 2.5, and > 2.5 m sec^{-1}); for 9 to 16 May, 49, 32, and 19; for 16 to 20 May, 56, 40, and 4; for 30 May to 6 June, 47, 40, and 13; and for 12 to 18 July, 58, 33, and 9. Mass median diameters were calculated from particle size distributions determined from cascade impactor samples collected during each period. Further details of air quality and meteorological conditions are available (*2*).
11. *Atmosphere-Biosphere Interactions: Toward a Better Understanding of the Ecological Consequence of Fossil Fuel Combustion* (National Academy of Sciences, Washington, D.C., 1981).
12. An upper-limit approximation of the effect of the canopy is the product of the single-surface deposition rate and the leaf area index (~ 6 m^2, foliar surface area per square meter of ground surface area).
13. Wet deposition above and below the canopy is the product of the growing season (1 April to 25 October 1977) precipitation or throughfall amount and the volume-weighted mean concentration in each. The sampling plus analytical reproducibility for wetfall is ± 5 to 15 percent, whereas rainfall amounts are accurate to ± 5 percent (*2*). We calculated the dry-deposition flux by using a simplified two-layer canopy (above and below canopy closure) for which deposition rates were measured separately (*2*), the temporal variation in the leaf area index, and the mean dry-deposition rates measured to inert surfaces (Table 1, relative standard errors of the means range from 0.23 for Mn, 0.24 for Zn, 0.33 for Pb, to 0.47 for Cd). Foliar leaching is calculated as the difference between the wet-deposition flux below the canopy (throughfall flux) minus the sum of the wet- plus dry-deposition fluxes to the canopy itself.
14. S. E. Lindberg, *Atmos. Environ.* **15**, 1749 (1981).
15. W. A. Hoffman, Jr., S. E. Lindberg, R. R. Turner, *Environ. Sci. Technol.* **14**, 999 (1980).
16. R. J. Lantzy and F. T. McKenzie, *Geochim. Cosmochim. Acta* **43**, 511 (1979).
17. We thank L. A. Barrie, J. N. Galloway, T. C. Hutchinson, D. S. Shriner, M. J. Levin, L. K. Mann, and F. L. Ball for helpful discussions and technical assistance. Research sponsored by the Office of Health and Environmental Research, U.S. Department of Energy, under contract W-7405-eng-26 with the Union Carbide Corporation. Publication 1839, Environmental Sciences Division, Oak Ridge National Laboratory.

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Global Sea Level Trend in the Past Century

Abstract. *Data derived from tide-gauge stations throughout the world indicate that the mean sea level rose by about 12 centimeters in the past century. The sea level change has a high correlation with the trend of global surface air temperature. A large part of the sea level rise can be accounted for in terms of the thermal expansion of the upper layers of the ocean. The results also represent weak indirect evidence for a net melting of the continental ice sheets.*

Sea level change is of current interest because of its possible sensitivity to climate change. It has been suggested, for example, that global warming due to increasing atmospheric CO_2 could melt the marine West Antarctic ice sheet, raising the global sea level 5 to 6 m (*1*). A sea level rise of as little as 15 cm may double the probability of damaging storm surges on the coast of Britain (*2*). Such a rise would also cause substantial beach erosion and the intrusion of seawater into low-lying areas that are now freshwater regions.

Many processes affect the sea level position measured on shorelines. Among the most important are eustatic sea level changes due to changes in the ocean water volume, caused mainly by the melting or growth of ice sheets, and isostatic adjustments of the earth's crust, caused mainly by ice sheet growth or decay and the associated change in the ocean water mass (*3, 4*). Tectonic movement and river sedimentation can generate local sea level trends comparable to eustatic and isostatic changes, as can changes in ocean currents and prevailing winds, although such trends are of limited duration.

We used tide-gauge measurements to estimate global sea level change in the past century. Data from more than 700 stations were obtained from the Institute for Oceanographic Science, Birkenhead, England. We excluded stations with records shorter that 20 years (a majority of the stations) and stations in seismically active areas such as the Pacific coast of Japan and in rapidly subsiding localities such as Galveston, Texas, and the Mississippi delta. The remaining 193 stations were divided into 14 regions on the basis of geographic proximity and the expected similarity of isostatic or tectonic behavior.

We reduced the individual station records to a common reference point by fitting a least-squares regression line to sea level as a function of time and by defining the zero point to be the value of the regression curve for 1940. The annual mean sea level curves for stations within a geographical region were then averaged to yield a mean sea level curve for each region. We obtained the global mean sea level curve by averaging the regional mean sea level curves, weighting each region equally (excluding the isostatically uplifting region of Scandinavia). Sea level trends, obtained by fitting a regression line through the mean sea level curves, are summarized in Table 1.

We also attempted to remove the long-term (usually 6000-year) sea level trends from the station data in order to obtain short-term sea level fluctuations, which are perhaps more appropriate for correlation with global climate variations in the past century. The cause of the long-term trend is uncertain. It has been argued that as much as 90 percent of it is residual isostatic uplift of continents due to the removal of the Wisconsin ice sheets (*4*). However, the long-term trend may contain a eustatic component, for example, due to a change in volume of the Antarctic or Greenland ice sheets.

The estimates for long-term sea level change are based on ^{14}C dating of measured positions of shoreline indicators in the geologic records, for example, mollusks, corals, and brackish-water peats (*5*). The 6000-year time interval was chosen to be as large as possible without approaching the period of the North American and European ice sheets, thus minimizing the effect of errors in the estimated sea level trend. However, for Scandinavia, where there has been a high rate of isostatic uplift since the last deglaciation, we used as a time interval the last 2500 years to improve the likelihood of obtaining a linear trend applicable to the past century. We corrected the dating of sea level positions to use the recent 5730-year estimate for ^{14}C half-life (*6*), rather than the 5570-year standard that had been used for most of the records (*5*), and to account for past atmospheric ^{14}C fluctuations (*7*).

The raw global sea level trend that we obtain for the past century is a rise of about 12 cm per century. After subtraction of the long-term trend, the result is a rise of 10 cm per century or 1 mm/year. The inferred long-term trend of 2 cm per century is very small as compared to the long-term trends of 1 m per century or more that are common at times of continental ice sheet growth or decay (*3*). Evidently the past few millennia have remained too warm to permit ice sheet formation on the North American or

Eurasian continents but too cold for substantial melting of the Greenland or Antarctic ice sheets.

We find that sea level rose in the past century in every geographical region except Scandinavia, and, after correction for long-term trends, sea level rose in every region except the west coast of South America where the change is smaller than the uncertainty based on the 95 percent confidence limit. We thus believe that this sea level rise is a true global trend and not, for example, a result of some regional variation in the geoid.

The sea level trend we find is similar to that obtained by Gutenberg [1.1 mm/year for the period 1807 to 1939 (*8*)], Fairbridge and Krebs [1.2 mm/year for 1900 to 1950 (*9*)], and Lisitzin [1.12 mm/year for 1807 to 1943 (*10*)], even though their studies were based on a much smaller number of tide-gauge stations. However, this rise is much less than the value recently reported by Emery [3 mm/year for 1935 to 1975 (*11*)]. Emery included stations in regions of known strong local uplift and subsidence (for example, Scandinavia and the east coast of Japan) and weighted each station equally; the result was that Scandinavia, Japan, and the east coast of the United States were heavily weighted. Emery (*11*) also reported a large increase in sea level (7 mm/year) in the decade 1966 to 1975 on the southeast coast of North America, and Wanless and Harlem (*12*) claim that sea level rose 10 to 14 cm in the past decade in south Florida; their result is based on the displacement of intertidal organisms on seawalls and bridge pilings. Our data (Fig. 1) show the steep rise in that region for 1966 to 1973, but the rise was not global and sea level decreased in the eastern United States for the period 1973 to 1977. Evidently long-term trends cannot be estimated from changes over 5- to 10-year periods for a single region. Even the global average curve (Fig. 2) has notable short-term variability.

The global sea level trend for the past century has some similarity to the trend in global surface air temperature (*13*). To quantify this, we computed the correlation coefficient between our global sea level curve and the global temperature curve of Hansen *et al.* (*13*), obtaining 0.6 when the annual mean curves are used for both quantities and 0.8 when the 5-year running means are used. Most of the positive correlation arises from the general increase in both sea level and temperature. Since one might expect a time lag between temperature change and sea level, we fitted a linear relation between our sea level curve and the global temperature trend,

$$S(t) = aT(t - t_0) + b \qquad (1)$$

where S and T are the 5-year means of global sea level and temperature, respectively, and t is time. The parameters a and b were obtained by least-squares linear regression, and the time lag t_0 was chosen to minimize the variance between Eq. 1 and the sea level curve. The results were $a = 16$ cm K^{-1}, $b = 0.3$ cm, and $t_0 = 18$ years.

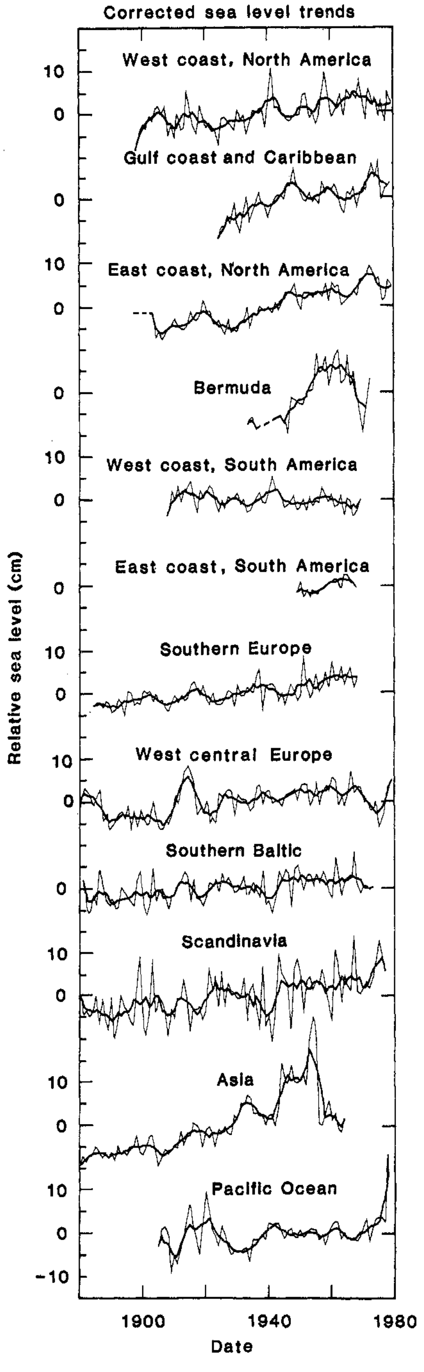

Fig. 1. Regional mean sea level trends. The heavy lines are 5-year running means. Long-range (6000-year) trends have been subtracted.

The sea level and temperature records are too short to allow much significance to be attached to this relationship. Nevertheless, it is interesting that the time lag of 18 years is of the order of the thermal relaxation time for the upper layers of the ocean, that is, the layers that are mixed in a time less than or comparable to the thermal relaxation time. This result suggests that part of the sea level rise may be attributable to thermal expansion, a possibility we can test by using the heat fluxes from the one-dimensional model of Hansen *et al.* (*13*). The two primary parameters or "tuning knobs" in that model are the equilibrium sensitivity (say, ΔT_{eq} for doubled atmospheric CO_2) and the rate of mixing of heat into the ocean beneath the mixed layer (specified by a diffusion coefficient k). Figure 2 illustrates the sea level change obtained for three values of the model's equilibrium sensitivity, k being constrained in each case to the value providing the best agreement with the observed global temperature trend of the past century. With the commonly accepted value for ΔT_{eq} of ~ 3°C, about half the observed sea level change is accounted for in terms of the thermal expansion of seawater. We tried other models for the mixing of heat into the ocean and obtained similar results.

We conclude that a large part, but probably not all, of the sea level rise of the past century is due to thermal expansion of the upper ocean. These results therefore also provide weak evidence for a decrease in the volume of the nonocean reservoirs of water. Some ground-water levels are known to have receded recently. For example, the huge Ogalalla reservoir in the high plains of the central United States may have dropped by several tens of meters (*14*), equivalent to a drop of a few millimeters of sea level. It seems possible that a sea level rise of a few centimeters could be accounted for in terms of a lowering of global groundwater levels. On the other hand, the trapping of water behind dams in the past century may have reduced sea level by 1 to 2 cm (*15*). Thus the net change in the water reservoirs of land areas is probably not a major cause of sea level change.

The magnitude of the global sea level rise therefore suggests that there has been a small net melting of the ice sheets. Although the present evidence is weak, continuation and refinement of this type of analysis is potentially important. As yet, direct observations of ice

Table 1. Sea level trends, 1880 to 1980, including correction for long-term (6000-year) trends.

Region	Sea level trend, 1880 to 1980			Corrected sea level trend, 1880 to 1980		
	Number of stations	Linear trend (cm/100 years)	95 percent confidence limit (cm/100 years)	Number of stations	Linear trend (cm/100 years)	95 percent confidence limit (cm/100 years)
West coast, North America	16	10	2	1	8	3
Gulf Coast and Caribbean	6	23	4	4	16	5
East coast, North America	32	30	2	30	15	2
Bermuda	1	26	16	1	20	16
West coast, South America	8	19	31	2	−3	3
East coast, South America	5	4	11	2	16	11
Africa	2	32	31	0		
Southern Europe	15	32	2	7	7	2
West central Europe	7	13	2	5	4	2
Southern Baltic	21	4	2	14	5	2
Scandinavia	47	−37*	3*	10	10	3
Asia	9	4	3	2	22	4
Australia	9	13	3	0		
Pacific Ocean	15	19	3	6	6	4
Global mean	193	12	1	86	10	1

*Not included in the global average.

sheet mass balance are not sufficiently accurate to establish even the sign of any trend (*16*).

A key application of the global sea level trend concerns the potential destruction of the marine West Antarctic ice sheet (*1*). It can be argued that the ice sheet is not close to disintegration, because it survived the Altithermal (~ 5000 years ago) when the global mean temperature was perhaps 1°C warmer than today. However, sea level, as well as temperature, must affect the ice sheet's stability. As indicated above, sea level has been flat, perhaps even rising slowly, over the past 5000 years. With the 10-cm rise of the past century, sea level must now be at or near its highest level since the previous interglacial, the Eemian 120,000 years ago (*17*).

Thus it is not inconceivable that the situation is near a point at which continued warming and rise of sea level could cause rapid, highly nonlinear disintegration of the ice sheet (*1*). We should emphasize that we have no evidence for such a process. Indeed, the sea level change we have deduced appears to be linear with temperature, and largely a result of the thermal expansion of seawater. Nevertheless, since sea level is at a high point and rising, the West Antarctic ice sheet warrants close attention.

Continued rise of sea level is likely in the near future, if predictions of global warming (*13*) are correct. The thermal expansion of seawater may raise sea level about 20 to 30 cm in the next 70 years (*18*); if slow ice sheet melting increases this by the same factor as in the past 100 years, a sea level rise of about 40 to 60 cm would occur by 2050. Thus we believe that substantial sea level change may occur even without rapid collapse of the West Antarctic ice sheet.

There is a clear need for improved observations. A direct measure of ice sheet growth or decay could be obtained from observations of ice sheet areal coverage and the altitude of the upper surface of the ice sheets that can be monitored by satellite. It is also desirable to measure the changes in the temperature profile in the ocean; these measurements would provide more explicit information on heat penetration and would serve as a crucial test of ocean-atmosphere climate models.

Finally, there is a need for observations and studies of sea level in the 1980's. A sharp global warming trend has been under way since the mid-1960's (*13*), and the current growth of atmospheric CO_2 and trace gases virtually assures that this trend will continue (*19*).

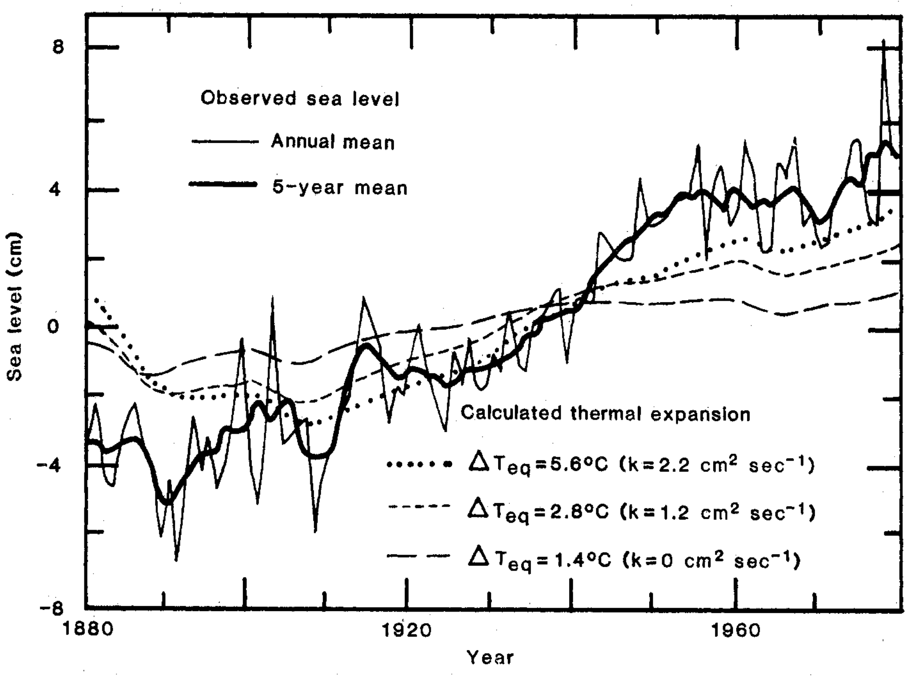

Fig. 2. Global mean sea level trend from tide-gauge data and comparison to the thermal expansion of the upper ocean obtained from the model of Hansen *et al.* (*13*) (see their equation 9 for the heat flux into the ocean). The radiative forcing used was CO_2 + volcanoes + sun [figure 5 in (*13*)], but a similar result would be obtained for other forcings that fit the observed global temperature trend; ΔT_{eq} is the equilibrium sensitivity of the model for doubled CO_2, and k is the diffusion coefficient beneath the mixed layer.

The sea level response to this warming should be carefully determined to aid our understanding of the processes and to allow early detection of any nonlinear response.

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References and Notes

1. T. Hughes, *Rev. Geophys. Space Phys.* **13**, 502 (1975); J. H. Mercer, *Nature (London)* **271**, 321 (1978).
2. J. R. Rossiter, in *The Sea*, N. M. Hill, Ed. (Interscience, New York, 1962), p. 595.
3. D. T. Donovan and E. J. W. Jones, *J. Geol. Soc. London* **136**, 187 (1979).
4. J. A. Clark, *J. Geophys. Res.* **85**, 4307 (1980).
5. N. A. Morner, Ed., *Earth Rheology, Isostasy and Eustasy* (Wiley, New York, 1980); W. Newman, unpublished data.
6. H. Godwin, *Nature (London)* **195**, 984 (1962).
7. H. E. Suess, *Radiocarbon* **20**, 1 (1978).
8. B. Gutenberg, *Geol. Soc. Am. Bull.* **90**, 618 (1979).
9. R. Fairbridge and O. Krebs, *Geophys. J. R. Astron. Soc.* **6**, 532 (1962).
10. E. Lisitzin, *Sea Level Changes* (Elsevier, New York, 1974).
11. K. O. Emery, *Proc. Natl. Acad. Sci. U.S.A.* **77**, 6968 (1980).
12. H. Wanless and P. Harlem, personal communication.
13. J. Hansen, D. Johnson, A. Lacis, S. Lebedeff, P. Lee, D. Rind, G. Russell, *Science* **213**, 957 (1981).
14. C. L. McGuiness, *U.S. Geol. Surv. Water Supply Pap. 1800* (1963).
15. M. I. L'vovich, in *World Water Resources and Their Future*, R. L. Nace, Ed. (Translation Board, American Geophysical Union, Washington, D.C., 1979), p. 264.
16. C. Bentley, J. Clough, K. Jezek, S. Shabtaie, *J. Glaciol.* **24**, 287 (1979); A. J. Colvill, *Polar Rec.* **18**, 390 (1977); S. Orvig, Ed., *World Survey of Climatology*, vol. 14, *Polar Regions* (Elsevier, New York, 1970), pp. 307–309.
17. W. S. Broecker and J. van Donk, *Rev. Geophys. Space Phys.* **8**, 169 (1970).
18. Thermal expansion in the next 70 years would be 20 cm for the slow energy growth (1½ percent per year) scenario of Hansen *et al.* (*13*), if the equilibrium sensitivity of the climate system is 2.8°C for doubled CO_2 and if heat is mixed diffusively into the thermocline with $k = 1.2$ cm^2/sec. Other trace gases (*19*) and a larger value for k [W. S. Broecker, T. H. Peng, R. Engh, *Radiocarbon* **22**, 565 (1980)] would increase the expansion.
19. A. Lacis, J. Hansen, P. Lee, T. Mitchell, S. Lebedeff, *Geophys. Res. Lett.* **8**, 1035 (1981).
20. We thank W. Newman for giving us access to his compilation of ^{14}C-dated sea level trends, and K. Emery, R. Fairbridge, and H. Wanless for helpful discussions.

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Escape of Hydrogen from Venus

Abstract. *Recombination of O_2^+ represents a source of fast oxygen atoms in Venus' exosphere, and subsequent collisions of oxygen atoms with hydrogen atoms lead to escape of about 10^7 hydrogen atoms per square centimeter per second. Escape of deuterium atoms is negligible, and the ratio of deuterium to hydrogen should increase with time. It is suggested that the mass-2 ion observed by Pioneer Venus is D^+, which implies a ratio of deuterium to hydrogen in the contemporary atmosphere of about 10^{-2}, an initial ratio of 5×10^{-5}, and an original H_2O abundance not less than 800 grams per square centimeter.*

Venus contains quantities of carbon and nitrogen similar to Earth, but hydrogen is deficient (*1*). The abundance of water on Venus is about 4.2 g cm^{-2}, which may be compared to the terrestrial value of 2.7×10^5 g cm^{-2}. Walker *et al.* (*2*) argued that the amount of H_2O on Venus should have been initially similar to that on Earth; they proposed that escape of H from a hot steamy atmosphere played a major role in Venus' early evolution. Lewis (*3*) thought that the H_2O content was low from the outset, because Venus formed in a warmer region of the solar nebula.

Analysis of processes influencing the current budget of Venus' hydrogen should shed some light on this issue. The upper atmosphere contains significant quantities of H, measured first by Mariner 5 (*4*). The density of H in outer regions of a planetary atmosphere provides direct information on the exchange of hydrogen with the interplanetary medium, and analysis of data for Earth (*5*) indicates an escape flux for H of 2.7×10^8 cm^{-2} sec^{-1}. The distribution of H is more complex for Venus, with at least two components (*6*), and interpretation is correspondingly ambiguous (*7*).

One component of Venus' exospheric H has a scale height of about 300 km, consistent with temperatures observed by Pioneer Venus (*8*). The second is more extensive, with a scale height of about 1000 km. Escape of atoms in the cold component is trivial, about 10 cm^{-2} sec^{-1}. Escape from the extended distribution may proceed more readily. The escape rate would be 1.5×10^6 cm^{-2} sec^{-1} if the velocity distribution were characterized by an effective temperature of 1000 K as suggested by the observed scale height.

Fast atoms comprising the extended component must be produced in the exosphere, either by acceleration of ambient H or by exothermic reactions involving hydrogen-bearing gases such as H_2. Proposed chemical sources (*9, 10*) include charge transfer of H with solar wind and a variety of reactions involving ionospheric ions, for example, reactions 1 through 6 in Table 1. We propose here an additional source, collisions of fast O with thermal H, reaction 7. Energetic O atoms, O*, are formed by the recombination of exospheric O_2^+ and CO_2^+, reactions 8 and 9. Recombination of O_2^+ proceeds mainly (*11*) by reaction 10, which represents a source for O* with an initial speed of 5.6 km sec^{-1}. Oxygen atoms formed in reaction 10 can escape directly from Mars (*12*). They are gravitationally bound to Venus but can induce significant escape of H by momentum transfer in reaction 7.

Consider an elastic collision of O* with stationary H and assume that the scattering is isotropic. The fraction of collisions leading to production of H atoms with speeds in excess of the escape velocity, 10.2 km sec^{-1}, is 6 percent for a O* velocity of 5.6 km sec^{-1} (*13*). The fraction is increased to approximately 15 percent if we account for the thermal velocity of H in an exosphere of temperature 300 K as implied by the cold H component observed by Mariner 5 (*6*).

The rate for escape of H due to reaction 7 on the dayside may be estimated as follows. Photoionization above the exobase leads to the production of O_2^+ and subsequently O*. The rate for production of O* is given by $2JN$, where J denotes the ionization rate (per second) and N is the column density of gas above the exobase (per square centimeter). Typical values for J range from 5×10^{-7} to about 1.2×10^{-6}, with higher values appropriate for solar conditions during the Venus flyby of Mariner 5 (*14*). The value of N is about 3×10^{14} (*15*), and the resulting source for O* is 8×10^8 cm^{-2} sec^{-1}. The fraction of exospheric O* that collides with H depends on the relative abundance of H and is approximately 2 percent for conditions during the Mariner 5 flight (*16*). The corresponding rate for escape of H from the dayside is about 10^6 cm^{-2} sec^{-1}.

The source of exospheric O* and H* may be even larger at night. The nightside ionosphere is variable (*17*), with densities as high as 10^5 electrons per cubic centimeter at altitudes between 140 and 150 km (*18*). The dominant positive ion at these altitudes is O_2^+, produced by reaction 11. The Pioneer data imply a recombination rate for the ions of about 3×10^8 cm^{-2} sec^{-1} (*18, 19*), which may be supplied either by transport from the dayside thermosphere or by in situ production (*17–20*). The corresponding rate for production of nightside O* is 6×10^8 cm^{-2} sec^{-1}. In contrast to the dayside, most of this source is located in the exosphere. The density of H in the lower nightside exosphere is 100

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