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PRINCIPLES OF ISOTOPE GEOLOGY

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explosion of hydrogen bombs from 1954 to 1963. The introduction of artificially produced tritium into the atmosphere and hydrosphere has been used to study mixing rates in the atmosphere and in the oceans and to trace the movement of groundwater (Dincer and Davis, 1967). Although water molecules containing tritium are subject to isotopic fractionation, this effect is generally negligible compared to variations in the rate of input and subsequent decay of tritium to stable $^{3}_{2}He$ by beta emission.

The oxygen of the atmosphere is strongly enriched in ^{18}O and has a $\delta^{18}O$ value of +23.5‰. This is the so-called "Dole effect" (Dole et al., 1954) which may be due to preferential removal of ^{16}O by respiration of plants and animals. The oxygen released during photosynthesis is derived from the water molecule and has a weighted average $\delta^{18}O$ value of +5‰ (Garlick, 1969). Since the fractionation factor for isotope exchange of oxygen between water and oxygen is very nearly equal to one (Urey, 1947), the oxygen of the atmosphere is not in isotopic equilibrium with water of the hydrosphere.

Atmospheric CO_2 has a $\delta^{18}O$ value of +41‰. The fractionation factor for isotope exchange of oxygen between CO_2 and H_2O (l) at 25°C is 1.0407 (Garlick, 1969) which indicates that atmospheric CO_2 is in approximate isotopic equilibrium with sea water.

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Snow and Ice Stratigraphy

The isotopic composition of oxygen (and hydrogen) of snow deposited in the polar regions of the Earth and at high elevations in mountains depends primarily on the temperature. For this reason $\delta^{18}O$ values of snow are strongly negative and display variations both in terms of seasonal temperature fluctuations and variations related to altitude and geographic latitude. Systematic measurements of $\delta^{18}O$ and δD have therefore been used to study flow patterns of glaciers, snow accumulation rates, and climatic variations in the past 100,000 years.

The seasonal variation of $\delta^{18}O$ values of snow can be used to date snow and firn layers by identifying successive summer layers as a function of depth. Snow deposited during the summer has less negative $\delta^{18}O$ values than snow deposited during winter when air temperatures are lower. Complications arise in areas of low precipitation, such as the interior of Antarctica, because snow may be removed by deflation and may then be redeposited elsewhere. As a result, the thickness of annual snow layers may be variable and gaps may exist in the record. Nevertheless, $\delta^{18}O$ profiles of snow and firn at various places in Antarctica have been used to determine the average rates of accumulation of water. For example, Epstein et al. (1965) found an average annual accumulation rate of 7 cm of water at the South Pole during the time interval 1958–[illegible], based on a $\delta^{18}O$ profile of snow (Fig. 18.4). This estimate is in good agreement with results obtained by identification of annual layers by conventional stratigraphic methods. It also agrees with the accumulation rate obtained by dating snow at the South Pole by means of ^{210}Pb (Chapter 16.5, Fig. 16.5). Similar studies of this kind have been published by Gonfiantini (1965), Deutsch et al. (1966), and others. Moreover, Hamilton and Langway (1967) demonstrated a significant correlation between the concentration of dust particles (0.6 to 3 microns in diameter) in an ice core from Greenland (site 2, close to Camp Century) and $\delta^{18}O$ values.

The seasonal fluctuations of $\delta^{18}O$ in snow are gradually eliminated due to homogenization of the isotope composition of the oxygen. This results from several causes, including melting and refreezing of water percolating downward through snow or [illegible],