The models also predict the presence of the powerful oxidants OH and H_2O_2, which would be expected to be present in the soil as well as the atmosphere. The Viking landers performed soil analyses to test for metabolic activity, but it is generally thought that they detected oxidants such as those mentioned above instead. Another experiment looked for organic molecules in the martian atmosphere but found none, a result explained as owing to their destruction by OH. The Mariner 6 and 7 orbiters observed small amounts of ozone in the winter hemisphere but not in warmer regions (*13*), where the odd-hydrogen oxidants would destroy it.

The early work described above was carried out just a few years before the chemistry of Earth's stratosphere suddenly aroused wide attention because of concerns about loss of ozone caused by aircraft exhausts and chlorofluorocarbons. This chemistry had been of interest since Chapman's work in 1930 on the presence of ozone, but the numerical methods were greatly refined by the work on Mars (and Venus). The same reaction set that is important for Mars also plays a key role in Earth's mesosphere (the region just above the stratosphere), except that there is no need to invoke the photolysis of H_2O_2 because O atoms are much more abundant than they are on Mars.

Despite its scarcity on present-day Mars, water carries great interest in a variety of fields (*14*). The presence of eroded channels and other erosion features requires the liquid to have been important earlier in the planet's history. Water is also a component of the polar caps. It must be considered in any scenario for the formation of the solar system. If life ever existed on Mars, liquid water must have been present. And the traces of water vapor in the atmosphere have the profound effects discussed above.

The odd-hydrogen reactions summarized in the figure are responsible for the stability of CO_2, the principal constituent of the martian atmosphere, and the absence of even traces of organic gases. The molecular hydrogen reported in (*1*) is another sign of the presence of odd hydrogen and its photochemical activity.

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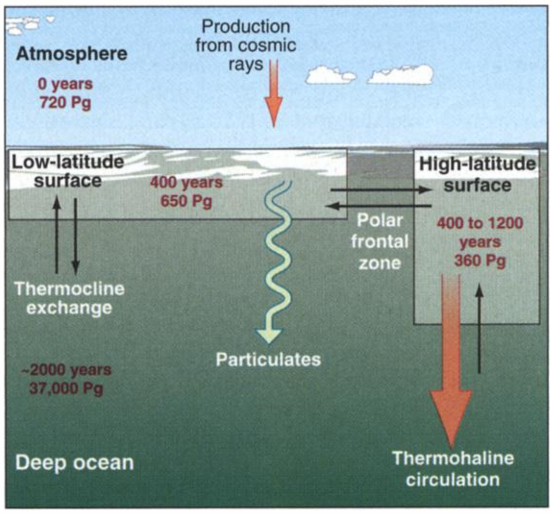
PERSPECTIVES: PALEOCLIMATE

Dating—Vive la Différence

Jess Adkins

Radiocarbon (^{14}C) is a potentially powerful chronometer in paleoclimate studies. Its half-life of 5730 years is well suited to date events from the modern to beyond the Last Glacial Maximum (about 20,000 years ago). However, realizing this potential has proven to be difficult. To know the true age of a carbon-containing sample, one must know the initial amount of ^{14}C at time zero. Because ^{14}C atoms are created in the atmosphere by cosmic rays and are exchanged among active carbon reservoirs on many time scales, this initial value is quite variable.

On page 1917 of this issue, Siani *et al.* (*1*) report a creative approach to circumventing these problems in dating marine sediment cores. The authors use volcanic ash layers to mark synchronous deposition of charcoal and planktonic foraminifera. The difference between radiocarbon dates from the charcoal (which reflects the ^{14}C content of the atmosphere) and the foraminifera (which reflect surface ocean conditions) from the same core is the "reservoir" age of the surface ocean. Using the contemporaneous charcoals to adjust for variability in initial ^{14}C, the planktonic ^{14}C dates can be placed on an absolute age scale, provided the charcoals have not been altered. Marine sediments are among the most important archives of past climate, and the ability to constrain the phasing between them and other records (such as ice cores and terrestrial data) is therefore of fundamental importance.



The ocean/atmosphere radiocarbon system. Today, surface waters between ~40°N and ~40°S have a uniform radiocarbon age of 400 years (the "reservoir age"). A leak of older waters from below pulls them away from the atmospheric value. Modern high-latitude surface waters are older and more variable. The exception is the North Atlantic, where ages remain at 400 years. This feature is the result of the position of the polar front close to the east coast of Greenland. In the past, the position of this front, and therefore the location of old reservoir ages to the north of it, have moved far to the south. Ages are for the inorganic carbon in the reservoir (organic carbon in the ocean is a small fraction of the total) and sizes are in petagrams (10^{15} grams).

To better understand Siani *et al.*'s approach, we need to know how radiocarbon is distributed among its various reservoirs in the ocean/atmosphere system (see the figure). ^{14}C data are reported as $\Delta^{14}C$, a normalized deviation from a standard, in units of per mil (‰) (*2*). New ^{14}C atoms are produced by cosmic rays in the upper atmosphere at a rate of about 2 atoms cm^{-2} s^{-1}. This radiocarbon is quickly oxidized to CO_2 and exchanges with the high- and low-latitude oceans. At high latitudes where new deep waters are formed, ^{14}C is carried to the abyss as dissolved inorganic carbon. Low-latitude surface waters exchange carbon by mixing across the thermocline (*3*), thus increasing the reservoir age away from the atmospheric value. Surface waters also export carbon as organic and $CaCO_3$ detritus, but this has a small effect on the oceanic ^{14}C budget.

Because the bulk of the radiocarbon is contained in dissolved

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inorganic carbon in the deep ocean, this is where 90% of the cosmic ray–produced ^{14}C decays. In contrast, the atmosphere is one of the smallest reservoirs of radiocarbon and is thus very susceptible to changes in the fluxes between them. If all deep-water formation were shut off, then the ^{14}C production rate would remain unchanged but the accessible carbon reservoir would be much smaller. This would cause atmospheric Δ^{14}C to rise at a rate of ~2‰/year. Changes of this type have indeed been observed in the atmospheric record.

Because the atmosphere's Δ^{14}C value is the basis for all radiocarbon ages, the most recent literature is periodically summarized in a calibration paper in which radiocarbon ages are compared with true "calendar" ages (*4*). The bulk of the data comes from tree rings and layered sediments, and the curve is well constrained back to about 14,600 years ago. Spot measurements of uranium-rich shallow water corals extend the record back to about 24,000 years (*5, 6*), and a recent stalagmite record extends all the way back to the limit of ^{14}C dating at ~40,000 years (*7*). These records demonstrate variations in atmospheric Δ^{14}C with large amplitudes at many time scales.

The variations cannot all be explained by changes in the ^{14}C production rate and must therefore reflect changes in carbon fluxes and hence climate (*8*). It is these climate changes that give rise to variations in the surface radiocarbon reservoir ages measured by Siani *et al.* (*1*). Together with the recent work of Waelbroeck *et al.* (*9*), the results allows us to paint a consistent picture of the timing between changes in North Atlantic sea surface temperature and the air over Greenland.

Waelbroeck *et al.* estimated the reservoir age by assuming that North Atlantic sea surface temperatures and air temperatures over Greenland had to be in phase. Using the difference between the ice sheet model and their sediment radiocarbon ages, they calculated reservoir ages for their cores. Siani *et al.* measured the ice versus sea surface temperature phase independently by using charcoal dates to constrain the past atmosphere. The two papers agree that the Younger Dryas, a ~1500-year-long return to cold climate during the last deglaciation, saw a twofold increase of the high-latitude reservoir age of the North Atlantic (*10*) and that this signal did not make it as far south as the Straits of Gibraltar.

However, they disagree about Heinrich event 1 (H1), when a massive discharge of icebergs into the North Atlantic about 16,000 years ago caused the polar frontal zone, the region between cold high-latitude waters and warmer subtropical waters, to shift far to the south. Waelbroeck *et al.* (*9*) assume that their core at 38°N (SU-81-18) did not have a change in reservoir age during H1. But Siani *et al.* measure a twofold change in this value in the Mediterranean, several degrees farther to the south.

The studies can be reconciled if we allow the reservoir age in Waelbroeck's core to increase by about 400 years during H1. This increase leads to a better fit between the start of the sea surface temperature increase in the core and the Greenland air temperature increase at the end of H1. In Waelbroeck *et al.*'s analysis (*9*), the air temperature increase occurs at the midpoint of the sea surface temperature rise.

Why do these shifts in the radiocarbon age of surface waters occur? A first answer is that the polar front moves past the core site (see figure caption). But that just moves the questions to why the polar waters themselves are older. Variability in both sea ice cover and radiocarbon ages of intermediate waters (the explanation favored by Siani *et al.*) is the most likely explanation to this question. However, a change in the exchange rate across the low-latitude thermoclime will also change the radiocarbon age and could have profound implications for the heat budget of the past ocean. Measuring the radiocarbon ages of intermediate waters of the past would be a very useful constraint to this problem.

References and Notes

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PERSPECTIVES: SIGNAL TRANSDUCTION

A New Thread in an Intricate Web

Mark von Zastrow and Keith Mostov

Most neurotransmitters, hormones, and growth factors activate cellular signaling pathways by binding to specific membrane receptors. These signaling pathways subsequently modulate biochemical networks comprising various cytoplasmic kinases, phosphatases, and guanosine triphosphate (GTP)–binding proteins. Information flow through these complex signaling networks requires a precise spatiotemporal organization of the appropriate signaling partners, many of which interact promiscuously when isolated from their native cellular environment. Thus, a critical aspect of cellular signaling is a matter of molecular choreography: getting the right proteins to the right place at the right time. It is not surprising, then, that membrane-trafficking pathways—which determine the structure and biochemical composition of the specialized membrane compartments in eukaryotic cells—have important effects on cellular signal transduction and, conversely, that signaling events can modulate membrane trafficking. Indeed, there appear to be many functional interactions between the otherwise distinct processes of signaling and membrane trafficking. This realization, which emerged over the last decade (*1*), has motivated a convergence between traditionally separate fields of cell biology. From this convergence emanates the question: How are signaling and membrane trafficking related at the molecular level? On page 1939 of this issue, Zheng *et al.* (*2*) describe a protein, RGS-PX1, that may be a new molecular thread in the intricate web that links signaling and membrane trafficking events.

Zheng and colleagues mined sequence databases for candidate proteins containing RGS (regulators of G protein signaling) domains. RGS domains are conserved in diverse organisms and have profound effects on cellular signal transduction triggered by seven-transmembrane G protein–coupled receptors (GPCRs). GPCRs trigger signaling by prompting guanine nucleotide exchange on the α subunit of heterotrimeric G proteins. This results in conversion of the "inactive" guanosine diphosphate (GDP)–bound α subunit to the "activated" GTP-bound form. RGS proteins are crucial for accelerating the conversion of the activated G

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