by 1 to 3 m and are interpreted to be the result of the sublimation of residual solid CO_2. Malin *et al.* estimate that if these observed annual scarp retreat rates were interpreted as an actual net loss of solid CO_2 from the south residual polar cap, the mass of the martian atmosphere would be increasing at a rate of about 1% per martian decade.

Are the large-scale changes in the current state of the martian atmosphere and the south polar cap suggested by these new observations reasonable? At least three considerations suggest that the answer may be yes. First, the mass of the atmosphere is extremely sensitive to the equilibrium vapor pressure of a permanent CO_2 deposit. In Leighton and Murray's model, a 1% change in the mass of the martian atmosphere would only require a 0.1% change in the long-term solar reflectance of a residual CO_2 deposit. Given this sensitivity, it may not be unreasonable to expect that interannual or longer term variations in the south residual polar cap's radiative properties might cause it to not be in precise vapor equilibrium with the present atmosphere.

Second, most previous measurements of the annual CO_2 cycle have been interpreted as consistent with annual repeatability, but the uncertainties in these measurements cannot preclude interannual variations in the mass of the martian atmosphere and south polar cap of the magnitudes suggested by Malin *et al.* Surface pressure measurements made by Viking Lander 1 over four Mars years show interannual differences on the order of 0.65% (*12*). Annual radiation budget measurements made by the Viking orbiters indicate a net annual imbalance at the south residual cap of -2 ± 5 W m^{-2}. This would correspond to a net annual gain or loss of about 20 ± 50 cm of CO_2 over the course of the year (*10*).

Third, comparisons between previously obtained images by the Mariner 9 and Viking orbiters have already shown large-scale changes in the extent of the bright south residual polar cap (*13*). The new MOC images may thus represent higher resolution views of previously observed interannual variations. Interestingly, the south residual cap observed by Viking appeared to contain more solid CO_2 than the one observed by Mariner 9 three Mars years earlier (*13*), implying that interannual changes can take place in both directions.

The suggestion that the martian carbon dioxide cycle may change from one year to the next adds an exciting new dimension to our study of the planet. During the next decade, we hope to see a quickening of the pace and a broadening of the scope of our exploration of the Red Planet. We hope to see additional long-term orbital observations of the present climate like those presented in these reports. Furthermore, landers and rovers in the polar regions and elsewhere on the planet should search for records of past climate change. From experience gained by studying the Earth's climate, we can anticipate that obtaining a deep understanding of martian climate processes and climate history will be a challenging endeavor. However, given what we are learning today, we have every reason to expect that the story that Mars has to tell us will be an interesting one.

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PERSPECTIVES: OCEAN CIRCULATION

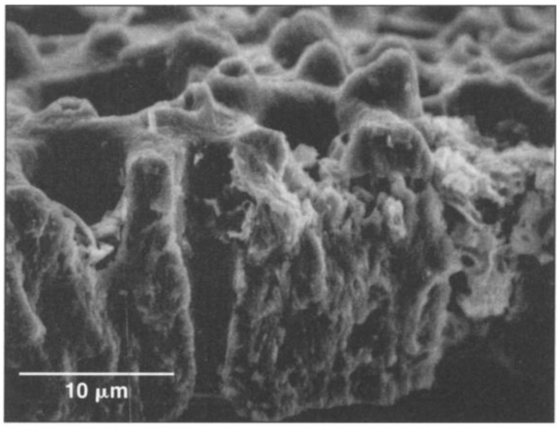
Thin Walls Tell the Tale

David Archer and Pamela Martin

Much of what we know about the oceans of the past has been gleaned from the microscopic shells of open-ocean plankton called foraminifera. We can reconstruct the temperature, chemistry, and circulation of the ocean from the distribution of species and their trace chemical makeup. On page 2152 of this issue, Broecker and Clark (*1*) show that foraminifera can tell us even more about the past. They use the thickness of their shell walls (see the figure) as an indicator of ocean chemistry and the carbon cycle.

The $CaCO_3$ shells dissolve and their walls get thinner when the concentration of carbonate ion, CO_3^{2-}, in seawater falls below the saturation value. Because the saturation value increases with increasing pressure, $CaCO_3$ is preserved on topographic highs such as mid-ocean ridges and completely dissolved in the deepest abyss. The concentration of CO_3^{2-} in seawater is intimately associated with pH and the concentration of dissolved CO_2 gas, and a reconstruction of deep ocean CO_3^{2-} can therefore provide clues about the circulation and carbon cycle of oceans past.

In today's ocean, convection in the North Atlantic carries high pH, high CO_3^{2-} waters into the abyss. The deep Pacific, in contrast, is unventilated, and its CO_3^{2-} concentration and pH are lower. Questions of ocean pH are at the heart of understanding the glacial/interglacial carbon dioxide concentration cycles of the atmosphere and hence the question of why there are ice ages at all. An increase in the pH of the whole ocean could explain the lowered glacial carbon dioxide concentrations during glacials (*2*).



The shell wall of a planktonic foraminifera of the species *Orbulina universa*.

CREDIT: PAMELA MARTIN

A number of methods have been used to measure the extent of past $CaCO_3$ dissolution. One potential indicator is the bulk $CaCO_3$ concentration of the sediment, but changes in $CaCO_3$ also depend on changes in $CaCO_3$ production or dilution by clays. Another is the relative abundance of the more dissolution-resistant species of foraminifera (*3*, *4*). Both of these methods tend to be insensitive to the initial stages of dissolution, which are thought to occur even when the bottom waters above the sediment are supersaturated.

Broecker and Clark (*5*) have developed a dissolution index based on the size distribution of $CaCO_3$ pieces to track the breakup of large foraminifera shells into fragments. The abundance of fragments with diameters of more than 63-μm tracks overlying water CO_3^{2-} nicely today. However, Broecker and Clark's most dogged critic, Broecker and Clark (*6*), showed that the initial size distribution of $CaCO_3$ must have been differ-

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ent during glacial times. Perhaps coccolithophorids, which make small plates of $CaCO_3$, were relatively more abundant then. Without a stable initial condition, size frequency loses its reliability as an indicator of dissolution.

The consideration of shell thickness was initiated by Lohmann (*7*), who heroically measured all three exterior dimensions of thousands of individual microscopic shells and weighed each of them to compute the thickness of the walls between the chambers. He found that the shell walls get systematically thinner with increasing water depth above the sediments. This is consistent with the effect of pressure on solubility, but the novelty was that the wall thinning began at depths where the overlying water is highly supersaturated with respect to $CaCO_3$. This is presumably the result of respiratory CO_2 driving down the pH of sediment pore waters. Here at last was a reliable indicator of the initial stages of $CaCO_3$ dissolution, which would be useful in regions in the middle water column where the column was supersaturated with respect to $CaCO_3$ (1 to 4 km)

Broecker and Clark streamlined the method by mechanically sieving a narrow shell size range and then weighing an ensemble rather than individual shells. They found a good correlation between shell thickness at the top of the sediment core and CO_3^{2-} concentration in the overlying water (*8*). Broecker and Clark now extend this work to the last glacial (*1*). The thickness results indicate that during this time, the intermediate Atlantic CO_3^{2-} concentration was considerably higher than it is today whereas that in the deep Atlantic was lower, leading to a dramatic gradient in CO_3^{2-}.

These results are apparently inconsistent with the distributions of geochemical proxies of dissolved nutrients, the $\delta^{13}C$ and the trace Cd concentrations of foraminifera, which uniformly show nutrient-rich waters at these sites. However, pH and nutrients may deviate from each other to some extent because CO_2 is transported through the atmosphere and nutrients are not. More startling is the inference that the gradient in deep ocean CO_3^{2-}, which is indicative of the flow path of ventilated water, was reversed during glacial time. If proven correct, this difference would have profound implications for the circulation of the deep ocean and the climate of the atmosphere above that drove it.

One of the longest sections in Broecker and Clark's report (*1*) is the discussion of caveats. One is the possibility that shells from different regions, climates, or CO_2 concentrations (*9*) may differ systematically in their shell weights before any dissolution has taken place. Broecker and Clark (*8*) have already documented an offset between the Atlantic and the Pacific and Indian present-day oceans. Even this possibility cannot explain all of the discordant results from different foraminiferal species in cores from the same region (*10*).

More measurements from around the world may or may not end up telling a coherent story. But however it turns out, Broecker and Clark are to be commended for making every effort to test and validate this novel technique. Thus does paleoceanography proceed by proxy.

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PERSPECTIVES: CLIMATE

Climate Variability and the Influence of the Sun

Joanna D. Haigh

The search for the signs of solar variability in climate records has historically been dogged by overstated claims and dubious statistics. In addition, ground-based measurements were insufficiently sensitive to detect variations in the Sun's radiative output. As a result, the subject has tended to be marginalized by the meteorological establishment. However, recent evidence from satellite data has shown that the term "solar constant" is a misnomer. Furthermore, climate change attribution requires the ability to separate natural climate-forcing mechanisms from human-induced factors. It is thus more important than ever that the influence of solar variability on climate is assessed and understood.

To detect a solar influence on climate, two key challenges must be met. First, reliable and consistent long-term indices of meteorological and solar parameters must be constructed. Second, any solar influence must be separated from other influences in what is an inherently highly variable and noisy nonlinear system. Even when a statistically robust solar signal emerges from the data, the mechanisms by which the solar influence acts remain to be explained, particularly because the irradiance variations are small and their direct effect must be amplified in the climate system. Hence, dynamical and physical feedbacks need to be considered. Among these effects may be the coupling of processes in the lower atmosphere to ocean circulation or to the middle atmosphere.

On page 2130 of this issue, Bond *et al.* (*1*) meet the first challenge mentioned above. They investigate the influence of the Sun on the climate of the North Atlantic region during the Holocene (from about 11,000 years ago to the present) using data extracted from deep-sea sediment cores. The level of solar activity is estimated from the concentrations of ^{14}C and ^{10}Be isotopes, which are produced by the action of galactic cosmic rays. Cosmic rays are more intense at Earth's surface when the Sun is less active. The climate is indicated by the concentrations of mineral tracers, which were deposited from drift ice circulating in the subpolar North Atlantic. Increases in the tracers indicate a southward expansion of cooler, ice-bearing water.

The authors show that each of the expansions of cooler water that occur roughly every 1500 years was associated with a strong minimum in solar activity. This remarkable result provides strong evidence that solar activity does indeed modulate climate on centennial to millennial time scales. The authors suggest that changes in ocean thermohaline circulation, associated with the supply of fresh, low-density water from the drift ice and coupled to atmospheric circulations, may amplify the direct effects of small variations in solar irradiance.

To learn more about potential mechanisms by which the solar signal may be amplified, we need to consider the natural modes of climatic variability in the climate system. The El Niño–Southern Oscillation phenomenon is the leading mode of natural variability in the tropics, although its influence is felt globally. At higher latitudes in the Atlantic, the leading winter mode is the North Atlantic Oscillation. These modes may be influenced by external forcing factors, which include natural components (such as solar variability or

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