stratospheric samples has a similar low value of $n_{\text{IM}} \approx 1.5 \times 10^{-7}$ at 700 nm; it increases to approximately 0.007 for the 15-km sample and to 0.02 for the 18-km samples at 300 nm. The measured $n_{\text{IM}}$ values for the stratospheric samples are in the same range as those for the surface samples. The stratospheric samples, however, show a greater wavelength dependence than the surface samples.

In contrast, a highly absorbing anthropogenic aerosol, elemental carbon, has an $n_{\text{IM}}$ value of $\approx 0.6$ throughout the visible spectrum (12). A white substance such as ammonium sulfate has an $n_{\text{IM}}$ value of $< 10^{-7}$ (13) in the visible, and representative light tan desert aerosols have $n_{\text{IM}}$ values that range from $\sim 0.003$ at 700 nm to $\sim 0.002$ at 300 nm (14).

The volcanic material measured by Pollack et al. (11) has smaller values of $n_{\text{IM}}$ throughout the visible spectrum; a synthetic aerosol modeled by Ivlev and Popova (15) that was used for stratospheric calculations has an $n_{\text{IM}}$ value of 0.005 through the visible; this value is larger than the stratospheric values for the wavelengths greater than 380 nm, a wavelength range that encompasses $> 90$ percent of the extraterrestrial solar flux.

In the absence of more detailed calculations, the modeled effects of the Mount St. Helens ash might be expected to be intermediate between those of the basalt aerosol (5) and those of the Ivlev and Popova aerosol (2, 4). In particular, some significant in situ stratospheric heating due to the absorption of solar radiation by the aerosol would be expected; for particles of the size range with the greatest stratospheric lifetime, a net surface cooling, though by an amount smaller than that modeled for basalt, would be expected.

E. M. Patterson
School of Geophysical Sciences, Georgia Institute of Technology, Atlanta 30332

References and Notes

Marine Macrophytes as a Global Carbon Sink

Abstract. Marine macrophyte biomass production, burial, oxidation, calcium carbonate dissolution, and metabolically accelerated diffusion of carbon dioxide across the air-sea interface may combine to sequester at least $10^7$ tons of carbon per year in the ocean. This carbon sink may partially account for discrepancies in extant global carbon budgets.

Approximately $5 \times 10^7$ tons of carbon are released to the atmosphere annually as CO$_2$ by the burning of fossil fuel (1). Carbon dioxide remaining in the atmosphere accounts for about 50 percent of the fossil fuel carbon released; about 40 percent is estimated to diffuse across the air-sea interface into the dissolved CO$_2$ pool of surface ocean water (2-4). It has been suggested by default that increasing biomass on land accounts for the remaining 10 percent (2, 4). An alternative argument is that there has been a net reduction of terrestrial biomass equivalent to 20 to 100 percent of the carbon liberated from fossil fuel (3, 5, 6).

Carbon sinks may fail to account for $0.5$ to $5 \times 10^7$ tons of carbon per year. Substantial carbon sinks that have been overlooked or erroneous evaluation of...
phyte ecosystems is only about 5 percent of the oceanic production, but the macrophyte biomass is about two-thirds of the oceanic biomass (Fig. 1). Moreover, the area covered by the plankton ecosystems (Fig. 1, O to Q) is $3.6 \times 10^8$ km$^2$, whereas the macrophytes (Fig. 1, R plus S) occupy only about $2 \times 10^8$ km$^2$ (9). Thus, the biomass per unit area occupied by the macrophytes is about 400 times that of the plankton. Ecosystems with a rapid turnover of stored carbon may be biologically important, but they are ineffective carbon sinks; the carbon has no significant reservoir in which to accumulate. Thus, forests on land are more effective sinks than are grasslands or cultivated lands (Fig. 1). By analogy, macrophyte ecosystems are more effective carbon sinks than are planktonic ecosystems. Controversy over the importance of the terrestrial biospheric carbon sink mostly deals with the changing biomass of forests (2-6). In other words, the "points" represented by A to E on Fig. 1 are not individual points but are the loci of points moving at a controversial rate and direction, roughly parallel to the constant-turnover lines.

Large-scale variations in macrophyte biomass are known on a time scale of decades. Perhaps the best documented, although poorly understood, example is the massive die-off of the sea grass Zostera marina in the North Atlantic Ocean. This plant decreased dramatically on both sides of the Atlantic in the 1930's and apparently recovered to near its earlier biomass by the late 1960's (10). If this biomass fluctuation and other more poorly documented variations are at all synchronous, then ecosystems R and S on Fig. 1 may also be loci of points shifting significantly through time.

The actual rate of carbon entry into the macrophyte reservoir is inadequately known. The sum of ecosystems R plus S (Fig. 1) suggests macrophyte production to be about $1 \times 10^9$ tons of carbon per year. De Vooys (11) reviewed primary production in aquatic environments and derived a value of $0.8 \times 10^9$ tons of carbon per year for "kelps, other weeds, angiosperms and coral reefs." Bunt (12) derived a value for "neritic waters," including plankton and benthic microflora, of $4 \times 10^8$ tons of carbon per year. The area of the sea floor covered by each of the major macrophyte communities may be the most poorly defined term in evaluating the macrophyte contribution to the global carbon budget, and the changing area of macrophyte ecosystems through time has not been addressed on a global scale.

Macrophyte production is likely to be at least the value of $10^9$ tons of carbon per year as derived from Whittaker and Likens (9) (Fig. 1). This figure is 20 percent of the annual fossil-fuel carbon input to the atmosphere (1), half the estimated direct carbon storage in oceanic CO$_2$ by gas diffusion across the air-water interface (2, 4), twice the amount some investigators attribute to an increase in the terrestrial biosphere pool (2, 4), and 20 to 100 percent of the rate of shrinkage of that terrestrial biospheric pool estimated by others (5, 6).

Little of the macrophyte biomass, either grass or algae, is directly grazed and converted to animal biomass (8, 13). Some enters the food web as detritus (8, 13, 14). Direct burial provides a net carbon sink of unknown magnitude, but detrital material that is oxidized and liberates CO$_2$ may also serve as a mechanism for CO$_2$ storage. Locally high CO$_2$ partial pressure generated on the sea floor or in the sediments by decaying macrophytes makes the surrounding water more capable of dissolving CaCO$_3$, thus adding CO$_2$ to the alkalinity pool of the oceans. Calcareous organisms are abundantly associated with, and often grow on, both sea grasses and algae (15); oxidation coupled with CaCO$_3$ dissolution may, therefore, be substantial in storing excess CO$_2$ in the form of increasing alkalinity.

Macrophytes that are transported to deep water and oxidized (13, 14, 16) induce a net CO$_2$ change on deep waters that are not equilibrating with the atmosphere except by exchange with surface waters. There is evidence for CO$_2$ liberation from organic-rich deep-sea sediments, in response to oxidation of organic material (17). This may or may not be associated with CaCO$_3$ dissolution. Either the shallow or deepwater oxidative pathways liberate essential nutrients back into the water for continued biochemical cycling while delivering the carbon to the oceanic CO$_2$ pool.

Organic carbon production in the ocean reduces the local CO$_2$ content of surface seawater and lowers CO$_2$ partial pressure. Because plankton production is a low-biomass, fast-turnover process.

Table 1. The CO$_2$ partial pressure of water impinging on coral reefs in the Houtman Abrolhos Islands, Western Australia.

<table>
<thead>
<tr>
<th>Date</th>
<th>Coral transect</th>
<th>Algal transect</th>
</tr>
</thead>
<tbody>
<tr>
<td>September 1979</td>
<td>162</td>
<td>114</td>
</tr>
<tr>
<td>March 1980</td>
<td>280</td>
<td>261</td>
</tr>
</tbody>
</table>
it does not ordinarily induce a large local reduction of CO₂. By contrast, macrophyte production can induce large local changes in CO₂ partial pressure, because the turnover rate of carbon from macrophyte photosynthesis is high due to carbon cycling in biomass. Coastal CO₂ anomalies are both temporally and spatially patchy, because macrophyte distribution and production are not homogeneous. Water-column CO₂ partial pressures, calculated from hourly pH and alkalinity sampling in a Western Australia seagrass bed, varied by about 30 μatm (10 percent) over 24 hours in response to community metabolism (Fig. 2). Mean CO₂ partial pressures for sites on a Western Australia coral reef in macroalgae varied between winter and summer by approximately 120 μatm and between sites by 20 to 50 μatm (Table 1). Because areas of high macrophyte production are often buffered by breaking waves, gas transfer coefficients at coasts are likely to be elevated above open ocean values.

Marine macrophyte organic carbon production, biomass storage, burial, oxidation, oxidation-induced CaCO₃ dissolution, and metabolically accelerated gas exchange across the air-sea interface are alternative expressions for a single, complex carbon sink. Although the components of this sink do exist, their quantitative significance on a global scale is not yet known.

S. V. Smith
Hawaii Institute of Marine Biology,
University of Hawaii, Kaneohe 96744

References and Notes