



Marine Macrophytes as a Global Carbon Sink

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stratospheric samples has a similar low value of n_{IM} ($\sim 1.5 \times 10^{-3}$) 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_{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_{IM} value of ~ 0.6 throughout the visible spectrum (12). A white substance such as ammonium sulfate has an n_{IM} value of $< 10^{-7}$ (13) in the visible, and representative light tan desert aerosols have n_{IM} values that range from ~ 0.003 at 700 nm to ~ 0.02 at 300 nm (14).

The volcanic material measured by Pollack *et al.* (11) has smaller values of n_{IM} throughout the visible spectrum; a synthetic aerosol modeled by Ivlev and Popova (15) that was used for stratospheric calculations has an n_{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.

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References and Notes

1. *Eos* **61**, 555 (1980).
2. J. A. Coakley, Jr., and G. W. Grams, *J. Appl. Meteorol.* **15**, 679 (1976).
3. Harshvardan and R. D. Cess, *Tellus* **28**, 1 (1976).
4. G. Fiocco, G. W. Grams, A. Mugnai, in *Proceedings of the Symposium on Radiation in the Atmosphere*, H. J. Bolle, Ed. (Science Press, Princeton, N.J., 1977), p. 74.
5. J. B. Pollack, O. B. Toon, C. Sagan, A. Summers, B. Baldwin, W. Van Camp, *J. Geophys. Res.* **81**, 1071 (1976).
6. G. Mie, *Ann. Phys.* **25**, 377 (1908).
7. K. F. Palmer and D. Williams, *Appl. Opt.* **14**, 208 (1975).
8. E. M. Patterson, in *Second Conference on Atmospheric Radiation* (American Meteorological Society, Boston, 1975), p. 177.
9. E. M. Patterson, D. A. Gillette, B. H. Stockton, *J. Geophys. Res.* **82**, 3153 (1977).
10. J. D. Lindberg and L. S. Laude, *Appl. Opt.* **13**, 1023 (1974).
11. J. B. Pollack, O. B. Toon, B. N. Khare, *Icarus* **19**, 372 (1973).

12. J. T. Twitty and J. O. Weinman, *J. Appl. Meteorol.* **10**, 725 (1971).
13. O. B. Toon and J. B. Pollack, *J. Geophys. Res.* **81**, 5733 (1976).
14. E. M. Patterson, *ibid.*, in press.
15. L. S. Ivlev and S. I. Popova, *Izv. Acad. Sci. USSR Atmos. Oceanic Phys.* **9**, 1034 (1973).

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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 10^9 tons of carbon per year in the ocean. This carbon sink may partially account for discrepancies in extant global carbon budgets.

Approximately 5×10^9 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 5×10^9 tons of carbon per year. Substantial carbon sinks that have been overlooked or erroneously evaluated of

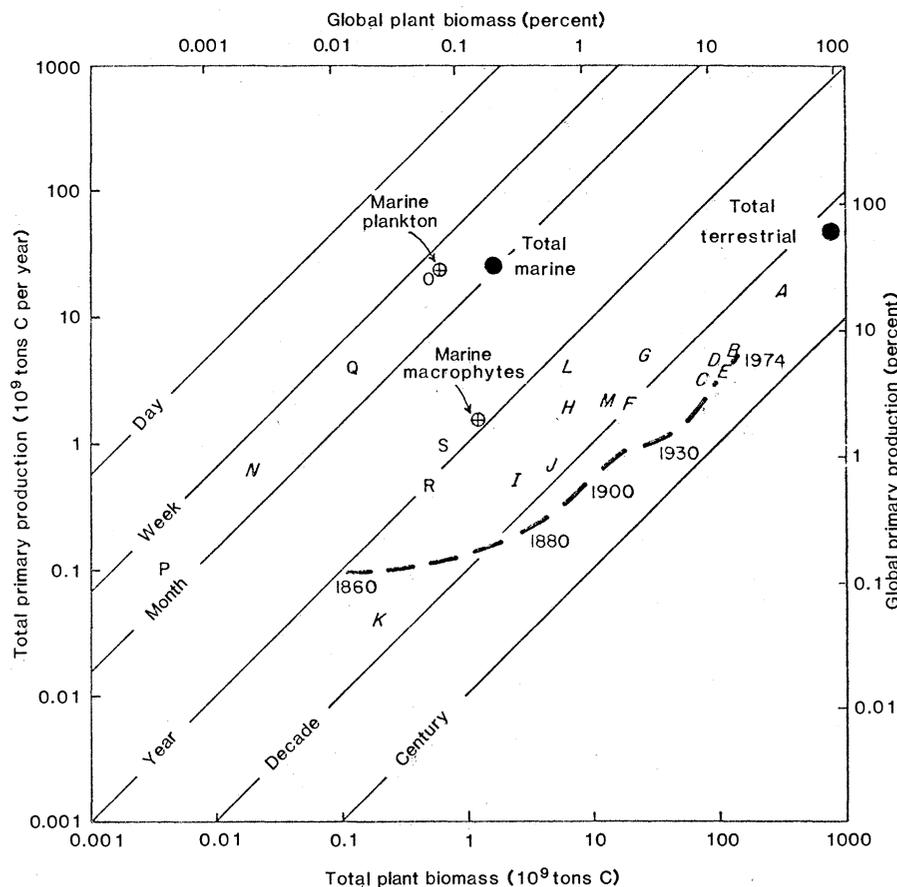


Fig. 1. Primary production, biomass, and turnover time for carbon in the biosphere, adapted from (9). Letters A to N (italic) indicate terrestrial ecosystems and remaining letters O to S (roman) indicate marine ecosystems: A is tropical rain forest; B is tropical seasonal forest; C is temperate evergreen forest; D is temperate deciduous forest; E is boreal forest; F is woodland and shrubland; G is savanna; H is temperate grassland; I is tundra and alpine meadow; J is desert scrub; K is rock, ice, and sand; L is cultivated land; M is swamp and marsh; N is lake and stream; O is open ocean; P is upwelling zones; Q is continental shelf; R is algal bed and reef; and S is estuaries. The cumulative fossil carbon input and the input rates (I) are shown as a dashed line. Since the source is not constant in size or input rate, the line represents a locus of points through time.

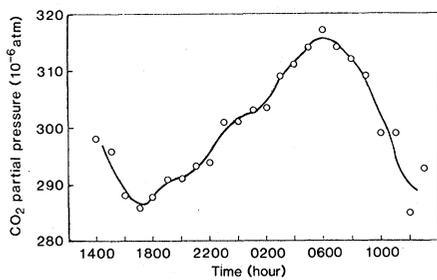


Fig. 2. Change in CO_2 partial pressure in a grass bed over a diel cycle (water depth = 7 m). The points are the hourly averages of 1- and 6-m samples. The line is a 2-hour moving average through the data points.

recognized sinks may explain the discrepancy. In this report, I have reevaluated aspects of marine biota as a carbon sink.

Plankton production in the ocean is apparently a small net carbon sink; for this reason, the biota of the ocean have been ignored in the global carbon budget (4). The marine biota I considered includes benthic submerged plants, both sea grasses and macroalgae, which are termed marine macrophytes. Grouping this collection of plants into a single category may seem arbitrary. Sea grasses are flowering plants which are ordinarily rooted in soft sediment and derive much of their nutrition from water in that sediment (7). Macroalgae primarily attach to hard substrata and derive their nutrition from the water column (8). Despite these differences I will discuss why these plant groups are more like one another than they are like other biological carbon sinks in the ocean.

Data from Whittaker and Likens (9) are presented in a format that is convenient for examining plant biomass, primary production, and mean turnover time, that is, the ratio of biomass to primary production (Fig. 1); each biospheric pool is treated as constant, with no change in either biomass or production. The annual release of fossil-fuel carbon and the cumulative release from 1860 to 1974 (1) are shown as production and biomass, respectively (Fig. 1).

Most terrestrial ecosystems (Fig. 1, A to N) have a relatively high biomass and a turnover time of years to decades. By contrast, most oceanic ecosystems (Fig. 1, O to S) exhibit a low biomass and a turnover time of less than a month, but there are two exceptions. Algal beds and reefs (Fig. 1, R) and estuaries (Fig. 1, S) have the highest biomass among the marine ecosystems, and each has a turnover time of about 1 year. Attached plants are important components in both of these macrophyte ecosystems.

The primary production of the macro-

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 $3.6 \times 10^8 \text{ km}^2$, whereas the macrophytes (Fig. 1, R plus S) occupy only about $2 \times 10^6 \text{ km}^2$ (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 1×10^9 tons of carbon per year. De Vooy (11) reviewed primary production in aquatic environments and derived a value of 0.8×10^9 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 4×10^9 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

Table 1. The CO_2 partial pressure of water impinging on coral reefs in the Houtman Abrolhos Islands, Western Australia.

Date	CO_2 partial pressure (μatm)	
	Coral transect	Algal transect
September 1979	162	114
March 1980	280	261

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,

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 macrophytes is slow due to carbon accumulation 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 rich 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 buffeted 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.

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References and Notes

1. C. D. Keeling, *Tellus* **25**, 174 (1973); R. M. Roty, in *The Fate of Fossil Fuel CO₂ in the Oceans*, N. R. Andersen and A. Malahoff, Eds. (Plenum, New York, 1977), p. 167.
2. H. Oeschger, U. Siegenthaler, U. Schotterer, A. Gugelmann, *Tellus* **27**, 168 (1975).
3. M. Stuiver, *Science* **199**, 253 (1978).
4. W. S. Broecker, T. Takahashi, H. J. Simpson, T.-H. Peng, *ibid.* **206**, 409 (1979).
5. B. Bolin, *ibid.* **196**, 613 (1977).
6. G. M. Woodwell et al., *ibid.* **99**, 141 (1978).
7. C. P. McRoy and R. J. Bardstate, *Limnol. Oceanogr.* **15**, 6 (1970); D. G. Patriquin, *Mar. Biol.* **15**, 35 (1972).
8. K. H. Mann, *Mem. Ist. Ital. Idrobiol. Dott Marco de Marchi Pallanza Italy* **29** (Suppl.), 353 (1972); K. H. Mann, *Science* **182**, 975 (1973); B. G. Hatcher, A. R. O. Chapman, K. H. Mann, *Mar. Biol.* **44**, 85 (1977).
9. R. H. Whittaker and G. E. Likens, *Brookhaven Symp. Biol.* **24**, 281 (1973).
10. E. Rasmussen, in *Seagrass Ecosystems—A Scientific Perspective*, C. P. McRoy and C. Helfrich, Eds. (Dekker, New York, 1977), p. 1.
11. C. G. N. De Vooy, in *The Global Carbon Cycle*, B. Bolin, E. T. Degens, S. Kempe, P. Ketner, Eds. (Wiley, New York, 1979), p. 259.
12. J. S. Bunt, in *Primary Productivity of the Biosphere*, H. Leith and R. H. Whittaker, Eds. (Springer-Verlag, New York, 1975), p. 169.
13. H. Kirkman and D. D. Reid, *Aquat. Bot.* **7**, 173 (1979); R. A. Congdon and A. J. McComb, *ibid.* **6**, 121 (1979).
14. B. C. Heezen, M. Ewing, R. J. Menzies, *Oikos* **6**, 170 (1955); D. B. Ericson, M. Ewing, B. C. Heezen, *Bull. Am. Assoc. Pet. Geol.* **36**, 489 (1952); R. M. Pratt, *Science* **138**, 492 (1962); R. J. Menzies and G. T. Rowe, *Int. Rev. Gesamten Hydrobiol.* **54**, 217 (1969); R. J. Menzies, J. S. Zaneveld, R. M. Pratt, *Deep-Sea Res.* **14**, 111 (1967); J. C. Ziemann, G. W. Thayer, M. B. Rob-

blee, R. T. Ziemann, in *Ecological Processes in Coastal and Marine Systems*, R. S. Livingston, Ed. (Plenum, New York, 1979); C. E. ZoBell, in *The Biology of Giant Kelp Beds (Macrocystis) in California*, W. J. North, Ed. (Cramer, Lehre, Germany, 1971), p. 269.

15. K. E. Chave, *J. Geol. Educ.* **15**, 200 (1967); L. S. Land, *J. Sediment. Petrol.* **40**, 1361 (1970); D. S. Patriquin, *ibid.* **42**, 687 (1972); S. V. Smith, *Limnol. Oceanogr.* **17**, 28 (1972); R. R. Seapy and M. M. Littler, *Pac. Sci.* **32**, 293 (1975).
16. T. Wolff, *Sarsia* **54**, 117 (1979).
17. R. A. Berner, *Am. J. Sci.* **264**, 1 (1966); B. J. Presley and T. R. Kaplan, *Geochim. Cosmochim. Acta* **32**, 1037 (1968); J.-L. Tsou, D. E. Hammond, R. M. Horowitz, in *Initial Reports of the Deep Sea Drilling Project*, B. C. Heezen

et al., Eds. (Government Printing Office, Washington, D.C., 1973), vol. 20, p. 861; T. Takahashi, L. A. Prince, L. J. Felice, in *ibid.*, p. 865.

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Fate and Transport of Dieldrin in Coralville Reservoir: Residues in Fish and Water Following a Pesticide Ban

Abstract. A model for the fate and transport of pesticide has been calibrated with field data from 1968 to 1978. The results have aided in a management decision to lift a commercial fishing ban in Coralville Reservoir, eastern Iowa. Dieldrin residues in fish, sediment, and water are all declining at about 15 percent per year. Approximately 50 percent of the pesticide load is exported from the reservoir in the outflow, 40 percent undergoes sedimentation, and 10 percent enters the fish.

Several models have been proposed to assess the fate and transport of pesticides in surface waters, including the Stanford Research Institute (SRI) model (1) and the Exposure Analysis Modeling System (EXAMS) (2). Previous models have not been extensively verified with field data, and they have not combined chemical fate modeling with biological effects. In this report, a model of pesti-

cide fate and bioconcentration is developed and applied to Coralville Reservoir, eastern Iowa, to assess the long-term fate and effects of dieldrin in the ecosystem. Results have aided the Iowa Conservation Commission in their decision to lift a commercial fishing ban.

The insecticide most widely used for corn rootworm and cutworm during the 1960's and 1970's was the chlorinated hydrocarbon aldrin, a derivative of hexachlorocyclopentadiene. Aldrin is rapidly converted to its epoxide dieldrin by biodegradation in natural waters (3). Because of the widespread use of aldrin in Iowa as an insecticide for corn, Iowa waters have become contaminated with dieldrin.

During the middle 1960's aldrin usage in Iowa reached a peak application of 6.5 million pounds (2.9 million kilograms) per year over a total of approximately 5 million acres (2 million hectares). Insect resistance to this chlorinated hydrocarbon became prevalent in the middle to late 1960's and swept from west to east across Iowa, and usage declined to approximately 3.5 million pounds (1.6 million kilograms) per year. In 1975 the Environmental Protection Agency (EPA) cancelled the registration of aldrin and dieldrin insecticides.

Public attention in Iowa to the problems of aldrin and dieldrin reached a peak in 1976, when the EPA and the Iowa Conservation Commission (ICC) placed a ban on commercial fishing in Coralville Reservoir. The primary commercial species affected by the ban was bigmouth buffalo, a bottom-feeding fish. In 1975 dieldrin residues in several big-

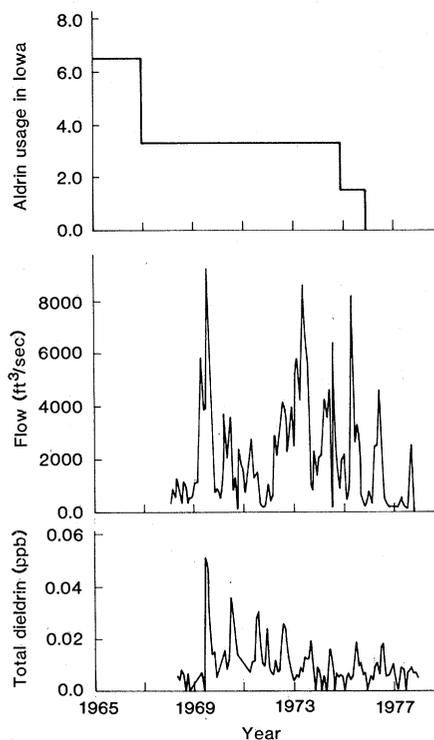


Fig. 1. Estimated aldrin usage (in million pounds) in Iowa and total dieldrin concentrations in the Iowa River at Iowa City, 1968 to 1978.