temperature pattern. The fit can be improved significantly by adjusting the phase relation at the onset (~A.D. 1100) and termination (~A.D. 1700) of the most recent clusters of radiocarbon anomalies (triplet episode) (*25*). This result is similar to that of Stuiver and Braziunas (*25*), who found that triplet episodes in the production rate ($\Delta Q/Q_t$) of ^{14}C were modeled differently than those of nontriplet intervals. Damon and Jirikowic (*20*) suggested that during the Hallstattzeit variations, Maunder-type anomalies may be exaggerated, which would lead to more extreme solar variation. Although the amplitude of the 50-year temperature trend during the triplet does not increase, the phase changes significantly at the beginning and end of the triplet. In the temperature reconstruction, the initial change of phase occurs at ~A.D. 1050 and corresponds to the timing of the peak of the first Maunder-type event in the triplet (*25*). This triplet interval is best fit by

$$\Delta T_{50} = 2.0 \cos\left(\frac{2\pi}{125}t + 0.96\pi\right) \quad (4)$$

This phase relation is terminated between A.D. 1650 and 1680, during the last Maunder-type event in the sequence. The phase change at the end of the triplet amounts to an additional 0.54π (~35 years). Feynman and Gabriel (*28*) found a similar change in phase in the 88-year Gleissberg cycle as the sun exited the Maunder minimum. They suggested that the sun exhibits periods of suppressed activity during which the phase of periodic variations is not conserved and that this behavior conforms to a chaotic solar dynamo. The appearance of phase changes in the 125-year period at the beginning and the end of the most recent set of major radiocarbon anomalies, as well as the maintenance of phase relations from ~A.D. 1050 to the termination of the prior triplet type event (~200 B.C.), supports this view and, in addition, suggests that the periodicities in the temperature reconstruction from the southern Sierra Nevada were solar-induced. The 600-year duration of anomalous phase (~A.D. 1050 to 1650) also corresponds to the average length of the Hallstattzeit variation (*19, 20*), whereas spectral peaks in the tree ring temperature reconstruction correspond to harmonics of the 2120-year recurrence interval between such events (*19, 20*). This comparison indicates that solar variability and climate may be closely related. In light of the findings (*25*) of significant power between 123 and 143 years in six of ten climate records (tree rings and glacier fluctuations) (*29*), it is likely that this solar-climate relation is at least marginally detectable globally.

REFERENCES AND NOTES

1. L. A. Scuderi, *Quat. Res. (NY)* **27**, 220 (1987); *ibid.* **34**, 66 (1990).
2. _____, *Nature* **325**, 242 (1987).
3. H. C. Fritts, *Tree Rings and Climate* (Academic Press, London, 1976).
4. W. Tranquillini, *Ecol. Stud.* **31**, 1–137 (1979).
5. K. R. Briffa *et al.*, *Nature* **346**, 434 (1990); K. R. Briffa *et al.*, *Clim. Dyn.* **7**, 111 (1992).
6. E. R. Cook, in *Methods of Dendrochronology*, E. R. Cook and L. A. Kairiukstis, Eds. (Kluwer Academic, Dordrecht, 1990), pp. 98–104.
7. D. A. Graybill, in *Climate from Tree Rings*, M. K. Hughes *et al.*, Eds. (Cambridge Univ. Press, Cambridge, 1982), pp. 21–30.
8. L. A. Scuderi, thesis, University of California, Los Angeles (1984).
9. The 38 series of ring widths that make up the chronology between A.D. 1 and 1981 range in length from 133 to 1183 years. The mean is 535 years; 70% of the series exceed 300 years in length, and 55% exceed 500 years. Chronology depth is 28 at A.D. 1800 and 19 at A.D. 1600. Depth varies from 9 to 19 between A.D. 1022 and 1600 with a minimum depth of 10 between A.D. 1 and 1021. Mean interseries correlation, a measure of the strength of the common signal in the chronology, is 0.397.
10. H. C. Fritts, T. Blasing, B. Hayden, J. Kutzbach, *J. Appl. Meteorol.* **10**, 845 (1971); K. Briffa and E. R. Cook, in (*6*), pp. 240–247; F. Serre-Bachet and L. Tessier, in *ibid.*, pp. 247–258.
11. S. W. Running and J. C. Coughlan, *Ecol. Modelling* **42**, 125 (1988).
12. G. C. Jacoby and R. D'Arrigo, *Clim. Change* **14**, 39 (1989).
13. V. C. LaMarche, Jr., *Science* **183**, 1043 (1974).
14. H. C. Fritts, J. Guiot, G. A. Gordon, in (*6*), pp. 178–185; G. A. Gordon and S. K. LeDuc, in *Seventh Conference on Probability and Statistics in Atmospheric Science* (Session 8, American Meteorological Society, Boston, 1981), pp. 129–133.
15. H. C. Fritts, *Reconstructing Large-Scale Climatic Patterns from Tree-Ring Data* (Univ. of Arizona Press, Tucson, AZ, 1991).
16. C. U. Hammer, H. B. Clausen, W. Dansgaard, *Nature* **288**, 230 (1980); R. B. Stothers, *ibid.* **307**, 344 (1984).
17. L. G. Drew, *Tree-Ring Chronologies of Western America: II. California and Nevada* (Chronology Series 1, Laboratory of Tree-Ring Research, Univ. of Arizona, Tucson, AZ, 1972).
18. H. C. Fritts and J. M. Lough, *Clim. Change* **7**, 203 (1985).
19. P. E. Damon and J. L. Jirikowic, in *Rare Nuclear Processes: Proceedings of the 14th Europhysics Conference on Nuclear Physics*, Bratislava, Czechoslovakia, 22 to 26 October 1990, P. Povinec, Ed. (World Scientific, Rivers Edge, NJ, 1992), pp. 177–201.
20. _____, in *Radiocarbon After Four Decades: An Interdisciplinary Perspective*, R. E. Taylor, A. Long, R. S. Kra, Eds. (Springer-Verlag, New York, 1992), pp. 117–129.
21. C. P. Sonett and S. A. Finney, *Philos. Trans. R. Soc. London* **330**, 413 (1990).
22. J. A. Eddy, *Clim. Change* **1**, 173 (1977).
23. M. Stuiver, T. F. Braziunas, B. Becker, B. Kromer, *Quat. Res. (NY)* **35**, 1 (1991).
24. C. P. Sonett and H. E. Suess, *Nature* **307**, 141 (1984).
25. M. Stuiver and T. F. Braziunas, *ibid.* **338**, 405 (1989).
26. _____, in *Secular Solar and Geomagnetic Variations in the Last 10,000 Years*, F. R. Stephenson and A. W. Wolfendale, Eds. (Kluwer, Dordrecht, the Netherlands, 1988), pp. 245–266.
27. M. R. Attolini, M. Galli, T. Nanni, *ibid.*, pp. 49–68.
28. J. Feynman and S. B. Gabriel, *Sol. Phys.* **127**, 393 (1990).
29. F. Rothlisberger, *10000 Jahre Gletschergeschichte der Erde* (Sauerländer, Aarau, Switzerland, 1986).
30. I thank R. Kerr, J. Drake, and two anonymous reviewers for comments that improved the content and clarity of this manuscript. Additional thanks to the Mount Whitney Ranger District, Lone Pine, CA, for continued access to sample sites, and to Boston University, which provided support for this study.

28 September 1992; accepted 15 December 1992

Export of North American Ozone Pollution to the North Atlantic Ocean

David D. Parrish,* John S. Holloway,† Michael Trainer,
Paul C. Murphy,† Gerry L. Forbes, Fred C. Fehsenfeld†

Measurement of the levels of ozone and carbon monoxide (a tracer of anthropogenic pollution) at three surface sites on the Atlantic coast of Canada allow the estimation of the amount of ozone photochemically produced from anthropogenic precursors over North America and transported to the lower troposphere over the temperate North Atlantic Ocean. This amount is greater than that injected from the stratosphere, the primary natural source of ozone. This conclusion supports the contention that ozone derived from anthropogenic pollution has a hemisphere-wide effect at northern temperate latitudes.

Transport of pollutants from populated and industrialized continental areas affects the chemistry and radiation balance of the global atmosphere (*1*). A pollutant of particular interest is ozone (O_3), because its photolysis initiates the oxidizing processes in the atmosphere (*2, 3*). Thus, the transport of O_3 from source regions affects the oxidizing capacity of the troposphere in the receptor areas. Additionally, O_3 is an important greenhouse gas whose atmospheric trends are only poorly known (*4*). Tropospheric O_3 has both natural and anthropogenic sources. The primary natural source is injection from the stratosphere (*5, 6*). The anthropogenic source is photochemical production from precursors emitted by industrial and transportation combustion sources (*7*). To understand the budget, and hence the effects of O_3, it is important to quantify and to compare the magnitude of these two sources.

The heavily polluted eastern coast of

North America is a particularly large source of O_3 and its precursors that can be transported to the temperate North Atlantic Ocean (8). This transport may have a major impact on the tropospheric chemistry over the North Atlantic Ocean, and even over the downwind continental region of western Europe. There is evidence that the average O_3 level at the surface in western Europe has more than doubled since preindustrial times (9) and that O_3 levels have increased throughout the lower troposphere over Europe in the last 50 years (*10, 11*). Transport from North America may play a role in this increase, particularly in the free troposphere.

To evaluate the significance of this transport, we measured O_3 and carbon monoxide (CO) levels during the summer of 1991 at three sites on the Atlantic coast of Canada (Fig. 1). CO is an anthropogenic pollutant that is relatively unreactive (lifetime about 1 month in summer) and has been used as an effective tracer of anthropogenic pollution (*12*). The measured relations between the concentrations of O_3 and CO in the air advected over the North Atlantic, coupled with the inventoried emissions of CO in eastern North America, provide the basis to estimate the amount of O_3 exported from North America.

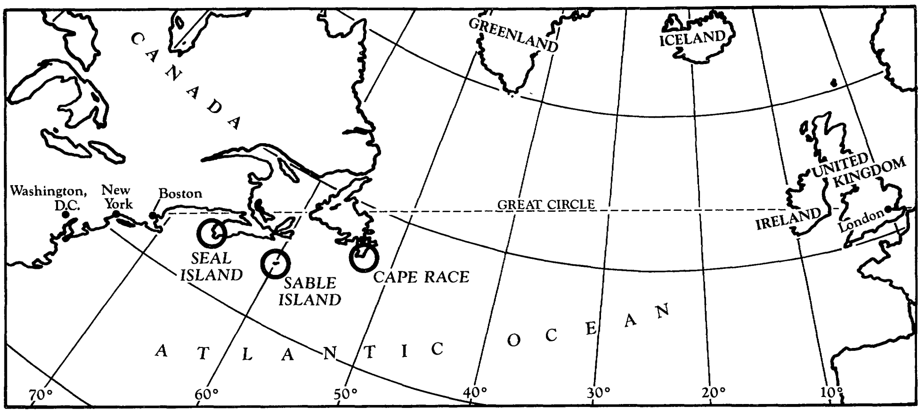

Fig. 1. Relation of the North Atlantic Ocean and the surrounding continents shown on a map projected on the great circle passing through Washington, D.C., and London. The three circles indicate the measurement sites.

During summer the atmospheric circulation in the temperate North Atlantic region is dominated by the Bermuda High; flow is typically to the northeast along the Atlantic Seaboard. The sites we sampled are spaced at approximately 500-km intervals downwind from the northeastern urban corridor of the United States and span approximately one-third of the distance from Boston to Ireland. Instruments to measure O_3 and CO levels [see (*13–15*) for experimental details] were placed in Canadian Coast Guard lighthouses on Seal Island, Nova Scotia (approximately 32 km east of Cape Sable, the southern tip of Nova Scotia), at Cape Race, Newfoundland, and in a Canadian Atmospheric Environment Service weather station on Sable Island, Nova Scotia. The sites are isolated from local pollution sources, with specific exceptions. Diesel generators located east and thus generally downwind of the instruments, power the installations on Seal and Sable islands, and limited vehicle traffic arrives at Cape Race. Influence from these sources was infrequent and easily identified from high variability over short time periods. Measurements from these periods have been removed from the data sets we discuss.

Measurements were made at all three sites from mid-July to mid-September, 1991. We calculated 1-min average O_3 and CO concentrations. Periods that showed high variability of either CO or O_3 levels were presumed to be contaminated by local sources and discarded. We then calculated 5-min averages and standard deviations. Any 5-min period with a standard deviation of CO or O_3 greater than 20 or 2 ppbv (parts per billion by volume), respectively, were also discarded. At no site were more than 4% of the data rejected by these two criteria. Winds were predominately from the direction of the local pollution source during periods of discarded data. Additional data were lost for some periods at each site as a result of equipment malfunction.

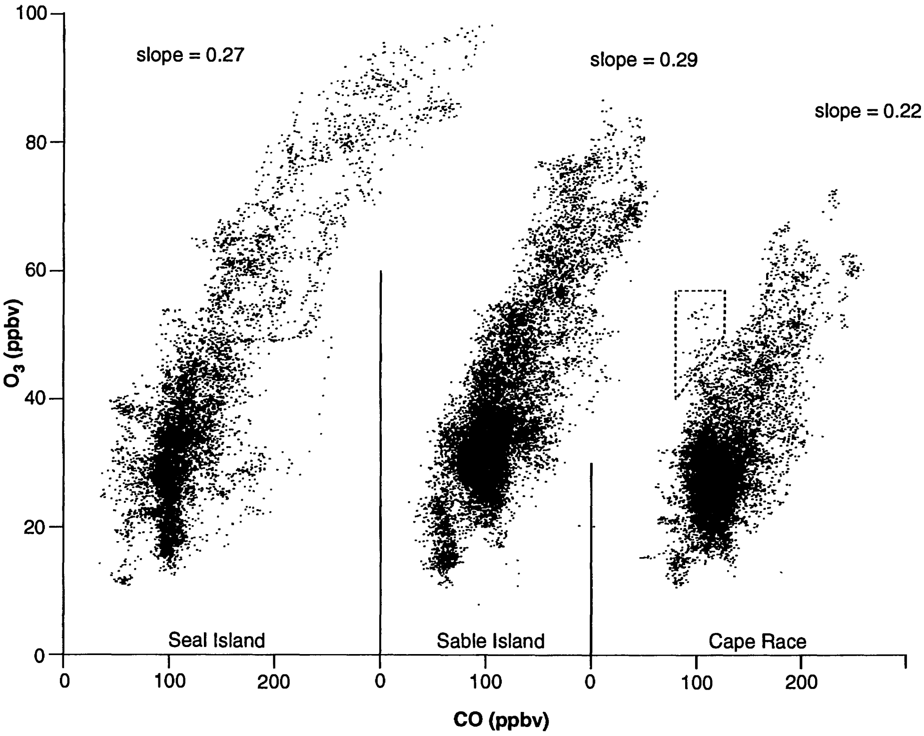

Fig. 2. Relation between O_3 and CO levels at each of the three sites. Each point indicates a 5-min average. For each site, the slope of the linear fit to the data is given; the standard errors in the slopes are all less than 0.01. The dashed line encloses the data points from the one pronounced episode of elevated O_3 attributed to stratospheric injection. At Seal Island the measurement period was 22 July to 22 September, at Sable Island it was 8 July to 21 September, and at Cape Race 13 July to 16 September, and the data sets represent 61%, 86% and 91% time coverage of the respective periods.

D. D. Parrish, J. S. Holloway, M. Trainer, P. C. Murphy, F. C. Fehsenfeld, Aeronomy Laboratory, National Oceanic and Atmospheric Administration, 325 Broadway, Boulder, CO 80303.

G. L. Forbes, Atmospheric Environment Service—Sable Island, 1496 Bedford Highway, Bedford, Nova Scotia B4A 1E5, Canada.

---

*To whom correspondence should be addressed.

†Also at Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, CO 80303.

At each of the three sites the data (Fig. 2) are characterized by generally constant, low levels of O_3 and CO interrupted frequently by episodes of correlated, elevated levels of both species. Linear regression analyses show that the variation of the data are similar at all three sites. The similarity of the derived slopes indicate that the relation between O_3 and CO levels is approximately constant over the 1000-km distance spanned by the sites.

The scatter in the data about the linear relation likely arises from several sources (in addition to experimental imprecision). One primary cause is that the ratio of O_3 to CO levels varied among episodes because of differences in the intensity of the photochemical processes that occurred in the respective air masses. For example, the air samples from Seal Island with the most elevated CO levels show relatively low O_3 levels. This pattern may be an indication of incomplete photochemical processing of the O_3 precursors in the air masses transported to this site, which is closest to the source regions. Indeed, air can be transported over the 500 km from Boston to Seal Island in less than 1 day under moderate wind conditions. Thus, some pollutants may be transported to Seal Island that have never been exposed to full sunlight.

Other factors must have also influenced O_3 levels. Injection of O_3 from the stratosphere would be indicated by episodes of elevated O_3 with low CO. Only one moderate episode was obvious. This was at Cape Race and is represented by the group of points between 40 and 60 ppbv O_3 with nearly constant CO levels at about 100 ppbv (Fig. 2). On 18 August 1991 in advance of the passage of Hurricane Bob up the Atlantic Coast, air with low levels of both O_3 and CO was transported into the region from the south. At Sable Island, O_3 and CO levels remained below 25 and 75 ppbv, respectively, for 36 hours. Such levels are characteristic of the tropical North Atlantic troposphere (*16*). The data from this episode correspond to the points in the lower left extreme of each data set of Fig. 2.

The slopes of the linear regressions in Fig. 2 provide an indication of the amount of O_3 photochemically produced from the precursors that were emitted into the air masses with the CO. That is, the amounts of O_3 and CO in these transported air masses were at a molar ratio of about 30%. Because the amount of CO emitted over North America is approximately known from emission inventories, and because the fraction of this CO that is transported to the North Atlantic can be estimated, it is possible to obtain an approximate budget for the amount of O_3 exported to the lower troposphere over the North Atlantic Ocean from North America.

A recent comprehensive inventory of emissions compiled for 1985 in North America (*17*) indicates that 40.4 million tons of CO per year was emitted from the states and provinces east of the Mississippi River; this amount is about two-thirds of the North American total. No significant seasonal cycle is present; therefore the CO emissions from eastern North America correspond to ~330 billion moles per summer. We assume that all of this CO is exported to the troposphere over the North Atlantic Ocean. Transport times across this region are a few days, much shorter than the 1-month lifetime of CO, and the prevailing winds are from the west. Thus, the quantity of O_3 photochemically produced from the precursors and transported to the North Atlantic is approximately 100 billion moles of O_3 per summer.

This simple budget estimate has significant uncertainties. First, the anthropogenic emissions of CO in eastern North America may be somewhat underestimated (*18–20*), and there are other, relatively minor sources of CO (*21*). Second, some of the exported O_3 may have been deposited or removed by chemical processes before it was measured. The relatively similar slopes found at the three sites suggest that loss of O_3 was not significant. Third, some of the CO emitted in eastern North America may be transported to other regions (for example, the Arctic or the Gulf of Mexico). Finally, the relation of O_3 to CO levels has been measured only at the surface and only to the northeast of the major source regions; at higher altitudes or when transport is more easterly or southeasterly, a somewhat different relation may hold.

Despite the uncertainties the O_3 export estimate is reasonable. The emissions inventory (*17*) estimates that 61 billion moles of nitrogen oxides per summer are emitted over eastern North America. Liu *et al.* (*7*) concluded that on average ten O_3 molecules are produced for each nitrogen oxide molecule emitted in this same region. The product of these numbers—610 billion moles—yields a rough estimate of the total O_3 produced in this region during the summer. Thus, the estimate of the export to the North Atlantic corresponds to 16% of the total produced. This is in reasonable accord with the exported fraction of 25% found in a model study (*22*) of one stagnant pollution period over eastern North America; the remainder was lost before transport by chemical reaction and surface deposition. Our export estimate is also in reasonable accord with a recent three-dimensional chemical transport model calculation of O_3 export from the United States (*23*), which found a total O_3 export about four times as large as our estimate. However, well over half of that total was ventilated vertically to the free troposphere over the western United States.

For comparison, the amount of O_3 injected from the stratosphere may be reasonably estimated from the annually averaged global flux of 5×10^{10} molecules per centimeter squared per second (*5, 6*). The flux is relatively low in the summer, but relatively high in the North Atlantic latitudes, and these factors will approximately cancel. The entire North Atlantic covers about 40×10^6 km^2, of which about 17×10^6 km^2 lie in the temperate region between Europe and North America. The above flux acting through the summer over this temperate region yields 110 billion moles, a quantity similar in magnitude to the estimate for continental transport, which is concentrated in this temperate region.

The stratospheric flux enters the upper troposphere, and a large fraction is removed by photochemical processes before it can be transported to the lower troposphere (*24*). The transport from the continent enters the lower troposphere. Therefore, during the summer, the O_3 budget in the lower troposphere over the temperate North Atlantic is dominated by O_3 that is photochemically produced from anthropogenic emissions. This conclusion is in accord with the good fit of the linear models to the data from the three sites.

In consideration of our results, the observed increase of tropospheric O_3 levels in Europe may be partially attributable to transport from North America. Likewise, transport from Europe could affect Asia and transport from Asia could affect North America, as has been indicated by measurements in California (*25*). The overall result would then be a general increase in the levels of O_3 in the temperate latitudes of the Northern Hemisphere. Such hemisphere-wide air pollution has been identified in satellite data (*26*) and is in accord with recent modeling studies (*27*).

REFERENCES AND NOTES

1. R. A. Duce *et al.*, *Global Biogeochem. Cyc.* **5**, 193 (1991).
2. J. A. Logan, M. J. Prather, S. C. Wofsy, M. B. McElroy, *J. Geophys. Res.* **86**, 7210 (1981).
3. A. M. Thompson, *Science* **256**, 1157 (1992).
4. R. T. Watson, H. Rodhe, H. Oeschger, U. Siegenthaler, in *Climate Change. The IPCC Scientific Assessment*, J. T. Houghton, G. J. Jenkins, J. J Ephraums, Eds. (Cambridge Univ. Press, Cambridge, 1990), pp. 1–40.
5. E. F. Danielsen and V. A. Mohnen, *J. Geophys. Res.* **82**, 5867 (1977).
6. J. D. Mahlman, H. Levy II, W. J. Moxin, *J. Atmos. Sci.* **37**, 655 (1980).
7. S. C. Liu *et al.*, *J. Geophys. Res.* **92**, 4191 (1987).
8. A. A. P. Pszenny, J. N. Galloway, R. S. Artz, J. F. Boatman, *Global Biogeochem. Cyc.* **4**, 121 (1990).
9. A. Volz and D. Kley, *Nature* **332**, 240 (1988).
10. J. A. Logan, *J. Geophys. Res.* **90**, 10,463 (1985).
11. J. Staehelin and W. Schmid, *Atmos. Environ.* **25A**, 1739 (1991).

12. J. Fishman and W. Seiler, *J. Geophys. Res.* **88**, 3662 (1983).
13. The O_3 instrument (Thermo Electron Corporation: Model 49) is based on ultraviolet absorption at 254 nm, and the CO instrument (Thermo Electron Corporation: Model 48) uses nondispersive infrared absorption. Each CO instrument was modified for improved sensitivity by adding a field lens (*14*) and more precise temperature control of the absorption cell. We checked the performance of each O_3 instrument twice daily by sampling zero air (ambient air passed through a charcoal trap) and an O_3 stream from the instrument's internal O_3 generator added both directly at the instrument and at the inlet. This latter procedure verified that the inlet line passed O_3 without loss. We measured the sensitivity of the CO instrument twice daily by addition of a 100 ppmv (parts per million by volume) CO calibration standard to the sampled air stream at a rate of 0.2% of the total flow. The zero of the CO instrument was determined hourly by sampling air passed through a heated catalyst bed that oxidized CO to CO_2. This procedure was used because water vapor absorbs in the same infrared region as CO and is thus an interferant in the CO measurement. Hence to correct for this interference, it is necessary that the water level be identical in the sampled ambient air and in the zero air. [See (*15*) for additional details.] During periods of rapid absolute humidity changes, we expect to have systematic errors in the measurements because the catalyst bed absorbs water, and it needs time (~30 min) to come into equilibrium with the ambient water vapor. However, in the marine boundary layer, where the present measurements were made, the water vapor is expected to be in near equilibrium with the sea surface temperature, and rapid water vapor changes are not expected. The instrument tests were performed automatically by computers that also collected, stored and telemetered the data back to the home laboratory by way of the GOES satellite.
14. R. R. Dickerson and A. C. Delany, *J. Atmos. Ocean. Tech.* **5**, 424 (1988).
15. D. D. Parrish, M. Trainer, M. P. Buhr, B. A. Watkins, F. C. Fehsenfeld, *J. Geophys. Res.* **96**, 9309 (1991).
16. S. R. Piotrowicz, C. J. Fischer, R. S. Artz, *Global Biogeochem. Cyc.* **4**, 215 (1990).
17. M. Saeger *et al.*, "The 1985 NAPAP emissions inventory (Version 2): Development of the annual data and modelers' tapes" (*EPA Rep. 600/7-89-012a*, U.S. Environmental Protection Agency, Washington, DC, 1989).
18. The estimate of the eastern North American source of CO may be low for two reasons. First, it has been suggested (*19*) that the emissions inventory for CO (*17*) has significantly underestimated the CO emissions. However, more recent results (*15, 20*) indicate that the error in the current inventory is likely to be small. Second, much of the CO emitted over western North America may also be transported over eastern North America and onto the North Atlantic.
19. W. R. Pierson, A. W. Gertler, R. L. Bradow, *J. Air Waste Manage. Assoc.* **40**, 1495 (1990).
20. M. P. Buhr, M. Trainer, D. D. Parrish, R. E. Sievers, F. C. Fehsenfeld, *Geophys. Res. Lett.* **19**, 1009 (1992).
21. CO is not only emitted directly, but it also is produced as a photochemical intermediate in the atmospheric oxidation of hydrocarbons emitted from anthropogenic and natural sources. This production of CO over eastern North America is difficult to quantify, but is expected to be considerably less than the direct emissions (*2*).
22. S. A. McKeen, E.-Y. Hsie, M. Trainer, R. Tallamraju, S. C. Liu, *J. Geophys. Res.* **96**, 10,809 (1991).
23. D. J. Jacob *et al.*, *ibid.*, in press.
24. S. C. Liu, D. Kley, M. McFarland, J. D. Mahlman, H. Levy II, *ibid.* **85**, 7546 (1980).
25. D. D. Parrish *et al.*, *ibid.* **97**, 15,883 (1992).
26. J. Fishman, K. Fakhruzzaman, B. Cros, D. Nganga, *Science* **252**, 1693 (1991).
27. A. M. Hough and R. G. Derwent, *Nature* **344**, 645 (1990).

28. The Atmospheric Environment Service of Environment Canada and the Canadian Coast Guard provided assistance and supporting data. This research has been funded as part of the Climate and Global Change program by the National Oceanic and Atmospheric Administration.

25 August 1992; accepted 24 December 1992

Fossilization of Soft Tissue in the Laboratory

Derek E. G. Briggs and Amanda J. Kear

Some of the most remarkable fossils preserve cellular details of soft tissues. In many of these, the tissues have been replaced by calcium phosphate. This process has been assumed to require elevated concentrations of phosphate in sediment pore waters. In decay experiments modern shrimps became partially mineralized in amorphous calcium phosphate, preserving cellular details of muscle tissue, particularly in a system closed to oxygen. The source for the formation of calcium phosphate was the shrimp itself. Mineralization, which was accompanied by a drop in pH, commenced within 2 weeks and increased in extent for at least 4 to 8 weeks. This mechanism halts the normal loss of detail of soft-tissue morphology before fossilization. Similar closed conditions would prevail where organisms are rapidly overgrown by microbial mats.

Fossils that preserve evidence of soft tissues yield a wealth of information over and above that provided by the normal remains of mineralized shells, bones, and teeth. Structural organic tissues such as cuticle may become incorporated into the fossil record as complex inert biopolymers. Tissues such as muscle, however, which are subject to rapid autolysis and metabolization by bacteria, usually can be preserved only by early authigenic mineralization (*1*, *2*). Occasionally such mineralization occurs quickly enough to preserve spectacular three-dimensional detail of muscles and other tissues and thus to give rise to some of the most remarkable fossils known. This type of preservation usually involves calcium phosphate (*3*). Recently studied examples have been found in the Santana Formation (Lower Cretaceous), Chapada do Araripe, Brazil (*4*–*6*), and the Cordillera de Domeyko (Upper Jurassic) of Chile (*7*). Here the striated muscle of both elasmobranchs and bony fishes displays details of the morphology of the sarcolemma, fibrils, sarcomeres, and the banding within them, and occasionally even rows of cell nuclei (*6*, *7*). Mineralization of fish soft tissue is a more widespread phenomenon than is generally realized, but many reports predate the availability of scanning electron microscopy (SEM), and the degree of detail that is preserved is unknown. Phosphatized muscle has also been reported in other groups, including pterosaurs from the Santana Formation (*8*), cephalopods from Solnhofen (Jurassic, Tithonian), Germany (*9*), and Christian Malford (Jurassic, Callovian), England (*10*), and crustaceans from Solnhofen and from within the gut of Santana fishes (*11*).

Attempts to understand the process of soft-tissue mineralization—the so-called "Medusa effect" (*5*)—and the conditions required to promote it have been based mainly on data from fossils. Experiments on mineralization in association with decaying shrimps (*1*) and fish (*12*) have resulted in the formation of calcium soaps, but not in extensive mineralization of soft tissue (*13*). Here we describe how the mineralization of soft tissue can be induced under experimental conditions in the laboratory, and we use the results to interpret its occurrence in the fossil record.

Freshly killed specimens of the shrimp *Crangon crangon* and the prawn *Palaemon* sp. (*14*) were placed in experimental vessels with 50 ml of standard artificial seawater (ASW) inoculated with water (50 ml/liter) and sediment (~0.5 ml/liter) from the Tay Estuary, Dundee, Scotland, and yeast extract (0.1 g/liter) (*15*, *16*). The vessels were incubated at 20° ± 0.5°C for 3 days to 25 weeks before sampling (*17*).

We carried out replicate experiments (five separate vessels, one carcass in each, for each sample point) under four different conditions (experiments A to D) to explore the mineralization process. At start-up, the O_2 in the water in experiments A, B, and C was at 50% saturation; in experiment D the water was deoxygenated (*18*). Initial decay reduced the O_2 concentration to just a few percent within 1 to 2 days. Rapid rediffusion of atmospheric O_2 was permitted in experiment A, which was not sealed; in experiment B the jar was fitted with a cap that permitted slow diffusion only; in experiments C and D the systems were sealed with anaerocult, which removes gaseous O_2

Department of Geology, University of Bristol, Wills Memorial Building, Queen's Road, Bristol BS8 1RJ, United Kingdom.