Foraminifera and Chlorophyll Maximum: Vertical Distribution, Seasonal Succession, and Paleoceanographic Significance

Richard G. Fairbanks; Peter H. Wiebe


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would appear that conditions are favorable for the transport of significant amounts of dust to the Enewetak region for 3 or 4 months of the year, beginning perhaps in February or March. If our measured atmospheric deposition rate of 4 \( \mu g \text{ cm}^{-2} \text{ month}^{-1} \) applies over a 3- to 4-month period, it would result in an accumulation rate of about 0.3 mm of wet sediment per 1000 years. This assumes an in situ sediment density (the ratio of the dry weight of sediment to the in situ volume occupied by the sediment) of 0.5 g cm\(^{-3}\) (9). The actual nonbiogenic sedimentation rates to the ocean floor in this region of the North Pacific are not well known but are probably about 1 mm per 1000 years (9, 10). We would expect that the atmospheric contribution of dust to Pacific deep-sea sediments would be even greater in the higher latitudes where transport conditions from Asia are more favorable.

R. A. Duce
C. K. Unni
B. J. Ray

Graduate School of Oceanography,
University of Rhode Island,
Kingston 02881

J. M. Prospero
J. T. Merrill

Rosenstiel School of Marine and
Atmospheric Sciences, University of
Miami, Miami, Florida 33149

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Ortner and Wiebe concluded that zooplankton biomass was enhanced about the DCM. These results confirm the conclusions of Hobson and Lorenzen (6) that the depth of the DCM in the North Atlantic and the Gulf of Mexico is influenced by the depth of the local pycnocline. Plant cells of the DCM utilize nutrients diffusing into nutrient-poor surface water (7). The flux of nutrients to shallow, nutrient-depleted surface waters is greatest in the pycnocline. The maximum depth to which the DCM follows the seasonal pycnocline is regulated by the minimum light intensity necessary for the growth of the particular phytoplankton taxa inhabiting the area. Chlorophyll accumulates in many oceanic environments to which 1 percent of the ambient light penetrates (7, 8).

Many species of planktonic foraminifera are markedly more abundant in the DCM than in surrounding waters, an indication of an association between foraminifera, other zooplankton, and algal cells of the DCM (Fig. 1) (9, 10). Apparently the DCM is a major food source zone, which is exploited by foraminifera. In addition, planktonic foraminiferal species have specific temperature tolerance ranges (11), and therefore the seasonal pycnocline can provide an ecologic niche for multiple species.

The results obtained from cruises Chain 125 and Knorr 53 point to an important correlation, which is amenable to modeling paleoceanographic conditions by means of transfer functions or algebraically less abstract methods (12). If temperature is a major factor regulating foraminiferal species composition (11), then for a large portion of the year the temperature at the DCM depth is the most useful predictor of foraminiferal vertical and seasonal distribution and oxygen-isotopic composition. In addition, it follows that the level of chlorophyll and the zooplankton concentration are related to the standing stock of foraminifera.

A DCM develops seasonally in the pycnocline in the northern Sargasso Sea and the slope water, but the DCM concentration is greater and its duration is shorter in the slope water than in the Sargasso Sea. The DCM is near or at the surface in the winter months and submerges from May to September in the slope water and from April through November in the Sargasso Sea. The DCM reaches a maximum depth in the slope water of 75 m in September and 85 m during July in the Sargasso Sea (4, 13) (Fig. 2). The annual temperature range at the DCM is attenuated by 66 percent in the slope water and 57 percent in the Sargasso Sea as compared to the overlying sea surface (Fig. 2). Slope water surface temperatures range from 9.8°C in April to 24.2°C in August as compared to a range of 9.8°C in April to 19.4°C at the end of October for temperatures at the DCM; the maximum difference between sea surface and DCM temperatures is 10°C in July. Sargasso Sea surface temperatures range from 18°C in March and April to 28°C in August, whereas the temperature at the DCM ranges from 18°C between March and April to 23.6°C in October. Here the maximum temperature difference is approximately 6°C in August. Winter and early spring surface temperatures are coincident with temperatures at the DCM. However, the sea surface maximum temperature leads the maximum annual temperature of the DCM by 3 months in the slope water and 2 months in the Sargasso Sea.

Transfer function paleotemperature predictions based on seabed foraminiferal samples (12) may be less accurate in regions where the DCM temperature departs significantly from the surface temperature. In many regions of the world’s oceans, the DCM occurs in the well-mixed layer during part of the year. Where sufficient geographic and seasonal data on chlorophyll concentrations in the water column are lacking, the temperature at the DCM can be mapped as the temperature at the top of the pycnocline. This level can be approximated as the stability maximum in the photic zone (about 0 to 80 m) where $\rho \sigma = Z$ is greatest ($Z$ is the depth, and $\rho \sigma$ is the specific gravity of water).

Where chlorophyll concentrations are high and relatively constant throughout the euphotic zone as, for example, in MOC-1-28 (Fig. 1a), spinoce, nonspinoce, or symbiont-bearing and symbiont-barren, foraminifera species show no depth differentiation within the euphotic zone (9, 10). The association of the DCM with the pycnocline may be the reason why earlier investigators related the depth distributions of foraminifera species to the density of seawater (14).

RICHARD G. FAIRBANKS
Lamont-Doherty Geological Observatory of Columbia University,
Palisades, New York 10964

PETER H. WIEBE
Woods Hole Oceanographic Institution,
Woods Hole, Massachusetts 02543

References and Notes
A Comparison of Thermal Observations of Mount St. Helens Before and During the First Week of the Initial 1980 Eruption

Abstract. Before and during the first week of the March–April 1980 eruptions of Mount St. Helens, Washington, infrared thermal surveys were conducted to monitor the thermal activity of the volcano. The purpose was to determine if an increase in thermal activity had taken place since an earlier airborne survey in 1966. Nine months before the eruption there was no evidence of an increase in thermal activity. The survey during the first week of the 1980 eruptions indicated that little or no change in thermal activity had taken place up to 4 April. Temperatures of ejected ash and steam were low and never exceeded 13°C directly above the vent.

Mount St. Helens, which is located in southwestern Washington (45°12′N, 122°11′W), started to erupt on 27 March 1980 after a dormant period of 124 years. To study this recent sequence of eruptions, we established a research site 9 km northwest of the summit at an elevation of 840 m. During the time period between 30 March and 4 April 1980, we monitored the thermal and seismic activity associated with the volcanic eruptions. We conducted the thermal monitoring by using an infrared scanner and an infrared thermometer (1), as a follow-up investigation to a survey conducted during a 5-day period in August 1979.

Our ground-based infrared survey of Mount St. Helens during the summer of 1979 was prompted by reports (2, 3) that this volcano was possibly the most active and most violent of the volcanoes in the coterminous United States. Mount St. Helens is younger than its three neighboring peaks, Mount Rainier, Mount Adams, and Mount Hood (2, 3). Although its existence can be traced back some 37,000 years, nearly all of the present cone was formed in the last 2500 years. Radiocarbon dates indicate that the dormant period has not exceeded 500 years; more recently, the dormant periods have lasted only 100 to 200 years. The frequency and volume of eruptions produced by Mount St. Helens are similar to those of Mount Vesuvius, Mount Fuji, and Mount Hekla. However, eruptions of Mount St. Helens have never produced as much material as the eruption of Mount Mazama (Crater Lake, Oregon) 6600 years ago.

Because of the possibility that Mount St. Helens might erupt before the end of this century (2), we included it, along with Mount Shasta and Mount Lassen (located in California), in ground-based infrared surveys during August and September 1979. The purpose of our surveys was to document changes in existing hot areas on each mountain and to look for any new hot areas that might have appeared since the 1966 infrared thermal survey of the mountain (4). The survey during the summer of 1979 seemed particularly timely since interest in Cascade volcanoes was renewed after reports of increased heat flow from Mount Baker (5), 290 km north of Mount St. Helens.

The infrared scanner that we used in our initial survey of Mount St. Helens (August 1979) and during the first week of the 1980 eruptive activity (30 March to 4 April 1980) produces a visual image on a cathode-ray tube (Figs. 1 and 2) of the infrared radiation emitted from the object observed. The infrared scanner quantifies the temperature differences on...