explain juxtaposed areas of manganese-poor orange varnish (from rock undersides and closed cracks) and manganese-rich black varnish (from rock surfaces and open cracks), the physicochemical model requires that water saturation and mobilization of manganese from the orange to the black varnish alternate with oxidizing conditions to fix the manganese. Under these conditions, swelling minerals would be modified to chlorite (*16*), but varnish clay minerals are composed of illite and montmorillonite (*1*), not chlorite. Our biological model associates orange varnishes with microenvironments where alkaline dust or soil is in contact with rock surfaces and causes them to be inhospitable to the manganese-concentrating microbes that are active in near-neutral *p*H environments, where black varnish forms. Other difficulties with the physicochemical model of varnish formation include the widely documented presence of iron-manganese films in nonarid environments and the noticeable absence of manganese-rich varnish around lichens on desert rocks, where the lichens generate the required *E*h-*p*H changes.

It appears that the phenomenon of desert varnish may be added to the expanding list of manganese concentrations that have been found to have a microbial origin. The irregular onset of bacterial colonization accounts for the puzzling inconsistency in varnish development from stone to stone on desert pavements. Thus it is hazardous for archeologists to use the degree of varnish development as the sole criterion to estimate the relative ages of individual stone artifacts in lithic assemblages.

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Oxygen-18 Enrichment of Planktonic Foraminifera Due to Gametogenic Calcification Below the Euphotic Zone

Abstract. *Empty shells of spinose planktonic foraminifera on the seabed are significantly enriched in oxygen-18 as compared with the shells of their living counterparts in surface waters. This enrichment is due to gametogenic calcification, which extracts calcium carbonate from deeper and colder waters as the shell sinks below the euphotic zone.*

Emiliani and subsequent investigators have amply demonstrated the usefulness of the oxygen isotopic composition ($\delta^{18}O$) of planktonic foraminifera as a tool for deciphering Cenozoic paleoenvironmental conditions (*1*). Yet the fundamental problem of whether these organisms secrete $CaCO_3$ in isotopic equilibrium with ambient water is still a matter of debate. Two different approaches have been used to estimate the extent to which isotopic equilibrium is achieved in planktonic foraminifera. The results obtained from the use of these two procedures have been contradictory.

The first approach, based on the study of well-dated Holocene surface sediments, yielded $\delta^{18}O$ values that were interpreted as indicating that the shell $CaCO_3$ was in isotopic equilibrium or close to isotopic equilibrium with surface-water conditions (*2–5*). These investigators assumed that the relationship between foraminiferal shell $\delta^{18}O$, seawater $\delta^{18}O$, and water temperature was the same as that for mollusks (*6*). The second approach is based on the isotopic analysis of plankton-tow specimens (*7, 8*). A plot of isotopic values versus temperature for most spinose planktonic foraminifera shows that the $CaCO_3$ of their shells is not in isotopic equilibrium with surface-water conditions but is slightly impoverished in ^{18}O when examined in light of the paleotemperature scale of Epstein *et al.* (*6*).

We recently studied the $\delta^{18}O$ variations in two common tropical spinose planktonic foraminiferal species, *Globigerinoides ruber* and *G. sacculifer*, in the upper 200 m of water of the subtropical and tropical Indian Ocean (*9*). With these measurements it was possible to evaluate the ^{18}O fractionation between foraminiferal carbonate and seawater as a function of temperature, because seawater temperature variations in this region are small and the life-span of *G. sacculifer* is on the order of a few weeks (*10*). The results (Fig. 1) show an excellent correlation between the calcite isotopic enrichment ($\delta^{18}_{foram} - \delta^{18}_{water}$) and the temperature of the surficial mixed layer (*11*). These data indicate that most of the calcite secretion occurred in this mixed layer (*12*). This result, which is consistent with the model of Fairbanks and Wiebe (*8*), suggests that foraminifera grow in the chlorophyll maximum since its depth is governed by the noticeable density increase which occurs at the base of the surficial mixed layer. Our results indicate that *G. ruber* and *G. sacculifer* could preferentially develop in the upper part of the chlorophyll maximum. Moreover, the $\delta^{18}O$ values of *G. ruber* and *G. sacculifer* shells are both isotopically lighter than equilibrium values by about 0.6 per mil (*6*), and the departures from equilibrium remain constant in the temperature range from 20° to 31°C.

Globigerinoides ruber shells with a diameter smaller than 250 μm have the

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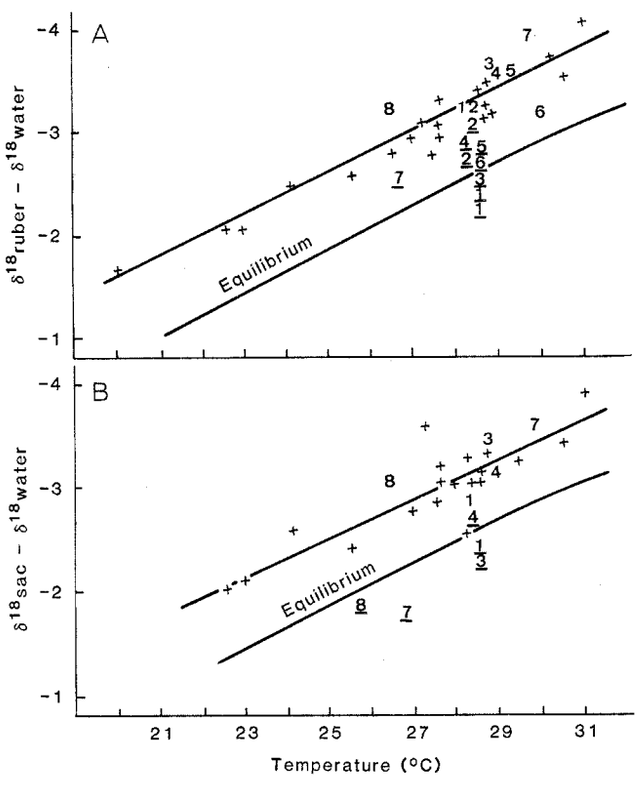
same low $\delta^{18}O$ value as larger shells. Similarly, no difference was observed between shells collected within and below the mixed layer (*9*). By contrast, *G. sacculifer* shells collected below the mixed layer have a $\delta^{18}O$ value consistently heavier by about 0.2 per mil than those taken within the mixed layer, an indication that this species continues to secrete a certain amount of $CaCO_3$ below the mixed layer. Our results thus indicate that, prior to reproduction time, the shell secretion of *G. ruber* occurs entirely within the mixed layer and that of *G. sacculifer* occurs mainly within the mixed layer. Our results confirm the isotopic disequilibrium observed in plankton-tow samples (*7, 8*), but they conflict with the data from core tops which are consistent with the idea that the planktonic foraminifera secrete their shells in isotopic equilibrium with ambient seawater.

In an effort to resolve this conflict, we compared the $\delta^{18}O$ values of shells of these two species in plankton tows from the upper 200 m of water with the corresponding values of their empty shells in the underlying sediment. We observed significant $\delta^{18}O$ mean enrichments of about 0.78 and 0.92 per mil in the empty shells of *G. ruber* and *G. sacculifer*, respectively, as compared with the values for their shells from the upper 200 m (Fig. 1). We checked the Holocene age of each core top by analyzing the foraminiferal shells down-core to the last glacial maximum (~ 18,000 years before the present). Figure 2 displays some typical records of analyzed core-top sediments in the tropical Indian Ocean, where the Holocene $\delta^{18}O$ values are significantly heavier than the values for the plankton samples (*13*). Moreover, for most stations these $\delta^{18}O$ values are significantly heavier than those for calcite if secreted in surface waters during the coolest or highest salinity season (*14*). This result demonstrates that the observed ^{18}O enrichment cannot be due only to selective dissolution of thin-walled specimens or specimens which secreted their shells in surface water during the warmest season (*5, 15*). Mixing of glacial with postglacial Holocene shells by bioturbation within the sediments can also be ruled out as an explanation for these high $\delta^{18}O$ values, since bioturbation in the sediment has a mixing depth of about 15 cm (*16*) and we observed the presence of ^{18}O-enriched shells in core tops with a Holocene layer thicker than 1 m (Fig. 2).

One possible explanation for these results is that the *G. ruber* and *G. sacculifer* populations are constituted of thin-walled individuals secreting their shells out of isotopic equilibrium and of thicker-walled specimens secreting their shells essentially in isotopic equilibrium with surface waters, and that only the thicker-walled shells are preserved in the sediment. Such a model would have two consequences:

1) The excellent correlation between the measured $\delta^{18}O$ values and the mixed-



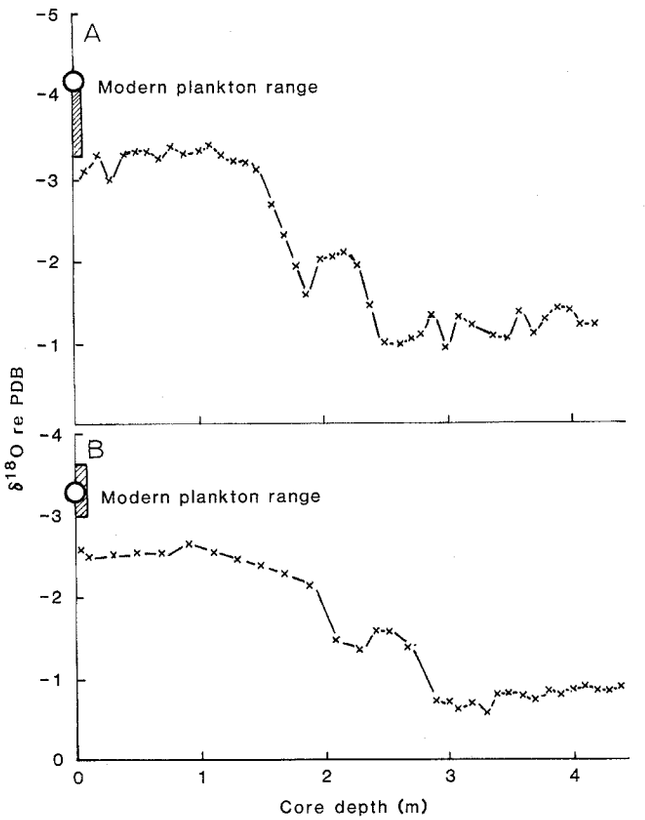


Fig. 1 (left). Variation of the oxygen isotopic enrichment between shell calcite and water versus temperature for living specimens of (A) *Globigerinoides ruber* and (B) *Globigerinoides sacculifer* in the tropical and subtropical Indian Ocean (*9*). The best-fit lines are $\delta^{18}_{ruber} - \delta^{18}_{water} = 2.55 - 0.20T$ ($r = .94$) and $\delta^{18}_{sacculifer} - \delta^{18}_{water} = 2.27 - 0.19T$ ($r = .87$). Crosses indicate values for living specimens without comparison with fossil specimens. Digits without underlining indicate values for living specimens from plankton tows versus the measured temperature. Underlined digits represent values for Holocene fossil specimens versus the mean annual temperature. Source locations are as follows: 1, 12°05.4′N, 73°54.0′E; 2, 13°07.6′N, 73°18.6′E; 3, 14°59.8′N, 72°19.8′E; 4, 15°31.8′N, 72°34.1′E; 5, 10°28′N, 75°14′E; 6, 11°29.7′N, 74°32.2′E; 7, 19°13.3′N, 60°40.9′E; and 8, 20°41.9′N, 59°34.1′E. The isotopic "equilibrium" line based on mollusks corresponds to the paleotemperature scale of Epstein *et al.* (*6*): $T = 16.5 - 4.3(\delta_c - \delta_w) + 0.14(\delta_c - \delta_w)^2$ where $\delta_c = \delta^{18}O_{calcite}$. Fig. 2 (right). Oxygen-18 isotopic composition (relative to the Pee Dee belemnite standard (PDB) of living *G. ruber* as compared to that of fossil shells of this species in Holocene sediment in two cores from the northern Indian Ocean: (A) core MD 77171: 11°45.6′N, 94°09′E (station MD 13-36); (B) core MD 77194: 10°28′N, 75°14′E (station MD 13-59). The circles on the *y*-axes indicate the measured δ values of planktonic foraminifera collected in plankton tows during the cruise.

layer temperature from our continuous sampling of plankton in the upper 200 m of water in the northern Indian Ocean during two periods of about 30 days (Fig. 1) could have been obtained only if we had always collected and analyzed samples with the same percentages of thin-walled and thick-walled shells. This seems highly unlikely.

2) If two populations existed within a single species, one with thin-walled shells and another with thick-walled shells, we would expect a bimodal distribution of their wall thickness. Quantitative measurements by Bé (*17*) demonstrated that *G. sacculifer* has a unimodal distribution of wall thickness when it occurs in the water column or surface sediment. We thus reject this hypothesis, which is not supported by field observations.

Another possible explanation for the observed enrichment in ^{18}O of *G. ruber* and *G. sacculifer* in the sediment as compared to living specimens in epipelagic waters is that these organisms continue to calcify their shells below 200 m during a late stage of their life cycle. Independent observations from laboratory-cultured *G. sacculifer* indicate that this species secretes additional $CaCO_3$ during gametogenesis for a period of about 14 to 16 hours prior to full gamete release (*17*). This gametogenic calcification deposits calcite over the last-formed chambers, with the layer varying from a thin veneer to a "calcite crust" up to 16 μm thick.

Preliminary analyses of cultured *G. sacculifer* indicate that the amount of calcite secreted during gametogenesis is sufficient to alter significantly the $\delta^{18}O$ values of the adult shells (*18*). Moreover, the fact that gametogenesis in spinose planktonic foraminiferal species in laboratory culture is preceded invariably by the sinking of the organism and spine loss (*19*) provides strong support for our hypothesis that gametogenic calcification in the ocean would take place at considerable depth below the euphotic zone. The spine shedding or nuclear division during gametogenesis, or both, results in the loss of the buoyancy mechanism, and thus the organism would inevitably sink from the epipelagic zone. We estimate that an adult *G. sacculifer* with a shell length of 500 μm and an initial weight of 20 μg would increase to 27 μg through gametogenic calcification and descend to an estimated depth of 450 m in 12 hours (*20*).

Gametogenic calcification during shell sinking to mesopelagic depth would extract $CaCO_3$ from waters considerably cooler than the epipelagic waters in which the foraminifera form the bulk of their shells. We calculate on the basis of shell size and wall thickness that about 18 percent of the total calcite of a *G. sacculifer* shell in surface sediment is secreted below the euphotic zone (*21*). We have also compared the mean weights of pregametogenic and postgametogenic shells of *G. sacculifer* and concluded that gametogenic calcification accounts for 22 percent, on the average, of the sediment shell weight, in good agreement with the preceding estimate (*22*).

Since our isotopic measurements show that the mean $\delta^{18}O$ value of *G. sacculifer* shell from the surface water is −3.11 per mil but −2.19 per mil from the sediment (*17*), we calculate that the 18 to 22 percent of the calcite secreted at depth has $\delta^{18}O$ values ranging from +2 to +1.07, indicating a mean temperature of secretion of 9° to 12.5°C, or mean depths ranging from 300 to 800 m in the northern Indian Ocean, assuming that gametogenic calcification occurs in isotopic equilibrium (*6*) in water of δ_w = +0.3 per mil (δ_w = $\delta^{18}O$ of seawater) relative to standard mean ocean water (SMOW). These calculated δ estimates are extremely sensitive to the ratio of gametogenic to pregametogenic calcite. A more accurate estimate for the gametogenic calcification depth can be obtained through better measurements of this ratio, either in laboratory culture or in sediment trap samples.

According to our model, larger individuals would settle faster during gametogenesis and thus would be encrusted with calcite secreted at lower temperatures (greater depth) than smaller individuals. For instance, a shell with a diameter of 800 μm and a weight of 82 μg would descend to 630 m in 12 hours (*20*), 180 m deeper than a 500-μm shell. Since no isotopic variations were found among the larger and smaller shells of the foraminifera collected from plankton tows, it is expected that larger individuals would be richer in ^{18}O than smaller individuals in the sediment assemblage. The magnitude of this effect will depend on various factors, such as the temperature gradient in the water column, the amount of calcite secreted at various depths, and also the effects of dissolution within the shells (*23*). Despite these possible biases, analyses of six size fractions in the 14 most common foraminiferal species in box core ERDC-92 (*24*) show that, as a general rule, the larger shells within each species (including *G. ruber* and *G. sacculifer* in surface sediment) tend to be enriched in ^{18}O. This result provides strong support for our model.

We therefore conclude that gametogenic calcification by planktonic foraminifera during their descent below the euphotic zone provides an explanation for the ^{18}O enrichment of fossil shells relative to their living counterparts in surficial waters. This isotopic enrichment balances in part the disequilibrium observed in planktonic foraminifera collected in plankton tows, but it is only by chance that the $\delta^{18}O$ values of planktonic foraminiferal shells in surface sediments are sometimes found to be close to isotopic equilibrium with surface-water conditions. In future paleoceanographic reconstructions with oxygen isotopes, investigators should thus take into account that gametogenic calcification introduces a bias in the $\delta^{18}O$ values, the amplitude of which may depend on the dissolution rate of the sediment.

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$$\delta^{18}O = \left[\frac{(^{18}O/^{16}O)_{sample}}{(^{18}O/^{16}O)_{standard}} - 1 \right] \times 10^3$$

relative to the Pee Dee belemnite (PDB) standard.
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12. The slope of the regression line of $\delta^{18}O$ versus the mixed-layer temperature is −0.20 for *G. ruber* and −0.19 for *G. sacculifer*, almost exactly the value predicted by Epstein *et al.* (*6*) in this temperature range, taking into account the quadratic term of the temperature equation. The correlation coefficients (r) are also at their highest values (Fig. 1). If the regression is made between $\delta^{18}O$ and the temperature (T) at 50 m, the correlation coefficients and the slopes of the regression lines decrease noticeably; the regression equations are $\delta^{18}_{ruber} - \delta^{18}_{water} = 0.97 - 0.16T$ ($r = .72$) and $\delta^{18}_{sacculifer} - \delta^{18}_{water} = 0.25 - 0.12T$ ($r = .61$). This poorer correlation is due to the fact that at those stations where the surficial mixed layer is shallower than 50 m, the temperature at 50 m is significantly lower than that of the mixed layer whereas the foraminiferal $\delta^{18}O$ value is similar to that of calcite secreted in warm waters. This result shows unambiguously that secretion occurs at (or close to) the temperature of the mixed layer and excludes the hypothesis that calcite is deposited at a constant depth (as shallow as 50 m), determined, for instance, by the extent of light penetration in the ocean.
13. In our measurements, the $\delta^{18}O$ values of foraminiferal shells from sediment samples are heavier than those of plankton samples by 0.78 per mil for *G. ruber* and by 0.92 per mil for *G. sacculifer*. A t-test performed on these differences gave t values of 8.79 for *G. ruber* and 10.16 for *G. sacculifer*. These high t values indicate that the mean differences are statistically different from zero, even at the confidence level of .995.
14. Mean values of $\delta^{18}O$ for foraminiferal shells from sediment samples are heavier than the calculated heaviest $\delta^{18}O$ values of shells secreted in surface water by 0.33 per mil for *G. ruber* and 0.60 per mil for *G. sacculifer*. A t-test performed on these differences gave t values of 4.77 for *G. ruber* and 11.73 for *G. sacculifer*. These high t values indicate that the mean $\delta^{18}O$ differences are statistically different from zero, even at the .995 confidence level.
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$$W = \frac{8}{3}\frac{\nu}{x}\left[-1 + \left(1 + \frac{1}{24}\frac{\rho_p\,\rho_E}{\rho_E}\,g\,\frac{x^3}{\nu^2}\right)^{1/2}\right]$$

where w is the speed of the particle; x is its mean diameter; ρ_p and ρ_E are the mass of a unit volume of the particle and seawater, respectively; g is the gravitational acceleration; and ν is the seawater kinematic viscosity [J. C. Brun-Cottan, thesis, University of Paris (1976)]. Numerical values to be used are as follows: $\nu = 1.5 \times 10^{-2}$ cm^2 sec^{-1}; $g = 981$ cm sec^{-2}; $\rho_E = 1.03$; $\rho_p \approx 1.2$ (this low value reflects the occurrence of water which fills pores and the inside of the shell). The diameter x must be given in centimeters, and w has the dimensions centimeters per second.
21. A comparison of the size of *G. sacculifer* shells, measured by Bé (*17*), from plankton tows and from surface sediment off Barbados, shows that, on average, the shells are 18 μm larger in the sediment than in the plankton tows. This difference exactly matches an observed increase of 9 μm in the thickness of the last chambers of the sediment specimens compared to those of the plankton specimens. We devised a simple model considering the pregametogenic adult shell as enclosing an immature shell (identical to those collected between 0 and 10 m) within larger final chambers (identical to those collected between 10 and 75 m). Assuming that the shell is constituted of spherical chambers, gametogenic calcification adds 22 percent to the pregametogenic shell and thus accounts for 18 percent of the weight of the postgametogenic shell, as found in the sediment.
22. Bé (*17*) compared the total weight of 50 pregametogenic specimens of *G. sacculifer* from a surface plankton tow collected off Barbados with the total weight of 50 postgametogenic specimens of *G. sacculifer* from the surface sediment from the same general location. The 50 specimens in each group were of the same size range and included the following shell lengths: 9 samples, 500 μm; 13 samples, 520 μm; 11 samples, 540 μm; 7 samples, 560 μm; 6 samples, 580 μm; and 4 samples, 600 μm. The total weights of the pregametogenic and postgametogenic groups were 1330 and 1700 μg, respectively. Thus, gametogenic calcification added 28 percent to the weight of the pregametogenic shells.
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Radar Detection of Cloud-Seeding Effects

Abstract. *The effects on precipitation of artificially seeding clouds with Dry Ice have been monitored from cloud to ground with a radar that has a wavelength of 8.6 millimeters.*

The evaluation of the effects of artificial ice nucleants on clouds and precipitation is a difficult task that requires, in general, careful physical measurements and statistical evaluations (*1*). We describe here a powerful physical technique that utilizes an 8.6-mm-wavelength radar (*2*) with color display for detecting the effects of artificial seeding from cloud to ground (*3*).

To test the utility of the radar, we carried out a series of cloud-seeding trials in December 1979 and February 1980. The radar was located at Grayland, Washington, on the Pacific coast. In each trial an aircraft was used to seed layers of supercooled cloud with Dry Ice along tracks oriented perpendicular to the wind direction and located at various distances upwind from the radar. The antenna of the radar was pointed vertically in order to detect the seeded tracks and unseeded portions of the cloud as they moved overhead. After seeding, the aircraft made a series of passes at different altitudes through the seeded and unseeded clouds in order to obtain detailed measurements of their microstructures (*4*).

During the course of the 2-month experiment, 108 tracks were seeded. We describe here the results obtained on 1 day.

On 20 February 1980 a broken, nonprecipitating, altocumulus cloud deck was situated over Grayland. Cloud tops

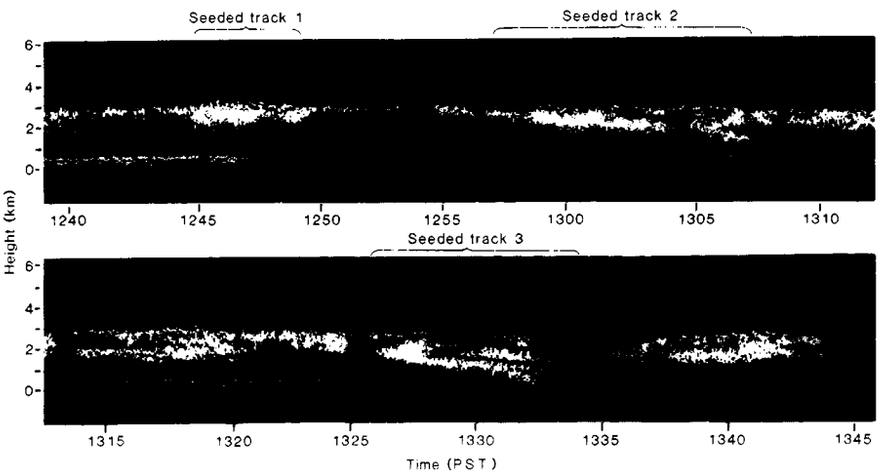

Fig. 1. Time-height display of radar echo pattern. The echo extending from ~ 0 to 0.5 km above ground is due to radar "ground clutter" and not to cloud. Clouds are located between ~ 5000 and 10,000 feet (~ 1.5 to 3 km). The gray areas represent the weakest radar echoes, black areas regions of intermediate radar echo strength, and white areas the strongest radar echoes. The third seeded track has precipitation falling from it that reached the ground at the radar site (see Fig. 3).

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