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# LETTER

# *In situ* $^{10}$Be-$^{26}$Al exposure ages at Meteor Crater, Arizona

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**Abstract**—A new method of dating the surface exposure of rocks from *in situ* production of $^{10}$Be and $^{26}$Al has been applied to determine the age of Meteor Crater, Arizona. A lower bound on the crater age of 49,200 ± 1,700 years has been obtained by this method.

FIFTEEN IMPACT CRATERS with associated meteorites are now known worldwide (SHOEMAKER, 1983); the ages of these craters provide a basis for estimating the flux of large meteoroids in the neighborhood of Earth. Ages of formation of many of these craters can be directly obtained from the well-studied method of measuring the terrestrial age of the meteorites (e.g., SUESS and WÄNKE, 1961). A new method of dating the surface exposure of crater materials using *in situ* produced cosmogenic nuclides can also be used to estimate ages of young impact craters and, potentially, can be applied even where meteorites have not been discovered. In order to explore the application of this latter technique and to determine the most geomorphically stable sites within and around the craters, we have undertaken a study of the exposure history of rocks at Meteor Crater, Arizona. A lower bound on the crater age of 49,200 ± 1,700 years has been obtained from *in situ* produced $^{10}$Be (half-life = $1.5 \times 10^6$ y) and $^{26}$Al ($7.05 \times 10^5$ y) concentrations in large ejecta blocks.

Meteor Crater (35°02′N, 111°01′W) is the largest known crater on the Earth with associated meteorites (Canyon Diablo) and is exceptionally well exposed and well explored (SHOEMAKER, 1963; SHOEMAKER and KIEFFER, 1974). The crater, which is 1.2 km in diameter and about 200 m deep, is late Pleistocene in age (SHOEMAKER, 1963). Since the crater was formed, a fallout deposit (mixed debris unit of SHOEMAKER, 1963) and upper units of the ballistic ejecta deposit (stratified debris units of SHOEMAKER, 1963) have been stripped from its rim. (See Fig. 1 for the distribution of the measured units.) The lower parts of the crater walls are mantled with Pleistocene talus and debris flow deposits, and the crater floor contains Pleistocene and Holocene alluvium and lake beds up to 30 m thick. From a variety of observations on erosion and deposition at the crater, including cavernous weathering of dolomite blocks on the rim, BLACKWELDER (1932) suggested that the crater was formed between the times of the Tahoe and Tioga glaciations of the Sierra Nevada. SHOEMAKER and KIEFFER (1974) found that the cratering event occurred during an interval in which the ground water table was much higher than at present and that two distinct Pleistocene units occur on the lower crater walls: (1) an old deposit of talus that rests on bedrock, allogenic breccia, or locally on the fallout deposit in the crater, and (2) a younger sequence of debris flow deposits that rests unconformably on the old talus and occupies deep gullies cut into the talus. They suggested that the old talus is mid-Wisconsin in age and that the debris flow deposits are late Wisconsin.

A variety of chronometric techniques has been applied that help delimit the age of Meteor Crater and its history of erosion and deposition. In principle, a date for the impact event can be obtained from the terrestrial age of the Canyon Diablo meteorite determined from concentrations of cosmogenic radionuclides. No reliable $^{14}$C (half-life = 5730 y) data are available because of the meteorite's long terrestrial age and the low production rate of $^{14}$C in large iron meteorites. Kaye (1963) measured $^{59}$Ni (half-life = $7.6 \times 10^4$ y) in 4 pieces of Canyon Diablo and obtained a terrestrial age of $<4 \times 10^4$ y, based on the highest value of 0.76 ± 0.06 dpm/g Ni. This low limit on the age was obtained because the production rate of $^{59}$Ni was underestimated. Although the shielding depth of Kaye's samples is not known, a more meaningful upper limit of the terrestrial age is $<2.5 \times 10^5$ y, based on a recent theoretical calculation of $^{59}$Ni production rates by SPERGEL et al. (1986).

Attempts to date the crater directly have yielded a more definite age. A radiocarbon age of 24,000 ± 2,000 y was obtained by IVES et al. (1964) from gastropod shells from a sample of lake beds in the crater floor. This sample was collected from the dump on the collar of a shaft that penetrates the lake beds on the floor of the crater; consequently, its stratigraphic position in the 30 m thick lake bed sequence is unknown. As it was collected from the surface of the dump, probably from material removed at a late stage of excavation, the sample probably came from considerable depth in the lake beds. To the extent that the carbon analyses from the gastropod shells were from the atmospheric $CO_2$ dissolved in the lake waters, the $^{14}$C age represents a lower limit on the age of the crater. SUTTON (1985) obtained an age of 49,000

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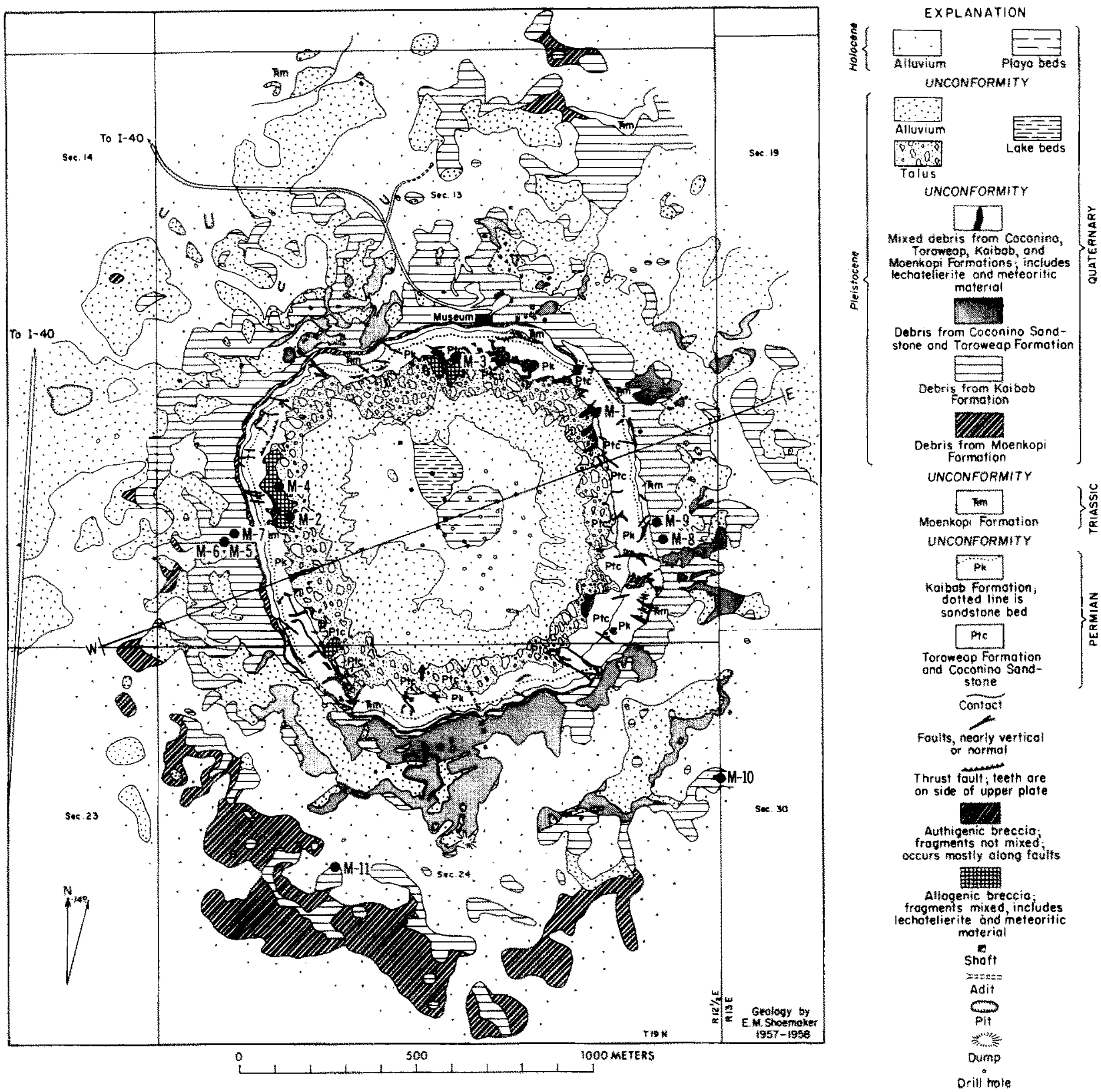


FIG. 1. Geological map of Meteor Crater, Arizona (SHOEMAKER, 1963), showing locations of samples M-1 through M-11.

± 3000 y for the impact event from a detailed investigation of the thermoluminescence of quartz in shocked rocks from breccia in the floor of the crater and shocked fragments derived from the fallout layer. In an accompanying paper PHILLIPS et al. (1991) report a mean cosmogenic $^{36}$Cl (half-life = $3.01 \times 10^5$ y) exposure age of 49,200 ± 700 y for several samples collected from the summits of large blocks of dolomite on the rim of the crater.

The use of *in situ* produced $^{10}$Be-$^{26}$Al to determine exposure ages and erosion rates of terrestrial rocks and sediments has been developed and applied to geomorphological studies for several years (LAL and ARNOLD, 1985; NISHIIZUMI et al., 1986, 1989, 1991; LAL, 1991). The production rates of these two nuclides and the production ratio have been obtained from studies of very late Pleistocene glacially polished surfaces in the Sierra Nevada (NISHIIZUMI et al., 1989). The production rates of $^{10}$Be and $^{26}$Al have been independently verified by studies of Antarctic rocks (NISHIIZUMI et al., 1991).

In the present study, we collected four rock samples from the crater walls and six rock samples from the summits of large blocks on the crater rim in a layer of ejecta derived from the Kaibab Formation of Permian age. Sandy dolomite of the Kaibab is by far the most resistant to weathering of all the rock types at Meteor Crater. Hence, the dolomite, which forms ledges on the crater walls and prominent knobs in the ejecta, was expected to have the longest continuous exposure at the surface. Our objective was to determine the ages of the oldest surfaces that we could find, both within the crater and

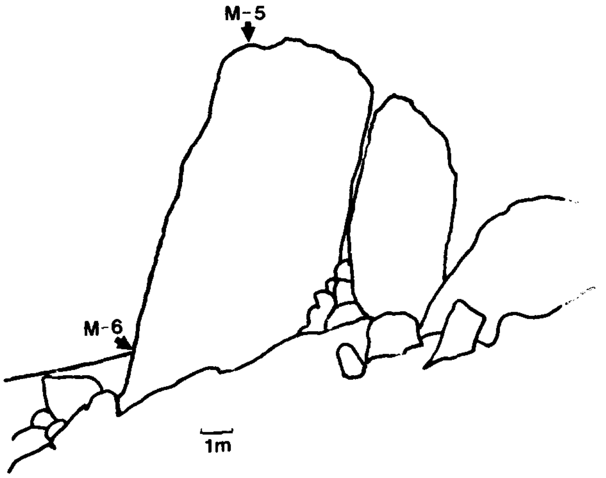


FIG. 2. Line drawing of Whale Rock based on a photograph. The view here is from the southeast; the "whale's" head and mouth are clearly visible. Foreground rocks and alluvium are included in the drawing. The site of M-5 can be seen near the summit of the rock; the exact site of M-6 cannot be seen in this drawing. It is around the back of the rock at the height of the arrow.

on the crater rim. The sample locations are shown in Fig. 1. All samples, except one (M-6), were collected from relatively flat surfaces on bedrock or on blocks derived from the Kaibab Formation; they consist chiefly of dolomite with various amounts of detrital and secondary quartz. Maximum sample depth from the exposed surfaces is 1.5 cm. M-1 was collected near the base of the Beta Member of the Kaibab on a pinnacle rising above the oldest Pleistocene talus on the northeast crater wall; M-2 and M-4 are from large blocks in the old talus unit on the west wall; M-3 is from a block at the head of the old talus on the north wall; M-5 is from the summit of Whale Rock (BARRINGER, 1910; about 9 m height and one of the largest exposed ejecta blocks) on the west rim of the crater; M-6 was collected from the side of Whale Rock (74° slope and about 1.2 m above the present level of alluvium at the base of the rock, as shown in Fig. 2); M-7 is from the summit of another block, 20 m east of Whale Rock; M-8 and M-9 (50 m north of M-8) are from large blocks on the east rim of the crater (M-9 is from the block known as Monument Rock; BARRINGER, 1910); M-10 and M-11 are from large blocks in the ejecta blanket south of the crater. The blocks sampled in the talus are from the lower part of the Alpha Member of the Kaibab, as are the large blocks that we sampled in the ejecta. All rocks sampled lay at depths greater than 10 m just prior to the cratering event and were therefore shielded from cosmic-ray production of $^{10}$Be and $^{26}$Al. Samples M-7 to M-10 were taken at the same sites that were sampled for $^{36}$Cl measurements; they correspond, respectively, to $^{36}$Cl samples MC-5, MC-1, MC-2, and MC-4 (PHILLIPS et al., 1991). The Coconino sandstone, which was also deeply buried prior to impact, has been stripped from most of the crater rim. Remaining blocks of Coconino ejecta, as well as bedrock Coconino outcrops in the crater walls, probably have been deeply denuded since the crater wall was formed and were not sampled. Detailed sample descriptions are given elsewhere (SHOEMAKER et al., unpubl. data).

The samples were processed to obtain a pure quartz phase, and Be and Al were then chemically separated and purified (KOHL and NISHIIZUMI, unpubl. data). The Al concentrations in quartz separated from these dolomite samples were higher than in most of the quartz separates we have studied. $^{10}$Be and $^{26}$Al measurements were performed at the University of Pennsylvania tandem accelerator using AMS (accelerator mass spectrometry; KLEIN et al., 1982; MIDDLETON et al., 1983). The results are shown in Table 1, which also lists the altitudes of the samples and the Al concentrations in the purified quartz. The $^{10}$Be and $^{26}$Al concentrations were calculated from the amount of Be carrier added ($\sim$1.5 mg), the Al content in the quartz, and the $^{10}$Be/Be and $^{26}$Al/Al ratios measured by AMS. The average measured $^{26}$Al/$^{10}$Be ratio, 6.33 ± 0.19, is in excellent agreement with the previously measured production ratio of 6.1 ± 0.4 (NISHIIZUMI et al., 1989), as expected for an exposure time that is short compared to the half-lives of $^{10}$Be and $^{26}$Al. Exposure ages (Table 2) were calculated for these samples using as the production rates of $^{10}$Be and $^{26}$Al at sea level (>50° latitude), 6.0 and 36.8 atom/year g $SiO_2$, respectively (NISHIIZUMI et al., 1989), and correcting for sample altitude, latitude, and exposure geometry. These correction factors are based on cosmic-ray measurements, and their associated errors are expected to be less than 10% (LAL and PETERS, 1967); they are discussed in detail by LAL (1991) and NISHIIZUMI et al. (1989). In the present work, we use geographic latitude to calculate exposure ages since the present geomagnetic latitude is not the same as that during the last 50,000 y. We estimate an overall uncertainty of about ±10% in the production rates of these nuclides primarily because of uncertainties in the age of the glacial polishing of the rock used to determine them (NISHIIZUMI et al., 1989). This 10% error has not been included in the errors quoted in this paper for the exposure ages.

The $^{10}$Be and $^{26}$Al results can be used to derive two limiting

Table 1. $^{10}$Be AND $^{26}$Al IN METEOR CRATER SAMPLES

| ID | Altitude (m) | $SiO_2$ Wt(g) | Al (ppm) | $^{10}$Be ($10^6$ atom/g $SiO_2$) | $^{26}$Al ($10^6$ atom/g $SiO_2$) | $^{26}$Al/$^{10}$Be |
|---|---|---|---|---|---|---|
| M-1 | 1680 | 30.26 | 250 | 0.448 ± 0.031 | 2.74 ± 0.22 | 6.12 ± 0.65 |
| M-2 | 1630 | 21.17 | 320 | 0.436 ± 0.037 | 2.47 ± 0.20 | 5.67 ± 0.66 |
| M-3 | 1630 | 21.34 | 300 | 0.253 ± 0.031 | 1.80 ± 0.15 | 7.09 ± 1.05 |
| M-4 | 1630 | 26.01 | 260 | 0.255 ± 0.032 | 1.60 ± 0.14 | 6.27 ± 0.97 |
| M-5 | 1730 | 21.48 | 310 | 0.921 ± 0.051 | 5.71 ± 0.52 | 6.20 ± 0.66 |
| M-6 | 1725 | 50.20 | 90 | 0.306 ± 0.013 | 1.68 ± 0.10 | 5.49 ± 0.40 |
| M-7 | 1730 | 35.38 | 230 | 0.833 ± 0.053 | 5.60 ± 0.38 | 6.72 ± 0.63 |
| M-8 | 1730 | 20.80 | 340 | 0.687 ± 0.074 | 4.97 ± 0.20 | 7.24 ± 0.84 |
| M-9 | 1730 | 35.76 | 350 | 0.543 ± 0.020 | 3.63 ± 0.19 | 6.69 ± 0.42 |
| M-10 | 1700 | 35.94 | 390 | 0.842 ± 0.028 | 5.73 ± 0.35 | 6.80 ± 0.48 |
| M-11 | 1700 | 27.93 | 640 | 0.940 ± 0.031 | 5.00 ± 0.29 | 5.32 ± 0.35 |
| | | | | | Average | 6.33 ± 0.19 |

Table 2. CALCULATED EXPOSURE AGES

| ID | $^{10}$Be Age (years) | $^{26}$Al Age (years) | Average (years) |
|---|---|---|---|
| M-1 | 24,800 ± 1,700 | 24,700 ± 2,000 | 24,700 ± 1,300 |
| M-2 | 24,900 ± 2,100 | 22,700 ± 1,900 | 23,800 ± 1,600 |
| M-3 | 15,000 ± 1,800 | 16,900 ± 1,500 | 16,000 ± 1,400 |
| M-4 | 14,600 ± 1,800 | 14,500 ± 1,400 | 14,500 ± 1,100 |
| M-5 | 49,500 ± 2,800 | 50,700 ± 4,800 | 49,800 ± 2,400 |
| M-6 | 24,100 ± 1,000 | 22,000 ± 1,300 | 23,000 ± 1,500 |
| M-7 | 44,700 ± 2,900 | 50,400 ± 3,500 | 47,600 ± 4,000 |
| M-8 | 36,800 ± 4,000 | 44,600 ± 1,800 | 40,700 ± 5,500 |
| M-9 | 29,100 ± 1,100 | 32,400 ± 1,700 | 30,700 ± 2,300 |
| M-10 | 46,200 ± 1,600 | 52,700 ± 3,300 | 49,400 ± 4,600 |
| M-11 | 51,600 ± 1,700 | 45,800 ± 2,700 | 48,700 ± 4,100 |

case histories for surface exposure as described by LAL (1991) and NISHIIZUMI et al. (1991). These involve either steady-state erosion with continuous exposure or a sudden exposure assuming no subsequent erosion. The actual history of any rock surface is probably a mixture of these two cases. Exposure ages, as they are discussed in this paper, have been calculated assuming no erosion and therefore are defined to be minimum exposure ages. These ages also assume no snow cover on the rocks sampled. Correction for snow cover is negligible at present, but in times of more extensive glaciation this could possibly be important. Such a correction would be in the direction to increase the actual exposure times that our samples must have experienced in order to have accumulated their present levels of radioactivity, further defining the ages given here as minimum ages. The error of the ages due to sample thickness is less than 2% in this study.

Four sets of exposure ages were found, ~49,000 y (M-5, 7, 10, and 11), ~31,000–41,000 y (M-8 and 9), ~24,000 y (M-1 and 2), and ~15,000 y (M-3 and 4). The age of ~49,800 y, found for the summit of Whale Rock (M-5) and the similar ages found for sites M-7, 10, and 11, record the approximate times that the summits of these large ejecta blocks were exposed after erosion had removed the fallout layer, the Coconino debris layer, and uppermost part of the Kaibab debris layer from the crater rim.

A young exposure age of 23,000 y was found 1.2 m above the base of Whale Rock (M-6), and provides a basis for calculating a denudation rate near Whale Rock. About 8 m of the matrix of the Kaibab ejecta was stripped from the flanks of Whale Rock in 27,000 y (average erosion rate of 30 cm/1,000 y). The mean rate of denudation in the last 23,000 y, on the other hand, has been only 5 cm/1,000 y. In fact, a late Pleistocene soil is formed in the alluvium at the base of Whale Rock, and the surface probably has been nearly stable for the last 10,000 y. Evidently, denudation rates were highest shortly after the crater formed. We think it is significant that the exposure ages of block summits high on the crater rim (sites M-5 and M-7) are approximately concordant with the exposure ages of block summits much farther from the rim (sites M-10 and M-11), where the Kaibab blocks probably were buried by no more than a few meters of crushed Coconino ejecta and unconsolidated fallout material. The Coconino debris layer and the fallout layer were much more easily eroded than the Kaibab layer and have been entirely removed from most of the crater rim. On the basis of the early erosion rate established at Whale Rock, we estimate that the summit of Whale Rock and of the large blocks at sites M-7, 10, and 11 may have been exposed within a few thousand years after the crater formed. Exposure of the summits of large blocks at sites M-8 and 9 appears to have been delayed by an additional 10,000 to 15,000 y. The approximately concordant exposure ages for M-5, 7, 10, and 11, which are in good agreement with the age of the shock event determined by the thermoluminescence method (SUTTON, 1985) and with several exposure ages obtained by the $^{36}$Cl method (PHILLIPS et al., 1991), probably are close to the age of the crater.

The relationship between our $^{10}$Be-$^{26}$Al ages and the $^{36}$Cl ages in the companion paper by PHILLIPS et al. (1991) for M-7 to M-10 is shown in Fig. 3. As described above, samples M-7 to M-10 correspond, respectively, to $^{36}$Cl samples MC-5, 1, 2, and 4 (PHILLIPS et al., 1991). Figure 3 shows a strong correlation between ages determined with the two methods.

The surface of a block in the old talus unit on the west crater wall (M-2) and the most stable bedrock surface that we could identify in the crater, which is above the old talus unit on the northeast wall (M-1), have exposure ages of about 24,000 y, close to the peak of the late Wisconsin pluvial maximum. These ages, at first, seem to support Blackwelder's interpretation that the old talus unit is of late Wisconsin age (Tioga, as used by BLACKWELDER, 1932). However, as the old talus unit rests directly on the easily eroded fallout deposit in many places, the base of this unit locally must be the earliest talus formed, and it must have been deposited shortly after the crater's formation. The 24,000 y exposure ages probably reflect the approximate time at which the sampled surfaces were exhumed or freshly eroded during the late Wisconsin pluvial maximum, which can now be identified as the interval during which the old talus was dissected and younger debris flows were deposited. Indeed, ~15,000 y exposure ages also were obtained from two blocks in the old talus unit, which suggests that the episode of dissection continued, perhaps intermittently, nearly to the end of the Pleistocene. It is interesting that these exposure ages of the crater walls are very similar to those obtained for the walls of Wolfe Creek Crater,

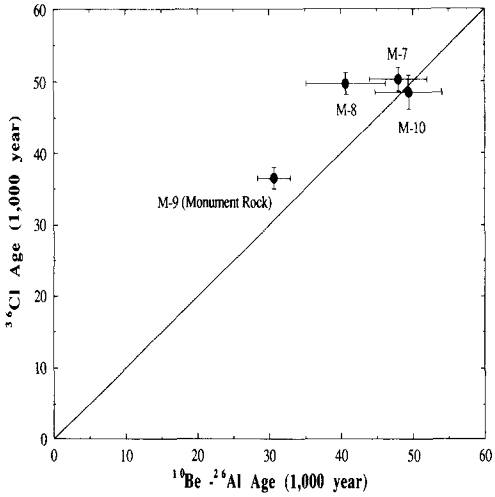


FIG. 3. The $^{36}$Cl ages of Phillips et al. (1991) are plotted vs. the $^{10}$Be-$^{26}$Al ages for M-7 to M-10 discussed in this paper. Samples M-7 to M-10 correspond, respectively, to $^{36}$Cl samples MC-5, 1, 2, and 4.

Australia, even though the age of Wolfe Creek Crater is ~300,000 y (SHOEMAKER et al., 1990). The knowledge of sampling strategies obtained from this work is being applied to meteorite craters in Australia (SHOEMAKER et al., 1990).

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