Reports

A 250,000-Year Climatic Record from Great Basin Vein Calcite: Implications for Milankovitch Theory

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A continuous record of oxygen-18 ($\delta^{18}O$) variations in the continental hydrosphere during the middle-to-late Pleistocene has been obtained from a uranium-series dated calcitic vein in the southern Great Basin. The vein was deposited from ground water that moved through Devils Hole—an open fault zone at Ash Meadows, Nevada—between 50 and 310 ka (thousand years ago). The configuration of the $\delta^{18}O$ versus time curve closely resembles the marine and Antarctic ice core (Vostok) $\delta^{18}O$ curves; however, the U-Th dates indicate that the last interglacial stage (marine oxygen isotope stage 5) began before 147 ± 3 ka, at least 17,000 years earlier than indicated by the marine $\delta^{18}O$ record and 7,000 years earlier than indicated by the less well dated Antarctic $\delta^{18}O$ record. This discrepancy and other differences in the timing of key climatic events suggest that the indirectly dated marine $\delta^{18}O$ chronology may need revision and that orbital forcing may not be the principal cause of the Pleistocene ice ages.

WE HAVE OBTAINED A DETAILED and well-dated record of $\delta^{18}O$ fluctuations in Great Basin ground waters for the period 50 to 310 ka (thousand years ago) from a calcite vein from Ash Meadows, Nevada. In this report, we describe this record, compare it with marine and Antarctic ice core $\delta^{18}O$ time series and with several continental proxy records, and discuss the implications of the data for the Milankovitch hypothesis of Pleistocene climatic change.

The Ash Meadows oasis is the principal discharge area of the Ash Meadows ground-water basin (Fig. 1). This ~12,000 km^2 basin is underlain by a thick section (~100 to >1000 m) of fractured Paleozoic carbonate rocks (*1–5*). Ground water moves through these rocks along calcite-coated fractures that are open to depths of at least 450 m below the top of this aquifer and at least 1300 m below land surface (*1*). The Spring Mountains and Sheep Range (Fig. 1) are the principal modern recharge areas. Water temperature, water chemistry, and spring discharge at Ash Meadows have not varied significantly during the last 50 years (*4*), although water discharge was clearly greater during the late Pliocene to middle Pleistocene than it is today (*6*).

The calcite vein sample that we analyzed (DH 2) was collected from ~21 m below water table in Devils Hole (Fig. 1), a 1- to 5-m-wide open fault zone near the center of the modern spring lineament at Ash Meadows. Devils Hole extends to more than 100 m below water table (*7*), which is about 15 m below land surface.

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We dated growth layers in DH 2 with both the ^{230}Th-^{234}U and ^{234}U-^{238}U methods (Table 1 and Fig. 2). We were able to use the ^{230}Th-^{234}U method to date the vein calcite because dissolved Th was essentially absent (<0.01 ppb) in Devils Hole ground water. To obtain ^{234}U-^{238}U dates, we back calculated the initial $^{234}U/^{238}U$ ratio in the ground waters from which the calcite was precipitated using the ^{230}Th-^{234}U dates (Table 1) and the $^{234}U/^{238}U$ ratios measured in the vein calcite. The back-calculated initial ratios (Table 1, column 8) for individual laminae of DH 2 range between 2.53 and 2.85 (Table 1); the average value [2.70 ± 0.07 (1σ)] agrees closely with $^{234}U/^{238}U$ ratios measured (*8*) in waters from the seven largest artesian springs at Ash Meadows [2.76 ± 0.09 (1σ)]. That this ratio has remained almost constant for nearly 300,000 years is consistent with the large size of the drainage basin and the known homogenization of local differences in modern water chemistry during flow of ground water to the Ash Meadows oasis (*1, 4*). Because of the small variation in the calculated initial ratios during the past 300,000 years and the similarity of these ratios with those in modern spring waters, we believe that the calculated ^{234}U-^{238}U ages are valid. That DH 2 has always behaved as a geochemically closed system is indicated by (i)

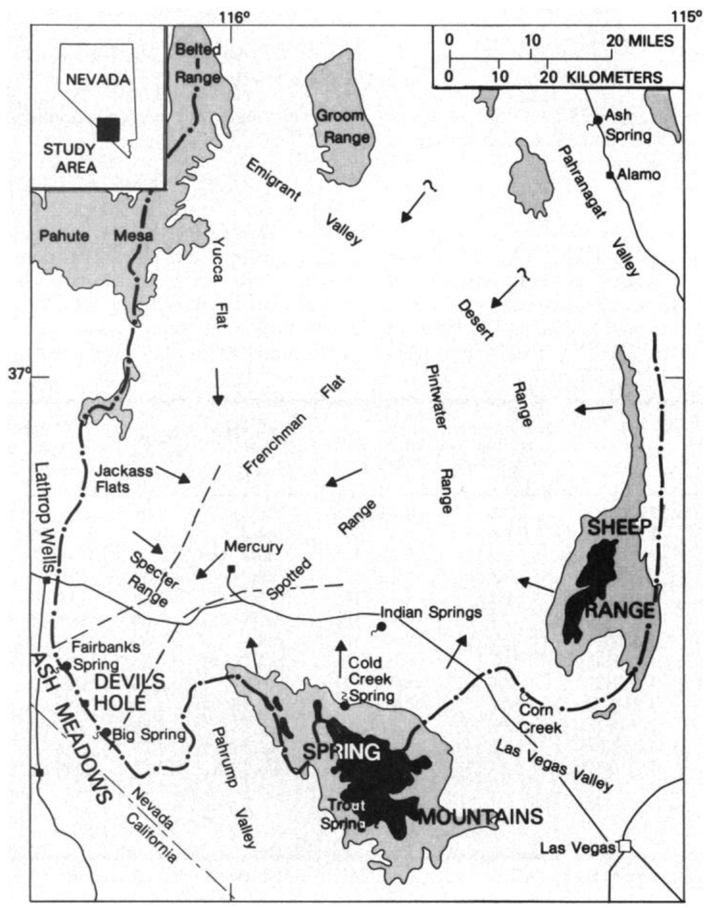

Fig. 1. Index map of southcentral Great Basin. Major mountains are shaded: heavy shading denotes altitudes of 2400 to 3600 m; light shading denotes altitudes of 1800 to 2400 m; ridges <1800 m are designated by name only. Dashed and dotted line, boundary of Ash Meadows ground-water basin (*1*). Arrows denote general direction of ground-water flow in aquifer as inferred from potentiometric maps (*1*). Dashed line is a major trough in potentiometric surface of carbonate aquifer.

the absence of recrystallization (*8*) or depositional hiatuses in the pure calcite (less than 0.5% insoluble residues); (ii) the consistency of the 13 paired ^{230}Th and ^{234}U dates, which, with only two exceptions, are in agreement with the stratigraphic order determined from field relations; and (iii) concordance of the ^{230}Th and ^{234}U ages, which are within one standard deviation of measurement error for 11 of the 13 laminae.

We used a weighted least-squares regression to derive a linear age versus distance relation for each clock (Fig. 2), and in each case better than 97% of the variance was explained by the relation. From the slope of the regression relation the average growth rates were 0.54×10^{-3} mm/year for the ^{230}Th clock and 0.56×10^{-3} mm/year for the ^{234}U clock. The extrapolated ^{230}Th age at the exposed (outer) face of DH 2 is 51 ± 2.8 ka, and the age at 140 mm, the full thickness of the sample, is 308 ± 10.2 ka; thus DH 2 provides a record spanning about 257,000 years. The ^{234}U age regression (Fig. 2) yields an age of 58 ± 6.0 ka at the free face and 310 ± 8.6 ka at full thickness, a ~252,000-year growth period. Because we compare our record with ^{230}Th-^{234}U dated marine and continental records, we principally discuss the ^{230}Th ages hereafter.

The curtailment of the DH 2 record at about 51 ka is puzzling. The water in Devils Hole is slightly supersaturated with respect to calcite (*9*), and calcite precipitated on most calcite crystals that we placed in the hole (*9*). For a growth rate of 0.55×10^{-3} mm/year, about 30 mm of sample is missing. There is no evidence that the sample had a complex growth history caused by episodes of breaking and reprecipitation as we have observed in other samples from Devils Hole. Sparse incipient borings occur on the outer free face of DH 2; a coating of organic matter may have inhibited calcite precipitation in parts of Devils Hole during the past 50,000 years.

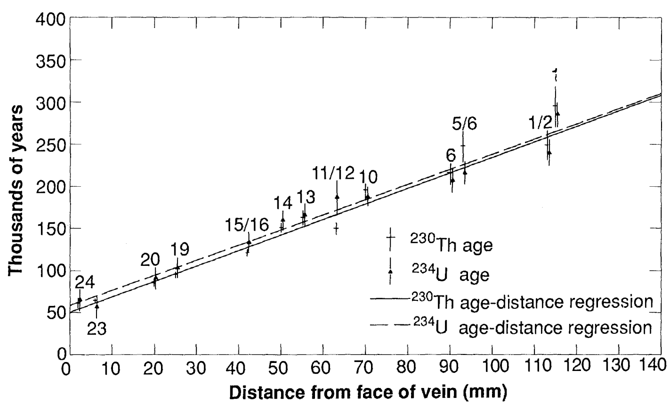

Fig. 2. Uranium-series ages of DH 2, Devils Hole, Ash Meadows, Nevada [numbers above data symbols refer to the laminae groups (Table 1)]. For clarity, the ^{234}U-^{238}U ages are plotted slightly to the right of the ^{230}Th-^{234}U ages. Weighted least-squares regression was used to derive the following age-distance (x, in millimeters) relations: ^{230}Th-^{234}U age (ka) = 1.839 (±0.086) x + 51.313 (±2.771) and ^{234}U-^{238}U age (ka) = 1.802 (±0.092) x + 58.398 (±5.951), where ka = 10^3 years ago. The percent variation explained by the regression is 97.6 and 97.2, respectively.

Table 1. Analytical data and calculated U-series ages of laminae from Devils Hole 2 (DH 2), Ash Meadows, Nevada. We subdivided DH 2 into 24 sections (laminae groups) cut parallel to faint laminations near the top and base of the vein. Activity ratios reported with one standard deviation propagated error in parentheses for last significant digits. Ages were calculated with half-lives for ^{230}Th of 75,200 and ^{234}U of 244,000 years. The initial ^{234}U/^{238}U activity ratio (column 8) was calculated with the ^{230}Th age and the measured ^{234}U/^{238}U activity; the average of the initial ratios (2.70) was used to calculate the ^{234}U-^{238}U age.

Laminae group	Distance (mm)*	Uranium (ppm)	Activity ratios			^{230}Th age (10^3 years)	Initial ^{234}U/^{238}U	^{234}U age (10^3 years)
			^{234}U/^{238}U	^{230}Th/^{232}Th	^{230}Th/^{234}U			
24	2	0.416(10)	2.415(35)	25(4)	0.450(27)	61(6)	2.68	65(13)
23	6	0.498(10)	2.445(30)	128(18)	0.470(9)	64(2)	2.73	57(12)
20	20	0.466(10)	2.312(29)	99(10)	0.577(11)	85(3)	2.67	91(12)
19	25	0.432(8)	2.264(26)	213(36)	0.622(11)	95(3)	2.66	104(12)
15/16†	42	0.490(10)	2.163(27)	96(8)	0.730(14)	121(4)	2.64	134(12)
14	50	0.497(10)	2.079(24)	176(22)	0.825(15)	151(5)	2.66	160(12)
13	55	0.514(10)	2.060(24)	262(72)	0.859(16)	163(7)	2.68	166(12)
11/12†	63	0.544(12)	1.996(35)	>100	0.821(18)	150(7)	2.53	188(17)
10	70	0.541(10)	1.997(22)	344(90)	0.933(17)	196(9)	2.74	188(12)
6	90	0.423(8)	1.944(22)	309(85)	0.968(18)	216(12)	2.74	207(13)
5/6†	93	0.408(8)	1.917(23)	239(30)	1.018(19)	248(16)	2.85	217(13)
1/2†	113	0.439(9)	1.859(25)	143(23)	1.015(20)	249(17)	2.74	240(15)
1	115	0.428(8)	1.754(21)	120(15)	1.060(20)	296(25)	2.75	286(14)
						Average:	2.70 ±0.07 (1σ)	

*Distance from outer (free) face of vein to midpoint of laminae. †Indicates that two sections were combined for dating.

We obtained the δ^{18}O record (Fig. 3) from 110 samples across the 140-mm thick vein. The average sampling interval was 1.27 mm; this distance represents ~2,300 ± 100 years. Precision of the δ^{18}O data at 1σ is 0.07 per mil, reported relative to VSMOW (*10*) on a scale normalized to $\delta^{18}O_{SLAP} = -55.5$ per mil (*10*).

Several factors may have influenced the measured δ^{18}O fluctuations including (i) exchange of oxygen isotopes between ground water and the rocks composing the aquifer; (ii) variations of the ground-water temperature at which calcite precipitated; (iii) the relative importance of winter and summer recharge; (iv) variations in local surface temperature during (atmospheric) precipitation; (v) varying sources for the moisture entering the southern Great Basin; (vi) ocean water–related factors; and (vii) variations in ground-water travel time from recharge to discharge areas. The first three factors do not appear to be important in the Ash Meadows basin (*11–15*).

That δ^{18}O variations in precipitation are strongly correlated with local surface temperature is well established globally from more than two decades of observation (*14*); δ^{18}O in precipitation increases linearly with surface temperature from −60° to +15°C at middle- to high-latitude stations. Thus, increases (or decreases) in δ^{18}O in DH 2 (Fig. 3) should reflect increases (or decreases) in mean surface temperature in the recharge areas of Ash Meadows basin. Because only winter-spring precipitation appears to have contributed to recharge (*11*), we believe that the δ^{18}O variations in DH 2 reflect changes in mean winter and spring surface air temperature.

The δ^{18}O variations probably also reflect changes in winter moisture sources. Major modern storms entering the Great Basin derive their moisture from polar, maritime polar, or tropical Pacific air masses (*16*), each of which is likely to have a different δ^{18}O. The DH 2 δ^{18}O variations may also reflect the 1.3 to 1.8 per mil change in δ^{18}O of the oceans between full glacial and peak interglacial times (*17*), as well as accompanying changes in ocean surface water temperatures and humidity at the surface. We are unable to evaluate the relative effects of moisture source and ocean-water-air–related factors to the δ^{18}O variations in DH 2. In summary, as a first approximation, we accept the strong global relation between δ^{18}O

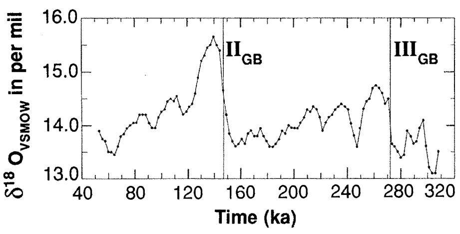

Fig. 3. Variations in $\delta^{18}O$ in DH 2 during the middle and late Pleistocene. Each dot represents an analyzed calcite sample; ages were assigned by the use of the ^{230}Th-^{234}U age-distance regression equation (Fig. 2). Lines labeled II_{GB} and III_{GB} are suggested Great Basin equivalents of Terminations II and III of the marine $\delta^{18}O$ chronology.

and mean annual surface air temperature (*14*) to indicate that the heaviest isotopic values displayed by the DH 2 $\delta^{18}O$ time series reflect relatively warm weather and the lightest values relatively cold weather.

The uranium-series clocks were activated upon the co-precipitation of uranium with calcite in Devils Hole. But the time when the climate-induced changes in $\delta^{18}O$ occurred in precipitation falling on the Spring Mountains and Sheep Range (Fig. 1), 70 to 100 km distant from Devils Hole, will be older than the ^{230}Th ages by the mean residence time of water in the aquifer. Unfortunately, the residence time is poorly known (*4*). Most likely, it has varied from perhaps just a few thousand years during a full glacial climate to 16,000 years or more during and after transition to interglacial times (*18*). Inexact knowledge of residence times clearly is a deficiency of the DH 2 record. However, this limitation has the effect of making our interpretations (discussed below) conservative.

The $\delta^{18}O$ data indicate that two episodes of rapid warming began in the southern Great Basin about 272 ± 8.5 and 147 ± 3.1 ka (Fig. 3, III_{GB} and II_{GB}). We suggest that these transitions represent Great Basin (GB) equivalents of terminations III and II of the marine $\delta^{18}O$ record (Fig. 4). Because of ground-water residence times, these terminations may have occurred perhaps 16,000 years earlier than shown by the DH 2 record. The major warm intervals (interglacials or interpluvials) that followed the terminations are, in turn, succeeded by a progressive cooling trend on which are superimposed subsidiary colder (stadial) and warmer (interstadial) periods (Fig. 3). The sawtooth pattern of isotopic changes (*19*) in the marine record is clearly evident in the GB record.

The time intervals between adjacent peaks and between adjacent troughs for the relatively wet periods when ground-water residence times were likely to have been relatively short (60 to 120, 160 to 250, and 280 to 310 ka) suggest that the obliquity solar insolation cycle (41,000 years) (*20*) and perhaps the precession cycles (19,000 and 23,000 years) occur in the data, although other factors are evidently involved. Approximately 125,000 years separates terminations II_{GB} and III_{GB} (Figs. 3 and 4); and if termination I occurred about 12 ka (Fig. 4), then 135,000 years separate terminations I and II_{GB}. These intervals nearly match the third and fifth terms of the eccentricity cycle, 123 and 131 ka (*20*). A spectral analysis of our entire data set suggests that cycles with periods (in order of decreasing power) of 126.5, 62.1, 32.2, 42.5, 25.3, 27.6, 34.5, 50.6, and 23 thousand years are present (*21*).

The close similarity in the shapes of the marine, Antarctic ice core, and DH 2 $\delta^{18}O$ curves (Fig. 4) suggests that DH 2 recorded global paleoclimatic conditions. Although the overall patterns are clearly similar, the major peaks and troughs that appear to be related in the different records (Fig. 4) do not occur at the same time. Four major features are shown by the marine and DH 2 curves, two of which are also displayed by the Antarctic record: (i) the major trough in isotope stage 4 of the marine record at about 65 ka, (ii) termination II (and II_{GB}), (iii) the major trough at about 225 ka on the marine curve, and (iv) termination III (and III_{GB}).

All three $\delta^{18}O$ curves show a major trough, presumably a major stadial, at about 65 ka (Fig. 4). We believe that ground-water residence times were perhaps only a few thousand years during major stadial and full glacial climates (*18*). Thus, the actual occurrence of this event in the Ash Meadows basin may be only slightly older than indicated by DH 2.

A cornerstone for the acceptance of the Milankovitch theory of Pleistocene climate change has been the coincidence of the timing of termination II (about 130 ka), as indicated by the marine $\delta^{18}O$ curves (Fig. 4), with projections of solar insolation (Fig. 5) (*22, 23*), which indicate that the period 135 to 125 ka was marked by unusually high summer insolation and unusually low winter insolation at latitudes above 40°N. Such conditions have been considered to be particularly favorable for global deglaciation. The DH 2 $\delta^{18}O$ data, however, indicate that termination II (II_{GB} on Figs. 3 and 4) occurred no later than 147 ka and, because of residence times, possibly 16,000 years earlier. earlier.

Conversely, periods of unusually low summer insolation, accompanied by high winter insolation at latitudes north of 40°—for example, at 115 and 230 ka (Fig. 5)—should favor rapid glaciation (*22*). Ruddiman and McIntyre (*24*) have suggested that the major stadial at about 225 ka on the marine $\delta^{18}O$ curve (Fig. 4) reflects a short-lived glaciation during one 23,000-year precession cycle. This time (231 ka) represented one of the strongest summer insolation minima of the past 1 million years (Fig. 5). Ruddiman and McIntyre dated this event on the basis of an ash layer that occurs in Pacific deep sea cores V19-28 and V19-29 just below the stadial maximum. The age of this ash was thought to be approximately 230 ka, but the uncertainty in this age is large (*25*). In contrast, we believe that this stadial

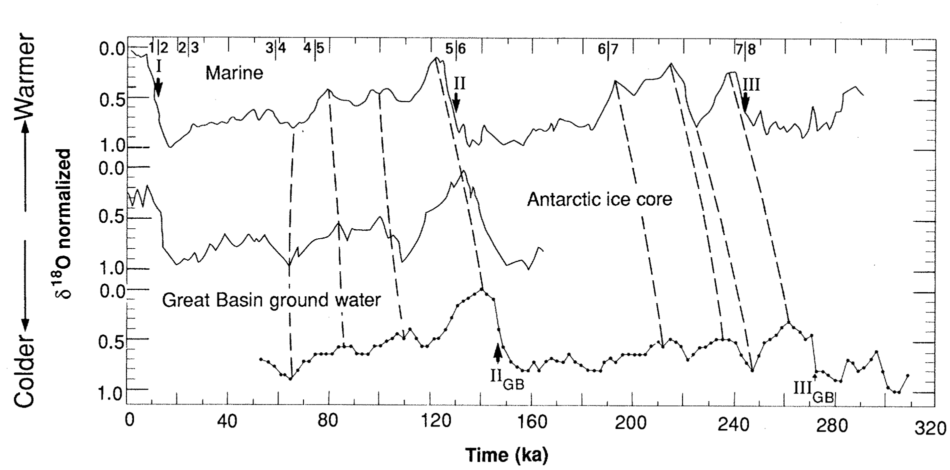

Fig. 4. Comparison of marine (*30*), Antarctic ice cap (*13*), and DH 2 $\delta^{18}O$ records for the middle-to-late Pleistocene; records are normalized to permit easy comparison. Numbers below upper margin are isotope stage boundaries of marine chronology (*30*). The marine $\delta^{18}O$ record is the "final" orbitally tuned record of Martinson *et al.* (*30*); minor fluctuations in this record between 18 and 22 ka are not shown at this scale. Absolute variation of $\delta^{18}O$ is <1.8 per mil in the marine record (*30*), 8.2 per mil in the Antarctic record (*13*), and 2.5 per mil in DH 2. Points labeled II and III denote terminations (major deglaciations) in the marine $\delta^{18}O$ record; II_{GB} and III_{GB} denote terminations suggested by the DH 2 record for the Great Basin. Ages of features shown by Great Basin curve are minimum values because of ground-water residence times. Vertical and sloping dashed lines connect major peaks and troughs believed to be correlative.

Table 2. Comparison of ages in thousands of years ago of key climatic events belived to be correlative in the DH 2, marine, and Antarctic (Vostok) $\delta^{18}O$ time series. Errors for DH 2 (in parentheses) are the one standard deviation of prediction derived for the regression relation (Fig. 2); errors are ±5,000 years for the marine core (*30*), and ±10,000 to 15,000 years for the base of the Antarctic record (*13*); IS, isotope stage.

Event	DH 2		Antarctic	Marine
	^{230}Th	^{234}U		
Major stadial in IS 4	65(2.3)	72(5.4)	65	66
Termination II	147(3.1)	152(3.3)	140	130
Major stadial in IS 7	248(7.4)	251(5.9)		225
Termination III	272(8.5)	275(7.0)		244

is correlative with the major trough on the DH 2 curve at about 248 ka because: (i) the three major peaks (interstadials) in isotope stage 7 of the marine $\delta^{18}O$ curve are matched by three peaks in the DH 2 curve, with the major stadial occurring between the two oldest peaks; (ii) on both the marine and DH 2 curves, this major stadial is about 60% of the magnitude of a full glacial–interglacial excursion (if termination II is a guide); and (iii) on both the marine and DH 2 curves, this major stadial occurred within about 25,000 years, roughly one precession cycle. If the proposed correlation is correct, then the DH 2 record indicates that this stadial occurred at least 23,000 years before the date indicated by the marine $\delta^{18}O$ curve. The insolation curves (Fig. 5) cannot account for a severe stadial at about 248 ka; instead these curves imply that a deglaciation should have occurred.

Termination III is poorly defined in comparison to terminations II and I on both the marine and the DH 2 $\delta^{18}O$ curves. It occurs at about 272 ka on the DH 2 curve (Figs. 3 and 4), or about 28,000 years earlier than indicated by the marine data.

In summary, the difference in timing of key features in the marine and DH 2 curves increases with time from perhaps a few thousand years at about 65 ka to 17,000 years at termination II, to about 28,000 years at termination III (Table 2). Because of the poorly known ground-water residence times in the Ash Meadows basin, the cited differences are minima.

Several factors affecting the DH 2 record could account for these differences. Underflow into the Ash Meadows basin from the northeast (Pahranagat Valley, Fig. 1) may constitute as much as 35% of the discharge at Ash Meadows (*2*). If Pahranagat Valley ground waters were about 100,000 to 120,000 years older than those derived from the Spring Mountains and Sheep Range, then the early (that is, pre-130 ka) date of termination II_{GB} could reflect fortuitous mixing of termination III_{GB} waters (from Pahranagat Valley) with pre–termination II waters from the rest of the Ash Meadows basin to yield a single early $\delta^{18}O$ shift (II_{GB}). This scenario is unlikely because no waters in the Ash Meadows basin nor in Pahranagat Valley are now older than 30 to 40 ka.

We believe that there is not a systematic error in the dating of the DH 2. The ^{230}Th and ^{234}U dates are concordant and were from closely spaced samples across the calcite vein. Also, our data are consistent with a variety of ^{230}Th-^{234}U–dated proxy paleoclimatic records that also indicate a pre-140-ka age of termination II (*26–28*).

Might the $\delta^{18}O$ record of DH 2 reflect regional but not global climatic conditions? We find it difficult to believe that the close resemblance of the DH 2 $\delta^{18}O$ time series to the marine and Antarctic records is a coincidence. Furthermore, $\delta^{18}O$ data from a second vein sample resembles the marine $\delta^{18}O$ time series back to at least 500 ka (*29*).

Which of the three $\delta^{18}O$ chronologies (Fig. 4) provide an accurate estimate of the timing of termination II? The Antarctic ice core from Vostok was dated with a two-dimensional ice-flow model (*13*). Such dating is independent of isotopic or orbital chronologies, but involves assumptions about rates of snow accumulation and the amount that ice is thinned with burial. The estimated age of the bottom of this core (165 ka) has an error of 10,000 to 15,000 years (*13*); because this indirect dating has not been checked with another dating technique, we consider the Antarctic record to be the least reliable. The marine record has been dated by both direct and indirect means, including a combination of ^{231}Pa-^{230}Th dating, assumptions of constant sedimentation rate between dated horizons, and correlation with high sea-level terraces on Barbados (*19, 27*). Combined ("stacked") oxygen isotope records from various oceans have been "tuned" to variations of solar insolation (*30*). The final chronology is be-

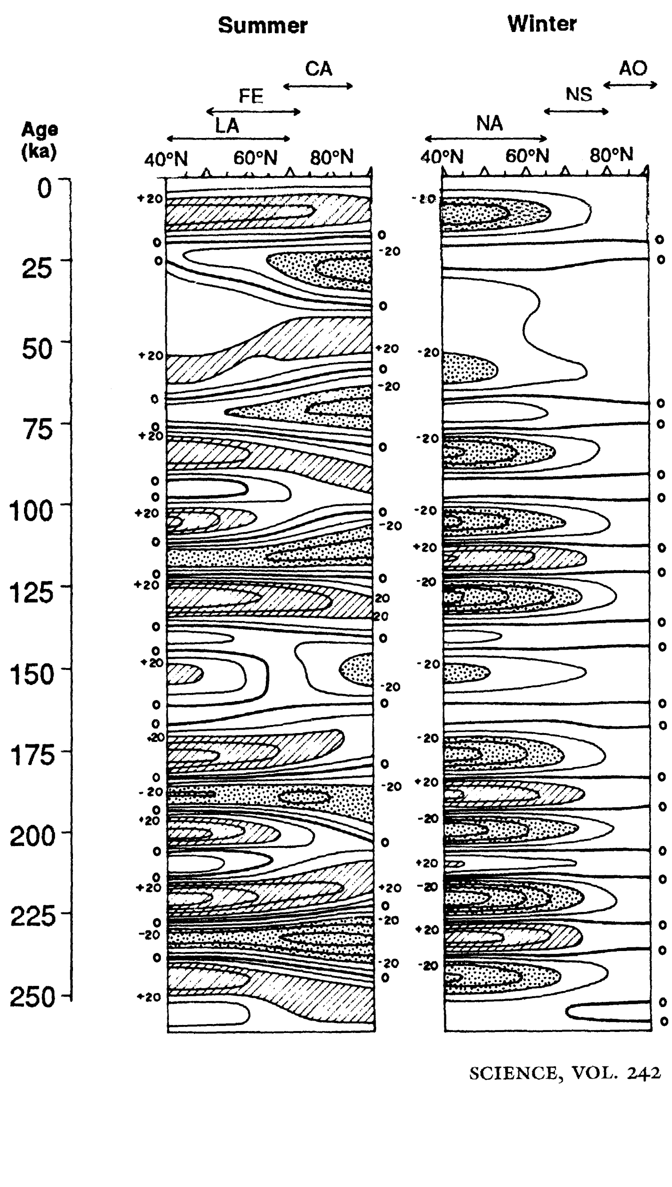

Fig. 5. Variations of insolation during the last 250,000 years with latitude in the Northern Hemisphere. Insolation values are shown as departures from modern values for the winter and summer caloric half-years. Diagonal pattern, relatively high insolation; dotted pattern, relatively low insolation. CA, circum Arctic; FE, Fennoscandian; LA, Laurentide; AO, Arctic Ocean; NS, Norwegian Sea; NA, North Atlantic Ocean. [From (*22*), after (*23*)]

lieved to have an average error of ±5000 years (*30*).

In contrast, we have dated the DH 2 record directly by two radioactive clocks that yield concordant ages. We consider it to be the best dated of the three records, even though it yields only minimum ages for key climatic events. Furthermore, a variety of other proxy records—including direct ^{230}Th-^{234}U dating of marine core V28-238, which yielded a date of 145 ± 9.5 ka for termination II—supports the DH 2 chronology (*26*); we conclude that the accuracy of the marine δ^{18}O chronology is questionable, particularly for the age of termination II.

Alternatively, might the marine and DH 2 chronologies both be correct, with the marine chronology lagging the continental record (DH 2) by at least 17,000 years? A lag of this magnitude appears unlikely. While ice-sheet influence on surface temperature and storm tracks (*31*)—and subsequently on δ^{18}O values of Great Basin precipitation—is likely to have been more rapid than its influence on the δ^{18}O of the oceans, a lag of 17,000 years between continental and the marine effects during deglaciation appears implausible; because of residence times, this lag is a minimum value. Still another possibility is that the marine δ^{18}O data are not a reliable indicator of all major changes of sea level. Johnson and Andrews (*28*) have suggested that termination II of the marine δ^{18}O record does not indicate an abrupt rise of sea level (because of melting of continental ice sheets), but rather an abrupt melting of vast Antarctic floating ice shelves, which like the continental ice sheets had a low δ^{18}O. They invoked this mechanism to explain the major discrepancy between sea-level curves, which suggest that there was a major rise before 140 ka, and the marine δ^{18}O record.

Regardless of the explanation for the discrepancy between the marine and DH 2 δ^{18}O chronologies, a reexamination of the Milankovitch theory of Pleistocene climates is in order because the DH 2 data, as well as other data (*26*), indicate that a major warming of climate occurred before 140 ka. If the astronomic projections of millenial variations in solar insolation are correct, then termination II_{GB} on the DH 2 curve is apparently not directly related to insolation changes. The insolation curves (Fig. 5) show no extreme deviations from average insolation between about 140 and 170 ka, and hence no apparent cause for termination II_{GB} as displayed by DH 2. Because of ground-water residence times, we leave open the possibility that the relatively high summer and low winter insolation centered at 175 ka (Fig. 5) may have caused termination II_{GB}, but if they did have an effect, the strong 175-ka deglaciation signal should have affected the marine record and, conversely, the 130-ka insolation pulse should have affected the Great Basin record. Additionally, there is no evidence on the insolation curves (Fig. 5) for the major stadial centered at 248 ka on the DH 2 curve; instead, they suggest that a deglaciation should have begun at this time.

We conclude, as have others (*26, 28*), that the Milankovitch theory, as currently formulated, does not account for the timing of several major climatic events from 100 to 300 ka and the pre-135 ka sea-level highs (*26*). However, the DH 2 proxy record does contain periodicities close to those of obliquity and precession and in this respect is supportive of insolation variations as a factor in middle to late Pleistocene climatic change. Causes other than or in addition to insolation, for example, strong internal feedback processes (*32*) involving global temperature, ice volume, bedrock deflection, and ocean currents, may have triggered the last interglacial (termination II) as well as other major climatic shifts displayed by the DH 2 δ^{18}O curve between 100 and 310 ka.

REFERENCES AND NOTES

1. I. J. Winograd and W. Thordarson, *U.S. Geol. Surv. Prof. Pap. 712-C* (1975); W. W. Dudley, Jr., and J. D. Larson, *ibid. 927* (1976).
2. I. J. Winograd and I. Friedman, *Geol. Soc. Am. Bull.* **83**, 3691 (1972); J. K. Osmond and J. B. Cowart, in *Uranium Series Disequilibrium: Application to Environmental Problems*, M. Ivanovich and R. S. Harmon, Eds. (Clarendon, Oxford, 1982), pp. 226–228.
3. I. J. Winograd and A. C. Riggs, *Geol. Soc. Am. Abstr.* **16**, 698 (1984).
4. I. J. Winograd and F. J. Pearson, Jr., *Water Resour. Res.* **12**, 1125 (1976).
5. I. J. Winograd and F. N. Robertson, *Science* **216**, 1227 (1982).
6. R. L. Hay, R. E. Pexton, T. T. Teague, T. K. Kyser, *Geol. Soc. Am. Bull.* **97**, 1488 (1986); I. J. Winograd and G. C. Doty, *U.S. Geol. Surv. Open-File Rep. 80-569* (1980); I. J. Winograd and B. J. Szabo, *ibid. 85-697* (1986).
7. R. J. Hoffman, *U.S. Geol. Surv. Open-File Rep. 88-93* (1988).
8. I. J. Winograd, B. J. Szabo, T. B. Coplen, A. C. Riggs, P. T. Kolesar, *Science* **227**, 519 (1985).
9. L. Niel Plummer, personal communication.
10. Vienna standard mean ocean water (VSMOW) [T. B. Coplen, *Chem. Geol.* 72, 293 (1988)].
11. Ground-water temperatures along the principal flow paths in the Ash Meadows ground-water basin are relatively low, varying from 25° to 35°C. At such temperatures, isotopic exchange between water and rock is unlikely [R. N. Clayton, L. J. P. Muffler, D. E. White, *Am. J. Sci.* **266**, 968 (1968)]; this implies that the δ^{18}O of ground water reaching Devils Hole is identical to that which entered the aquifer in the Spring Mountains and Sheep Range (Fig. 1). The δ^{18}O of calcite precipitated from water is temperature dependent, decreasing approximately 0.2 per mil with each 1°C increase in temperature [I. Friedman and J. R. O'Neil, *U.S. Geol. Surv. Prof. Pap. 440-KK* (1977)]. Even though water temperature has remained constant in Devils Hole since 1932 (*4*), we cannot assume that it has during the past 300,000 years. We dismiss the possibility that variations in ground-water temperature significantly affected the δ^{18}O of DH 2 on two grounds. First, Benson and Klieforth (*12*) reported that the δ^{18}O of ground waters from an area immediately west of that shown in Fig. 1 increased by 1.8 per mil between 18.5 and 9 ka; this time interval spans termination I. The magnitude of δ^{18}O changes across terminations I and II were virtually identical in the marine record (Fig. 4), and nearly identical in the Antarctic record. Accordingly, we assumed that the δ^{18}O changes in ground water across termination II at Devils Hole were also ~1.8 per mil, and we measured an increase in δ^{18}O of 2.0 per mil (Fig. 3). Second, the similarity in the shape of the DH 2 and Antarctic ice core (Vostok) δ^{18}O curves (Fig. 4) provides strong circumstantial evidence that the δ^{18}O variations recorded by DH 2 principally reflect climatic changes, rather than changes in ground-water temperature: of the three records, the Antarctic record is the most direct indicator of climate change because it records variations in atmospheric temperature (*13*) that are unbuffered by passage through the marine or continental hydrosphere.

   There are large differences in the δ^{18}O of winter compared to summer precipitation, especially for middle to high latitudes (*14*). Might the major fluctuations (Fig. 3) reflect long-term variations in the importance of summer versus winter-spring precipitation as a source of ground-water recharge? Recharge from summer precipitation in the southern Great Basin is negligible today (*3*) as it is in other arid to sub-humid terranes [(*12*); M. Sophocleus and C. A. Perry, *J. Hydrol.* **81**, 297 (1985); E. S. Simpson, D. B. Thorud, I. Friedman, *Int. Assoc. Sci. Hydrol. Publ.* **94**, 112 (1972); B. M. Gallaher, thesis, University of Arizona, Tucson (1979)]. On the basis of these studies, we believe that summer recharge is also unlikely to have been important in the past. Possible increased summer monsoonal precipitation (*15*) during a pluvial maximum–interpluvial transition would, we suggest, have been fully intercepted in the soil zone by a likely increase in density of vegetation rather than significantly augmenting recharge.
12. L. V. Benson and H. Klieforth, *Am. Geophys. Union Geophys. Monogr.*, in press.
13. G. Lorius *et al.*, *Nature* **316**, 591 (1985).
14. V. Yurtsever and J. R. Gat, *IAEA Tech. Rep. 210* (1981), p. 103; J. Jouzel *et al.*, *J. Geophys. Res.* **92**, 14739 (1987).
15. W. G. Spaulding, *U.S. Geol. Surv. Water Res. Invest. Rep. 84-4328* (1984); ——— and L. J. Graumlich, *Nature* **320**, 441 (1986).
16. R. L. Weaver, *U.S. Dept. Commer. Weather Bur. Hydrometeorol. Rep. 37* (1962).
17. N. J. Shackleton, *Quat. Sci. Rev.* **6**, 183 (1987).
18. The ground water now moving through Devils Hole is probably of late Wisconsin age because (i) the ^{14}C content of Ash Meadows ground water (*4*), though difficult to interpret quantitatively, indicates that the water age is 10 to 30 ka (*4*); and (ii) the δ^{18}O of calcite currently precipitating in Devils Hole is about +14.0 per mil (T. B. Coplen, unpublished data), a value that closely matches that recorded by DH 2 in the period preceding the start of the last interglacial (155 to 190 ka) (Fig. 3). In contrast, δ^{18}O values in calcite were up to 1.7 per mil heavier during the last interglacial (135 to 145 ka). Post–late Wisconsin ground waters apparently have not yet reached Ash Meadows, presumably because postpluvial recharge constitutes (volumetrically) only a fraction of the pluvial water in storage in this vast aquifer system; that is, the pluvial waters apparently have not yet been flushed out of the aquifer. Because full-glacial conditions existed in this region until about 16 ka (*15*), the ground waters currently discharging at Ash Meadows are, at the least, about this age. Similarly, we assume that during the last interglacial, the residence time was also ~16,000 years. Thus, the transition in DH 2 at 147 ka (see line labeled II_{GB} on Fig. 3) most likely represents a climate change that occurred in the recharge area ~16,000 years earlier. Because late Wisconsin pluvial waters may continue to emerge at Ash Meadows for several thousand years, the residence time might be as long as 20,000 years during and immediately following a glacial-interglacial transition. Residence times were most likely much shorter, perhaps only a few thousand years, during full glacial times and major stadials for two reasons. First, ground-water recharge rates at such times may have been three to ten times as

great as modern recharge [M. S. Bedinger *et al.*, *U.S. Geol. Surv. Open File Rep. 84-738* (1984)]. We consider that the Ash Meadows system is approximately a constant volume system with residence times inversely proportional to recharge rates. Second, during pluvial and stadial climates significant recharge likely also occurred on the numerous ridges of intermediate altitude (<1800 m) (Fig. 1). These ridges, which are outcrops of the regional aquifer, are all much closer to the Ash Meadows oasis than the Spring Mountains and Sheep Range; hence, recharge from these ridges would have a much shorter travel time to Devils Hole.

19. W. S. Broecker and J. van Donk, *Rev. Geophys. Space Phys.* **8**, 169 (1970).
20. M. Ghil and S. Childress, *Topics in Geophysical Fluid Dynamics: Atmospheric Dynamics, Dynamo Theory, and Climate Dynamics* (Springer-Verlag, New York, 1987), tables 12.1–12.3.
21. J. M. Landwehr, personal communication.
22. W. F. Ruddiman and H. E. Wright, Jr., in *North America and Adjacent Oceans During the Last Deglaciation*, vol. K-3 of *The Geology of North America*, W. F. Ruddiman and H. E. Wright, Jr., Eds. (Geological Society of America, Boulder, CO, 1987), pp. 1–12.
23. A. L. Berger, *Quat. Res.* **9**, 139 (1978); in *Milankovitch and Climate*, A. L. Berger *et al.*, Eds. (Reidel, Boston, 1984), pp. 3–39.
24. W. F. Ruddiman and A. McIntyre, *Geol. Soc. Am. Bull.* **93**, 1273 (1982).
25. C. D. Ninkovich and N. J. Schackleton, *Earth Planet. Sci. Lett.* **27**, 20 (1975).
26. Several types of proxy records, in addition to DH 2, support a pre-140-ka initiation of the last interglaciation. Studies of sea-level fluctuations during the middle-to-late Pleistocene indicate a high sea-level stand at or above modern level at 135 to 140 ka, with a somewhat higher stand between 120 to 125 ka [W. S. Moore, in *Uranium Series Disequilibrium: Application to Environmental Problems*, M. Ivanovich and R. S. Harmon, Eds. (Clarendon, Oxford, 1982), chap. 18; T. M. Cronin, *Quat. Sci. Rev.* **1**, 177 (1983); C. E. Stearns, in *Quaternary Dating Methods* (Elsevier, New York, 1984), p. 53; J. Chappell, *Search* **14**, 99 (1983); P. Aharon and J. Chappell, *Paleogeogr. Paleoclimatol. Paleoecol.* **56**, 337 (1986); B. Pillans, in *Sea Surface Studies, A Global View*, R. J. N. DeVoy, Ed. (Croom Helm, London, 1987), chap. 9; L. F. Montaggioni and C. T. Hoang, *Paleogeogr. Paleoclimatol. Paleoecol.* **64**, 79 (1988)]. Moore noted that the 135-ka high sea level occurs in numerous areas and is commonly distinct stratigraphically from the 120-ka terraces; he concluded that the Milankovitch prediction cannot account for the sea-level record at 135 to 140 ka. [For contrasting views see (*27*); A. Kaufman, *Quat. Res.* **25**, 55 (1986); J. Chappell and N. J. Shackleton, *Nature* **324**, 137 (1986); R. L. Edwards, J. H. Chen, T.-L. Ku, G. J. Wasserburg, *Science* **236**, 1547 (1987); R. L. Edwards, J.-H. Chen, G. J. Wasserburg, *Earth Planet. Sci. Lett.* **81**, 175 (1986).] Given that several thousands of years are apparently needed for ice-sheet melting [J. Oerlemans, *Nature* **287**, 430 (1980)], we conclude that considerable sea-level data indicate that a major deglaciation was well under way by 140 ka, or 10,000 years earlier than indicated by termination II of the marine $\delta^{18}O$ chronology. Alternatively, if termination II actually occurred before 147 ka, as the DH 2 record indicates, why have no sea-level highs of say 150 to 170 ka been reported, especially from tectonically active regions, such as New Guinea, where the preservation of coral-bearing terraces is likely? We have no answer for this question. Additionally, the well-dated 125-ka sea-level high is seemingly not reflected in the DH 2 record (Figs. 3 and 4) unless it is recorded by the interstadial centered at about 110 ka, an event which presumably occurred thousands of years earlier in the Spring Mountains recharge area. Also, there are other major discrepancies—besides the 135-ka sea-level high—between sea-level curves and marine $\delta^{18}O$ curves. For example, relatively high sea levels are documented at times (40 to 100 ka) when the marine $\delta^{18}O$ record indicates that sea level should have been low [(*17, 28*); J. E. Mylroie and J. L. Carew, *Quat. Sci. Rev.* **7**, 55 (1988)]. Clearly, much remains to be learned in this area.

The most comprehensive recent attempt at direct dating of a deep-sea core is that of Kominz *et al.* [M. A. Kominz, G. R. Heath, T.-L. Ku, N. G. Pisias, *Earth Planet. Sci. Lett.* **45**, 394 (1979)]. They dated Pacific core V28-238 with ^{230}Th-^{234}U and obtained an age of 145 ± 9.5 ka for termination II and 260 ± 30 ka for termination III. Kominz *et al.* explicitly noted the major discrepancy between their ^{230}Th date and the widely accepted age of 127 ka for termination II.

Searles Lake, just 140 km southwest of Devils Hole, has a paleoclimatic record that extends for 3 million years [J. L. Bischoff, R. J. Rosenbauer, G. I. Smith, *Science* **227**, 1222 (1985); G. I. Smith, V. I. Barczak, G. F. Moulton, J. F. Liddicoat, *U.S. Geol. Surv. Prof. Pap. 1256* (1983)]. The 6/5 isotopic boundary (termination II) of the marine $\delta^{18}O$ record coincides with a major deepening, rather than the expected shallowing of this lake, whereas the 2/1 boundary (termination I) coincides with desiccation of Searles Lake, as expected. Thus, this continental record also does not mimic the marine record. The DH 2 $\delta^{18}O$ record (Figs. 3 and 4) clearly indicates that a major stadial had begun by 127 ka.

In high-latitude caves, periods of speleothem deposition and nondeposition are believed to indicate respectively interglacial and glacial climates. U-series dated speleothems from such caves located in areas overridden by or adjacent to an ice sheet suggest that the last interglacial stage began between 170 and 135 ka [M. Gascoyne, *Stud. Speleol.* **V**, 15 (1984); R. S. Harmon, D. C. Ford, H. P. Schwarcz, *Can. J. Earth Sci.* **14**, 2543 (1977); R. S. Lively, *Geology* **11**, 259 (1983); R. S. Harmon, *Natl. Speleol. Soc. Bull.* **41**, 102 (1979); M. Gascoyne, H. P. Schwarcz, D. C. Ford, *Phil. Trans. R. Soc. Lond. Ser. B* **301**, 143 (1983)].

In summary, there is abundant proxy marine and continental evidence, all ^{230}Th-^{234}U dated, that the transition to interglacial climates occurred sometime between 135 and 170 ka. No one of these proxy records would, by itself, constitute a challenge to the widely accepted 130 ka age for termination II. When considered together, however, they provide strong independent support for the pre-147-ka age for termination II indicated by DH 2.

27. W. F. Ruddiman *et al.*, *Quat. Res.* **21**, 123 (1984).
28. R. G. Johnson and J. T. Andrews, *Paleogeogr. Paleoclimatol. Paleoecol.* **53**, 107 (1986).
29. I. J. Winograd *et al.*, unpublished data.
30. D. G. Martinson *et al.*, *Quat Res.* **27**, 1 (1987).
31. J. E. Kutzbach, in *North America and Adjacent Oceans During the Last Deglaciation*, vol. K-3 of *The Geology of North America*, W. F. Ruddiman and H. E. Wright, Jr., Eds. (Geological Society of America, Boulder, CO, 1987), pp. 425–446.
32. H. LeTreut, J. Portes, J. Jouzel, M. Ghil, *J. Geophys. Res.* **93**, 9365 (1988); M. Ghil, A. Mullhaupt, P. Pestiaux, *Clim. Dyam.* **2**, 1 (1987); B. Saltzman, A. R. Hansen, K. A. Maasch, *J. Atmos. Sci.* **41**, 3380 (1984).
33. We thank R. J. Hoffman for assisting in the SCUBA exploration of Devils Hole, J. M. Landwehr for considerable help with the statistical analysis of our data, and P. T. Kolesar for his petrographic analysis of DH 2. We thank E. Rothfuss, P. Rowlands and D. Sada for permission to conduct research in Devils Hole and C. Lee for ediing and typing. The manuscript was greatly improved by the review comments of G. I. Smith, T. M. Cronin, L. V. Benson, and three anonymous reviewers.

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Io: Evidence for Silicate Volcanism in 1986

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Infrared observations of Io during the 1986 apparition of Jupiter indicate that a large eruptive event occurred on the leading side of Io on 7 August 1986, Universal Time. Measurements made at 4.8, 8.7, and 20 micrometers suggest that the source of the event was about 15 kilometers in radius with a model temperature of ~900 Kelvin. Together with previously reported events, these measurements indicate that high-temperature volcanic activity on the leading side of Io may be more frequent than previously thought. The inferred temperature is significantly above the boiling point of sulfur in a vacuum (715 Kelvin) and thus constitutes strong evidence for active silicate volcanism on the surface of Io.

Since the discovery of volcanic activity on Jupiter's satellite Io (*1*), the nature of the thermal activity has been studied through a combination of analyses of Voyager data and Earth-based infrared telescopic observations. The range of observed temperatures for volcanic areas on Io provides a simple way of separating the major components of infrared radiation coming from Io. Emission from the nonvolcanic, cold areas making up ~99% of Io's surface accounts for most of the flux in broadband measurements centered at 20 μm; the flux from volcanic areas with temperatures of 250 to 400 K dominates narrowband 8.7-μm measurements. Reflected sunlight with a small contribution from volcanic hotspots, primarily areas with temperatures ≳500 K, dominates the 4.8-μm spectral region. Because of these characteristics, the temporal and spatial variability of volcanic emissions from Io can be studied through measurement at several wavelengths of the total infrared flux from Io's disk as a function of time and orbital position (because Io is synchronously rotating, orbital position relates directly to the sub-Earth longitude on Io's surface).

Earlier studies of Io's thermal emissions

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