

Correlations between O^{18}/O^{16} Ratios and Chemical Compositions of Tektites¹

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The limits of oxygen isotope variation in tektites have been determined by analyzing 33 tektites that cover the widest possible spectrum of chemical compositions. δO^{18} values were obtained for 7 australites (8.9 to 11.0), 4 javaites (9.4 to 10.0), 3 indochinites (9.3 to 10.6), 5 philippinites (9.8 to 10.6), 8 moldavites (10.7 to 11.8), and 6 Ivory Coast tektites (13.0 to 13.6), as well as 2 samples of Darwin glass (14.4 to 14.9). Various tektite groupings based on chemical composition and geographic occurrence all show a systematic increase in O^{18} with decreasing SiO_2 content, similar to that previously found for the bediasites. These systematic correlations must arise either by (1) vapor fractionation of tektite material during impact melting or (2) mixing of a SiO_2 -rich igneous component and a low- SiO_2 component formed at a much lower temperature; perhaps the mixing occurred during weathering at the earth's surface or during hydrothermal alteration of silicic igneous rocks either on the earth or the moon.

INTRODUCTION

In three previous papers [Taylor and Epstein, 1962b; 1964; 1966] we showed that the O^{18}/O^{16} ratios of tektites from the Australasian, North American, and Czechoslovakian strewn fields are all very similar ($\delta O^{18} = 8.9$ to 11.3% , relative to SMOW, standard mean ocean water), whereas the Ivory Coast tektites are distinctly higher in O^{18}/O^{16} . In addition, samples covering essentially the entire known range of chemical variation in the Texas tektites (bediasites) were analyzed, and a simple inverse correlation between SiO_2 content and O^{18}/O^{16} ratios was found.

It has recently become possible to obtain tektites displaying a wide range in chemical composition from the strewn fields in Australia, Java, the Philippine Islands, Indochina, Czechoslovakia, and the Ivory Coast. These new tektite samples were analyzed for O^{18}/O^{16} and the data are given in Table 1 (see the references noted above for details of the analytical procedures). Also included in Table 1 are the first oxygen isotope analyses of Darwin glass from Tasmania (see Chapman *et al.* [1967] for a

description of this material) and of impactites from Meteor Crater, Arizona.

DISCUSSION OF THE ISOTOPE DATA

Relation to chemical groupings. The δO^{18} values of tektites are plotted as a function of index of refraction in Figure 1. Also plotted are a few previous analyses, including the bediasite data referred to above. The various tektite groups are delineated in Figure 1 principally according to geographic occurrence, but also with respect to the chemical groupings discovered by Chapman and Scheiber [1969].

The chemical groupings of Chapman and Scheiber are very nicely substantiated by the oxygen isotope data. Their HMg group, HCa australites, HNa/K australites, HCa philippinites, and HCu-HB indochinites each form a distinct grouping on a δO^{18} -refractive index diagram like Figure 1. Also, the two geographically separated sets of moldavites, from Bohemia and Moravia, plot in slightly different positions in Figure 1.

Note that *all* the tektite groups (except the Ivory Coast tektites) show a consistent positive correlation between index of refraction and δO^{18} . In the light of the well-established linear correlation between SiO_2 content and refractive index of tektites [Barnes, 1940], this implies that for each tektite group the higher SiO_2 ,

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TABLE 1. Oxygen Isotope Ratios of Tektites, Darwin Glass, and Impactites from Meteor Crater, Arizona

Sample Number and Locality	$\delta O^{18}, \text{‰}^*$	Index of Refraction	Type
Tektites			
1. AN-288 Snake Creek bore, Australia	8.93 \pm 0.17 [†] (4) [‡]	1.492	HCa
2. AN-148 Bloods Creek bore, Australia	9.14 (1)	1.501	HCa
3. T-224 Site A, Thailand	9.33 (1)	1.487	HCu-HB
4. J74 Sangiran, Java	9.41 \pm 0.10 (3)	1.496	HMg
5. J35 Sangiran, Java	9.43 \pm 0.03 (2)	1.513	HMg
6. AN-61 Lake Margaret, Australia	9.60 (1)	1.520	HMg
7. P192 Pasuquin, Luzon, Philippine Is.	9.81 \pm 0.04 (2)	1.525	HMg
8. J15 Sangiran, Java	9.92 (1)	1.529	HMg
9. J2 Sangiran, Java	9.96 \pm 0.14 (3)	1.539	HMg
10. AN-157 Bloods Creek bore, Australia	10.00 \pm 0.04 (2)	1.521	HCa
11. P290 Babalog, Luzon, Philippine Is.	10.04 \pm 0.09 (2)	1.506	HCa
12. Dalat, South Vietman DTM composite	10.27 \pm 0.05 (6)		
13. P195 Pangranayan, Philippine Is.	10.30 \pm 0.07 (2)	1.519	LCa/Al
14. P363 Santa Mesa, Luzon, Philippine Is.	10.40 (1)	1.525	HCa
15. P197 Ortigas site, Philippine Is.	10.60 \pm 0.08 (3)	1.534	HCa
16. T-220Dk Ubol, Thailand	10.64 \pm 0.02 (2)	1.513	HCu-HB
17. T5322 Slavetice, Moravia	10.68 \pm 0.07 (4)	1.489	HRb/Sr
18. AN-318 S. of Lake Wilson, Australia	10.69 \pm 0.02 (2)	1.531	HNa/K
19. AN-87 Williams Creek, Australia	10.76 \pm 0.14 (3)	1.542	HNa/K
20. T5324 Dukovany, Moravia	10.95 \pm 0.04 (3)	1.492	HRb/Sr
21. AN-245 Raspberry bore, Australia	11.03 \pm 0.07 (3)	1.536	HNa/K
22. T5314 Slavce, Bohemia	11.05 \pm 0.18 (4)	1.488	HRb/Sr
23. T4574 Radomilice, Bohemia	11.16 \pm 0.06 (3)		HSi
24. T5319 Nechov, Bohemia	11.25 \pm 0.05 (3)	1.489	HRb/Sr
25. Lhenice, Bohemia [Taylor and Epstein, 1966]	11.38 (1)	1.492	
26. T5317 Kroclov, Bohemia	11.70 \pm 0.03 (3)	1.494	LRb/Sr
27. T5296d Lhenice, Bohemia	11.81 \pm 0.09 (3)	1.495	LRb/Sr
28. T5658 Ivory Coast (2.93% MgO)	13.02 (1)		
29. T5702 Ivory Coast (2.86% MgO)	13.16 (1)		
30. T5653 Ivory Coast (4.26% MgO)	13.22 (1)		
31. T5649 Ivory Coast (3.85% MgO)	13.24 (1)		
32. T5686 Ivory Coast (4.20% MgO)	13.43 (1)		
33. T5706 Ivory Coast (3.56% MgO)	13.55 (1)		
Darwin Glass			
34. DG 3 Ten Mile Hill, Mt. Darwin, Tasmania	14.36 (1)	1.485	Dark
35. DG 24 Ten Mile Hill, Mt. Darwin, Tasmania	14.93 (1)	1.476	Light
Shocked Cocouino Sandstone, Meteor Crater			
36. SS-2, frothy lechatelierite, shaft 2, 105-ft depth	12.57 \pm 0.03 (2)		H
37. SS-3, bottom of crater, main shaft	12.57 \pm 0.04 (2)		H
38. SS-4, original Coconino sandstone	12.84 \pm 0.05 (2)		U
39. SS-6, south talus slope, bedding retained	13.02 \pm 0.02 (2)		L
40. SS-1, moderately shocked, south talus slope	13.81 (1)		M

* Oxygen isotope analyses are relative to standard mean ocean water (SMOW) in per mille.

[†] Analytical error shown is average deviation from the mean.

[‡] Numbers in parentheses indicate number of separate determinations.

Notes.

Samples 1 through 16, 18, 19, 21, 34, and 35 were obtained from Dean R. Chapman, who also supplied the indices of refraction and the chemical classifications, as follows: HCa, high calcium australites or philippinites; HMg, high magnesium group; HCu-HB, high copper, high boron indochinites; LCa/Al, philippinite with low Ca : Al ratio; HNa/K, australites with high Na : K ratios.

Samples 17, 20, 22 through 27, and 28 through 33 were obtained from W. H. Pinson, Jr., and the indices of refraction and chemical classifications are taken from *Philpotts and Pinson* [1966] and *Schnetzler and Pinson* [1964]: HRb/Sr, moldavites with high Rb : Sr ratios; LRb/Sr, moldavites with low Rb : Sr ratios; HSi, moldavite with highest SiO₂ content of any known tektite (84.5 wt%).

Samples 36 to 40 were obtained from Eugene M. Shoemaker and Susan Kieffer. H, highly shocked; M, moderately shocked; L, lightly shocked; U, unshocked parent material.

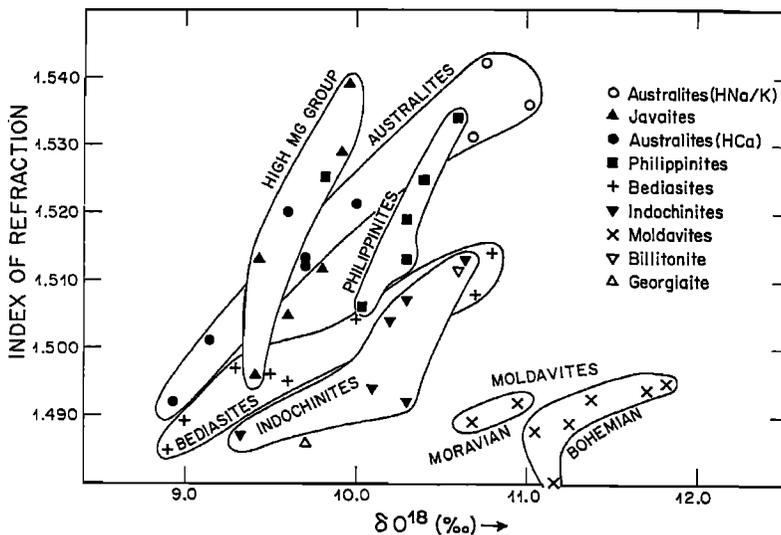


Fig. 1. Plot of δO^{18} versus index of refraction for the tektites analyzed in the present study. Also included are data on 8 bediasites, 2 javaites, 2 australites, 2 philippinites, 3 indochinites, a georgiite, and a billitonite from *Taylor and Epstein* [1964].

contents are associated with lower O^{18}/O^{16} ratios. Besides these systematic variations in δO^{18} within each group, the new data indicate that the total range in δO^{18} in every one of the Australasian tektite groups is within the range shown by the bediasites and the australites. It is truly remarkable that the O^{18}/O^{16} variation exhibited by the 700,000-yr-old australites is exactly the same as that shown by the 33-m.y.-old bediasites.

The HMg tektite group shows very little variation in O^{18}/O^{16} over a wide range of SiO_2 content, changing only from 9.41 to 9.96, whereas the SiO_2 content changes by almost 15 wt %. The HCa philippinites show a similarly small variation in δO^{18} , but all other tektite groupings show a much more rapid increase in O^{18} with decreasing SiO_2 , notably the bediasites, moldavites, and indochinites. Inasmuch as essentially the entire range of SiO_2 contents in the various tektite strewn fields are included in Figure 1, we believe that the correlations shown are a fundamental feature of tektite populations, and they are unlikely to be seriously modified by future work. For the same reason, it is considered unlikely that any other Australasian, North American, or Czechoslovakian tektites will be found to have O^{18}/O^{16} ratios appreciably different from those plotted in Figure 1. In other words, this figure probably

indicates the *limits* of δO^{18} in the three major tektite fields.

Unfortunately, the newly discovered microtektites [Glass, 1967] are too small to be readily analyzed for O^{18}/O^{16} , and, because their SiO_2 contents range to lower values than those of land tektites do, the actual range of δO^{18} might be larger than that shown in Figure 1. We may note, however, that the 'bottle-green' microtektites [Cassidy *et al.*, 1969] are simply an extension of the HMg group of *Chapman and Scheiber* [1969]; even if the HMg trend shown in Figure 1 is extrapolated to SiO_2 contents as low as 48 wt % (the lowest value known for a microtektite), the range of δO^{18} values of the HMg-'bottle-green' group would be within the 8.9 to 11.0‰ range shown by the other Australasian tektites. When sufficient microtektite material is available for analysis, it would be highly desirable to attempt O^{18} analyses on composite samples or perhaps even on unusually large individual specimens.

The new analyses of Ivory Coast tektites are very similar to one another, and they lie within the range of O^{18}/O^{16} previously determined for Ivory Coast samples [Taylor and Epstein, 1966]. These new samples cover the entire spectrum of chemical variations found in a suite of Ivory Coast tektites analyzed by W. H. Pinson. Their chemical variations are much

smaller than in the tektites of the major strewn fields listed in Table 1, however, so the small range (only 0.5‰) in O¹⁸/O¹⁶ is not unexpected. The SiO₂ contents of the Ivory Coast specimens, for example, vary from only 66.8 to 68.6 wt % [Pinson and Kolbe, 1969]. Note that the two Ivory Coast tektites that are lowest in MgO also have the lowest δO^{18} values. Although this trend is not completely clear, it is similar to the rather definite trends observed in the three major tektite groups.

Comparison with terrestrial impactite glass. In Figure 2, the tektite data are compared with available oxygen isotope analyses of various forms of terrestrial impactite glass and other glass of unknown origin (Libyan glass and Darwin glass). The impactite glasses cover a wide spectrum of SiO₂ contents and O¹⁸/O¹⁶ ratios, just as would be expected as a result of random fusion of the earth's surface materials by meteorite impact. The Ivory Coast tektites are distinctly different in O¹⁸/O¹⁶ from the other tektites, and it has been suggested that they are terrestrial impact glasses ejected from Bosumtwi crater [Cohen, 1961; Schnetzler *et al.*, 1966; Taylor and Epstein, 1966].

The oxygen isotope ratios of Darwin glass are typical of values obtained for terrestrial sandstones, shales, or most sedimentary and metasedimentary rocks. In terms of their O¹⁸/O¹⁶ ratios the Darwin glass samples have nothing in common with the nearby Australian tektites. In fact, on a δO^{18} refractive index diagram they exhibit a trend opposite to that shown by the australites (Figure 2).

Isotopic analyses of impact lechatelierite and shocked sandstone from Meteor Crater, Arizona, are practically identical with the parent Coconino sandstone. This result, added to the data presented by Taylor and Epstein [1964], indicates that impact melting at moderate temperatures produces little change in the O¹⁸/O¹⁶ ratios of the fused materials.

Comparison with igneous rocks. In Figure 3 we show the δO^{18} variations in tektites as a function of chemical composition, employing the quantity $\frac{1}{3} SiO_2 + K_2O - CaO - MgO - FeO$ (Larsen 'differentiation index'). Data on terrestrial igneous rocks from Figure 20 of Taylor [1968] are also plotted in Figure 3. The trends shown by the various tektite groups are completely different from the trends shown

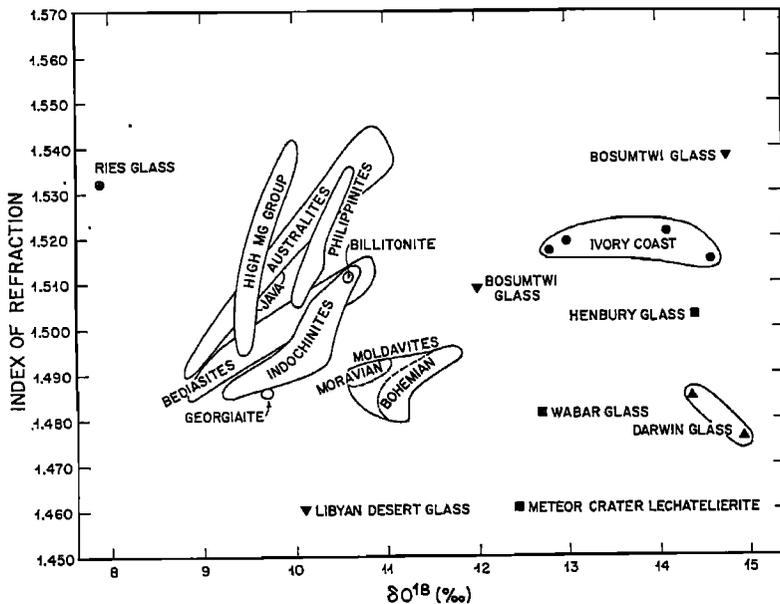


Fig. 2. Plot of δO^{18} versus index of refraction for Ivory Coast tektites and various samples of terrestrial impactite glass [Taylor and Epstein, 1962b; 1964; 1966]. These data are compared with the tektite fields delineated in Figure 1 and with the analyses of Darwin glass and Meteor Crater lechatelierite.

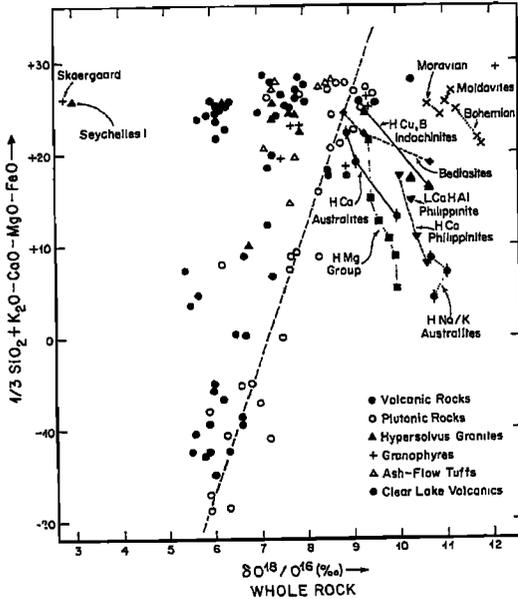


Fig. 3. Plot of δO^{18} versus Larsen 'differentiation index' for tectites and igneous rocks. The data on terrestrial igneous rocks are from Figure 20 of Taylor [1968]. The chemical data on the tectites were kindly supplied by D. R. Chapman.

by most of the igneous rocks. However, with increasing Larsen index, the tectite trends and igneous rock trends approach one another.

Isotopically, the SiO_2 -rich tectites are practically indistinguishable from SiO_2 -rich igneous rocks such as granites, granodiorites, granophyres, and many rhyolitic ash-flow tuffs. Obsidians and a rhyodacite from the Clear Lake, California, volcanic field also plot in the same range of δO^{18} , but most rhyolite obsidians have distinctly lower O^{18}/O^{16} ratios.

During igneous differentiation on the earth, O^{18}/O^{16} ratios either remain essentially constant, as in basalt-andesite-rhyolite associations, or they increase by only 2 to 3%, as in gabbro-tonalite-granodiorite provinces like the Southern California batholith (dashed line of Figure 3). We therefore conclude that, if the O^{18}/O^{16} ratios and chemical compositions of tectites reflect the nature of their parent materials before impact melting, these parent materials cannot be a suite of igneous rocks related to one another by simple magmatic differentiation. All known meteorites have δO^{18} values lower than 6.5, except types 1 and 2 carbonaceous chondrites and ureilites [Reuter *et al.*, 1965; Taylor

et al., 1965], so that we can extend the above conclusion to include the igneous meteorites as well as terrestrial igneous rocks.

Nevertheless, the coincidence of δO^{18} in the high- SiO_2 tectites and igneous rocks strongly suggests that a significant portion of the tectite parent materials are of igneous derivation, in particular the high- SiO_2 part. Igneous quartz is thus a very likely constituent of tectites, as there is abundant evidence (lechatelierite particles) that quartz grains were present in tectite parent materials.

Comparison with altered volcanic rocks, soils, and sediments. In Figure 4 the δO^{18} values of tectites are compared with those of terrestrial soils, shales, phyllites and pelitic schists (metamorphosed shales), and various SiO_2 -rich volcanic rocks. Observe that, except for a few of the altered volcanic flows and tuffs, the entire range of samples from the three major tectite fields is fairly well delineated from most of the terrestrial rocks plotted in Figure 4. It is very common for volcanic rocks, particularly ash-flow tuffs, to have undergone a low-temperature alteration process. This alteration occurs during and after cooling of the volcanic units as a result of interactions with late-stage gasses and aqueous solutions that pass through the rocks and exchange oxygen with the primary minerals and glass. The effects of this alteration process are (1) hydrated volcanic glass, (2) devitrified volcanic glass, and (3) deuteric or hydrothermal alteration of primary minerals, producing chlorite, epidote, clay minerals, etc. Because oxygen isotope fractionations between minerals and H_2O increase sharply with decreasing temperature, both surface weathering and low-temperature hydrothermal alteration usually produce an O^{18} enrichment in the affected rocks.

Except in localities where the meteoric waters are very low in O^{18} (only at high latitudes or high elevations [Craig, 1961]), the weathering of an igneous rock always produces an increase in O^{18}/O^{16} ratio in the rock. Weathered igneous rocks or mixtures of soils and fresh igneous rocks therefore qualify as parent materials of tectites on the basis of their O^{18}/O^{16} ratios, as pointed out previously by Taylor and Epstein [1964]. However, again excepting situations in which low- O^{18} meteoric waters are involved, hydrothermally altered igneous rocks are equally suitable tectite parent materials.

MIXING CURVES

Least-squares trend lines. The above discussion indicates that tektites could have formed from a parent material composed of a mixture of (1) fresh SiO₂-rich igneous rock and (2) a low-temperature alteration product of such rocks. Thus it is interesting to inquire more fully whether the oxygen isotope trend lines shown in Figures 1 and 3 could be mixing lines involving an SiO₂-rich component and an SiO₂-poor component.

In Figure 5, least-squares lines have been drawn through the data points for each tektite group. Included is a least-squares line through the bediasite data of *Taylor and Epstein* [1964]. Note that one peculiar moldavite has been discarded in deriving the least-squares line for the Bohemian moldavites; this sample has the highest SiO₂ content of any known tektite [*Schnetzler and Pinson*, 1964], and it clearly departs from the rather consistent trend exhibited by the other Bohemian tektites.

Not all of the tektites plotted in Figure 5 have been chemically analyzed, but, in view of the systematic trends shown, it is useful to plot those with known SiO₂ content on a $\delta^{18}\text{O}$ -SiO₂ diagram. New least-squares lines are drawn through these more restricted sets of data points in Figure 6, and the lines are extrapolated to higher and lower SiO₂ contents. There is a rough tendency for the least-squares lines to converge in the high-SiO₂ part of the diagram, and they markedly diverge at low-SiO₂ contents. These relationships are exactly what would be predicted if the lines represented mixing curves between (1) a high-SiO₂ igneous component having a relatively uniform $\delta^{18}\text{O}$ of about 8.0 to 10.0‰ and (2) a low-SiO₂ low-temperature component having a highly variable $\delta^{18}\text{O}$ of about 11 to more than 20‰.

Quartz is known to be very resistant to exchange and recrystallization in the presence of low-temperature aqueous fluids [*Taylor*, 1968; *Clayton et al.*, 1968], whereas feldspars and

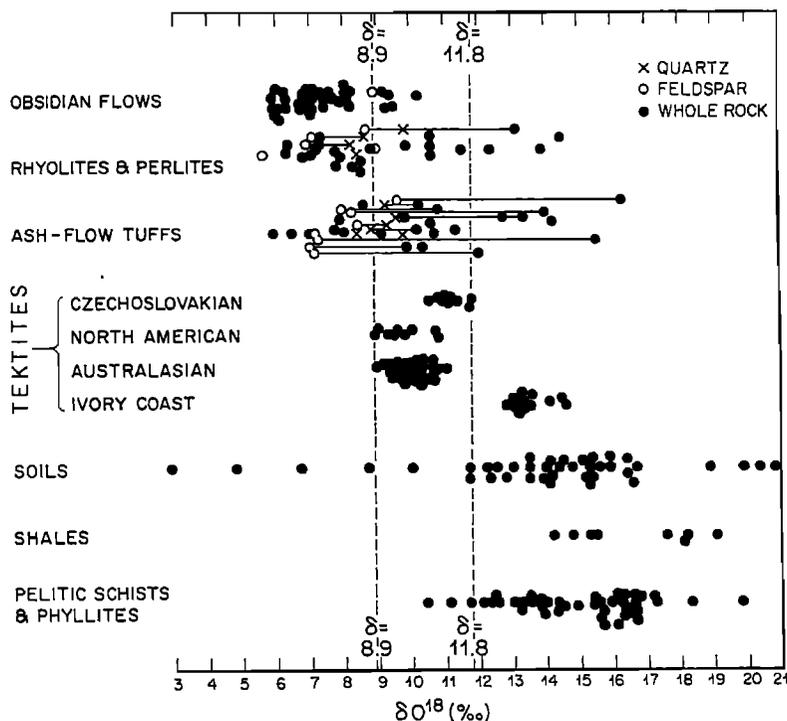


Fig. 4. Comparison of $\delta^{18}\text{O}$ data on tektites with data on silicic volcanic rocks [*Taylor*, 1968], soils [*Taylor and Epstein*, 1964; unpublished data by J. R. Lawrence], and sedimentary and metasedimentary rocks [*Taylor and Epstein*, 1964; *Garlick and Epstein*, 1967; *Savin and Epstein*, 1969]. Practically all of the volcanic rocks having $\delta^{18}\text{O}$ values higher than 9.5 have undergone O¹⁸ enrichment during low-temperature hydrous alteration.

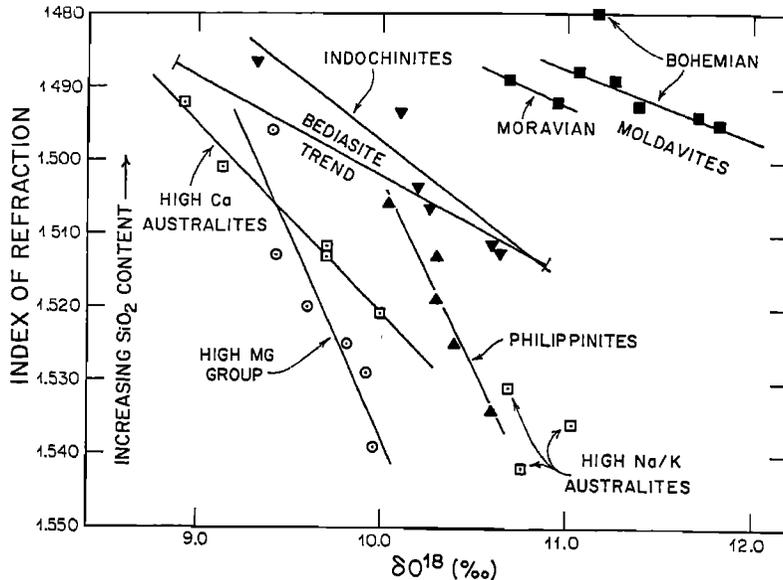


Fig. 5. Plot of $\delta^{18}\text{O}$ versus index of refraction for tektites, showing least-squares lines drawn through the data points obtained in the present study. Also included is the least-squares line through the bediasite data of Taylor and Epstein [1964].

mafic minerals are very susceptible to oxygen isotope exchange during hydrothermal alteration or weathering in forming chlorite, kaolinite, epidote, and other hydrous alteration minerals. The effects of such processes on granitic or rhyolitic igneous rocks would be to produce variations on a $\delta^{18}\text{O}$ - SiO_2 diagram that are similar to those shown in Figure 6. It should be noted that, if the trend lines of Figure 6 are extrapolated all the way to 100% SiO_2 (pure quartz), $\delta^{18}\text{O}$ values of 7 to 9‰ are obtained. This is a little lower than the value for typical quartz in plutonic granitic rocks but is almost identical with the quartz in most terrestrial rhyolites [Taylor and Epstein, 1962a; Taylor, 1968].

Possible mineralogy of tektite parent materials. If we accept the possibility that the least-squares lines of Figures 5 and 6 are mixing curves of the type outlined above, it is instructive to consider possible mineralogical compositions of tektite parent materials. Let us make the simplifying assumptions that the only 'igneous' mineral present is quartz and that the other minerals are all typical low-temperature alteration products. We also ignore any chemical variations arising from vapor fractionation of alkalis, SiO_2 , etc., that might occur during the formation of tektite glass. If we do

this, and if we remove all H_2O because of the anhydrous nature of tektite glass, then a kind of 'norm' can be calculated for any tektite chemical analysis; some typical examples are shown in Table 2, along with tabulations for a few microtektites that are very unusual in chemical composition.

Table 2 is useful primarily to test whether the kinds of mixing curves postulated on the basis of the oxygen isotope data are at all feasible. For example, if quartz is assigned a $\delta^{18}\text{O}$ of 8.0, the $\delta^{18}\text{O}$ of the rest of the assemblage can be calculated by material balance, utilizing appropriate $\text{O}^{18}/\text{O}^{16}$ ratios for the various tektites. When this is done, we obtain the following δ values for the nonquartz portion of the various tektites in Table 2: moldavite, 15.4; bediasite, 11.9; australite, 11.3; J86 HMG tektite, 11.4‰. Except for the moldavite, these values are considerably lower than those of weathering products formed at the earth's surface [Savin and Epstein, 1969]. This means that either (1) primary igneous minerals other than quartz are present in appreciable quantities in the tektite parent materials along with the weathering products or (2) the alteration process occurred at higher temperatures than that of the earth's surface.

In particular, observe that chemical variations in the HMg 'bottle-green' tektite group can be largely explained by adding chlorite and epidote and subtracting muscovite (or K feldspar) from the more 'normal' tektites. This is in the right direction to explain the steeper slope of the HMg least-squares lines in Figures 5 and 6 because in a given mineral assemblage chlorite and epidote are invariably lower in O¹⁸ than coexisting muscovite or K feldspar [Garlick and Epstein, 1967]. Similarly, this might also explain the steeper slope of the least-squares line for the HCa philippinites shown in Figures 5 and 6.

It is worth pointing out that many, if not most, granophyres that have apparently formed as late-stage differentiates of large layered gabbro plutons or diabase sills have δO^{18} values similar to tektites. As has been shown by Taylor [1967; 1968], these granophyres typically have

undergone a deuteric (hydrothermal) alteration that has enriched the whole rock in O¹⁸ while leaving the quartz nearly unaffected. Granophyres from the Precambrian Muskox and Duluth complexes in particular seem to have all the attributes of the postulated tektite parent material, at least in terms of the oxygen isotope relationships. They contain, however, exchanged alkali feldspar instead of muscovite, as do the examples in Table 2, and of course their quartz contents are lower than the average bediasite, moldavite, and australite shown in Table 2.

VAPOR FRACTIONATION IN TEKTITES

Walter [1967] has presented evidence that much of the chemical variation in tektites can be explained by fractional vaporization of the tektite liquid at the time of impact fusion. A single experiment by Walter and Clayton [1967] indicated that fusion of a sample containing

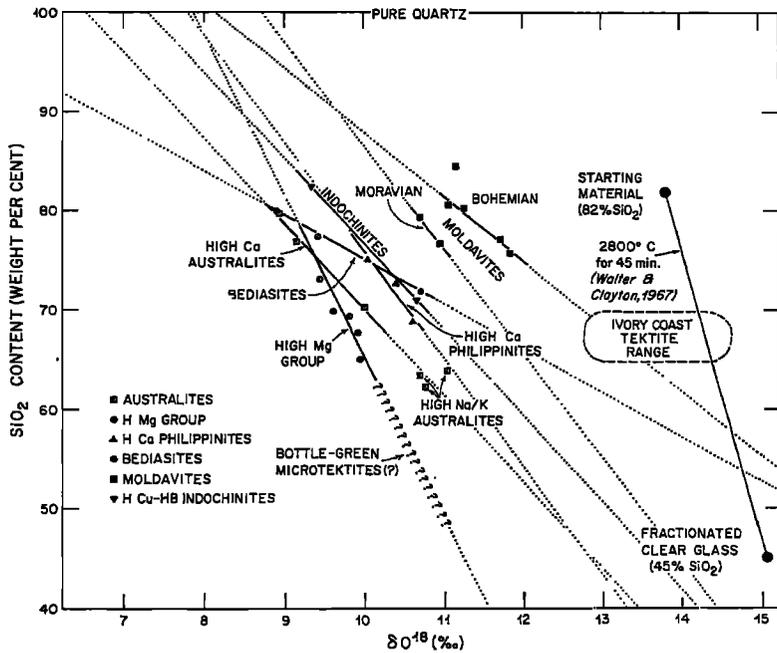


Fig. 6. Plot of δO^{18} versus SiO_2 content, showing least-squares lines drawn through data points for which chemical analyses are available [Chapman and Scheiber, 1969; Schnetzler and Pinson, 1964; Philpotts and Pinson, 1966]. The least-squares trend lines are very similar to those of Figure 5, which are based on a larger number of data points. Note that the bediasite line is drawn through data points for only two end member samples, BM-1 and B-90 [see Taylor and Epstein, 1964; Chao, 1963]. The lines are extrapolated toward higher and lower SiO_2 contents, illustrating the possibility that they represent mixing lines involving a high- SiO_2 , low- O^{18} component and a low- SiO_2 , high- O^{18} component. The lines are, however, qualitatively similar in slope to a vapor fractionation curve obtained by Walter and Clayton [1967] during 2800°C fusion of a high- SiO_2 glass.

TABLE 2. Calculated Mineralogical Compositions of Tektites (Weight per cent)

Mineral	Tektites				Microtektites		
	Average* Bediasite	Average† Moldavite	Average‡ Australite	HMg Tektite‡ J86	#63§	#64§	#66§
Quartz	56	57	49	41	10	25	29
Chlorite	2	1	3	18	59	45	10
Epidote¶		6	11	10	19	12	19
Muscovite**	17	24	18	12	1	3	5
Albite	13	4	11	7	4	5	9
Magnetite††	4	3	5	9	5	8	4
Kaolinite	6	3	1	1			22
Sphene	2	2	2	2	2	2	2

* Table 5, *Chao* [1963].

† Table 4, *Schnetzler and Pinson* [1963].

‡ Table 6, *Cassidy et al.* [1969].

§ Table 4, *Cassidy et al.* [1969], two 'bottle-green' samples (#63 and #64), and one high Al₂O₃ sample (#66).

|| Chlorite varies from Mg₃Al(OH)₂AlSi₂O₁₀ in high Al samples to Mg_{5.75}Al_{0.5}Si_{3.75}O₁₀(OH)₂.

¶ Epidote is Ca₂Al₂(OH)(SiO₄)₂.

** Muscovite is assumed to be a phengitic sericite, KMg_{0.5}Al₂Si_{3.5}O₁₀(OH)₂.

†† Magnetite calculated as FeO; note that the Fe could equally well have been assigned to the silicates, and if calculated as FeSiO₃, for example, the quartz contents of tektites are reduced to 53, 54, 45, 33, 6, 18, and 26, respectively.

Notes.

Calculations made as follows: all Na goes into albite; all Fe into magnetite; all K into muscovite; all Ti into sphene; all Ca left over from the sphene goes into epidote; all Mg left over from the muscovite goes into chlorite; any Al left over from albite, muscovite, epidote, and chlorite goes into kaolinite; the remaining Si goes into quartz.

82% SiO₂ at 2800°C for 45 min produced a lowering of SiO₂ content to 45% and an enrichment in O¹⁸ of about 1.2‰ (see Figure 6). This change is qualitatively in the same direction as we observe in the various tektite groups, and the rate of change of O¹⁸/O¹⁶ with SiO₂ content is quantitatively very similar to that observed for the HMg group and for the philippinites in the present work. The oxygen isotope variations are thus compatible with the vapor fractionation model of *Walter* [1967], as it is possible that variations in temperature or rate of heating might give the more rapid rate of change of δO¹⁸ with SiO₂ content demanded by the other tektite groups, particularly the bediasites and moldavites. More experiments along these lines would be valuable.

If the mechanism of fractional vaporization is the correct explanation of tektite formation, the tektite parent materials for all three major strewn fields must have an SiO₂ content of about 82 wt % [*Walter*, 1967] and an O¹⁸/O¹⁶ ratio of about 8.9 for the bediasites and Australasian tektites and about 10.7 for the moldavites. There are no even remotely widespread materials on the earth that have the above char-

acteristics and this might be construed as evidence for a lunar origin.

It is very difficult, however, to explain the chemical data on the HMg 'bottle-green' tektites by any process of vapor fractionation [*Cassidy et al.*, 1969, p. 1021], even though this is one of the groups quantitatively most similar to the experimental O¹⁸/O¹⁶ variations observed during vapor fractionation. In addition, *Chapman and Scheiber* [1969] have observed several other chemical trends among the Australasian tektite populations that are very difficult to ascribe to fractional vaporization of tektite material. Certainly, in view of the consistency in the patterns shown by all the tektite groups in Figure 6, whatever is the explanation of the trends it must surely apply to all. Hence, if vapor fractionation is deemed impossible as an explanation of even one of the groups in Figure 6, it must be considered doubtful for all.

Loss of H₂O during relatively low temperature fusion (1700°C) of hydrous silicates also produces an enrichment in O¹⁸ [*Taylor and Epstein*, 1964]. Also, even though vapor fractionation may not be the explanation of the entire chemical variations in tektites, it very

likely has occurred to a limited extent, and we can therefore expect that the original parent material of tektites is probably lower in O¹⁸ than the tektites themselves. The magnitude of this effect is unlikely to be more than about 1‰, but it is in a direction that increases the disparity between tektites and terrestrial soils or sedimentary rocks and reduces the isotopic differences between tektites and igneous rocks.

SUMMARY AND CONCLUSIONS

In summarizing the present study we must report that the oxygen isotope data still provide no definite answer to the elusive question of a lunar versus terrestrial origin for tektites. They do, however, place certain restraints on tektite parent materials, as follows:

1. It is practically inconceivable that the correlations observed in δO^{18} and SiO₂ content of tektites could be due to any process of magmatic differentiation involving simple crystal-melt equilibria.

2. At least a portion of the tektite material is almost certainly of igneous origin (the high-SiO₂ part); in particular, igneous quartz is probably an abundant constituent of the parent materials.

3. The consistent patterns showing an increase in O¹⁸/O¹⁶ ratio with decreasing SiO₂ content in several tektite groups must be due to (a) vapor fractionation of the liquid during impact fusion or (b) mixing between a high-SiO₂ igneous component and a low-SiO₂ low-temperature component formed during weathering at the earth's surface or during hydrothermal alteration at moderate temperatures.

4. Weathered or hydrothermally altered quartz-bearing terrestrial igneous rocks are therefore possible parent materials of tektites. This includes granites, granodiorites, granophyres, and silicic volcanic rocks, as well as possibly some sediments directly derived from igneous rocks, such as certain arkoses and graywackes. The great bulk of soils, sandstones, shales, and other sedimentary and metasedimentary rocks on the earth definitely do not qualify (except in the case of the Ivory Coast tektites).

5. The bediasites and Australasian tektites must have formed from practically identical parent materials. The similarities are even more

astonishing if the HNa/K australites were formed during a separate, older event than the other Australasian tektites, as is suggested by fission-track dating [Fleischer *et al.*, 1969]; these peculiar, low-SiO₂ tektites plot on the same trend as do the other australites (Figure 1). The parent materials of moldavites were apparently about 1‰ richer in O¹⁸ than the other tektite parent materials, but this still represents great uniformity considering the known variation of O¹⁸ in earth surface materials and meteorites. The tektite parent materials are *highly selected* in some fashion; they are far from being random samples of the SiO₂-rich rocks of the earth's surface. This is the only aspect of the oxygen isotope data that could prevent us from wholeheartedly accepting a terrestrial origin of tektites.

6. If tektites originate on the moon, and if the observed chemical and isotopic variations are a property of their parent materials, a low-temperature hydrothermal alteration event of some type must have affected the SiO₂-rich igneous rocks that are presumed to exist on the lunar surface. This conceivably could have happened during the degassing that probably would have accompanied volcanism, or perhaps during an interval when a transient atmosphere was present on the moon. We should note that the only extraterrestrial objects similar in O¹⁸/O¹⁶ ratio to tektites are the types 1 and 2 carbonaceous chondrites, which are the only meteorites that have clearly suffered a low-temperature hydrous alteration.

7. Regardless whether tektites are from the moon or the earth, there must be a sorting mechanism in which the high-SiO₂ portions are at least partially separated from the rest of the tektite parent material, in order to produce the spectrum of compositions observed in tektites. Such sorting processes obviously exist on the earth, and if the moon had a transient atmosphere they might also have existed there. It is perhaps also possible that meteorite bombardment, slumping, and downslope movement may produce sorting effects on the moon. However, the necessity for such a sorting process conceivably could be overcome if (a) the chemical variations in tektite materials are mainly due to quartz content and (b) a rather uniform low-temperature alteration affects the rocks. In such a case, the rocks would all be

enriched in O^{18} , but the high- SiO_2 rocks would be affected least because of the more abundant quartz, which is relatively inert in comparison with the coexisting minerals likely to be found in igneous rocks. In such an example the chemical variations could be a result of any plausible process, and the O^{18}/O^{16} ratios would be merely superimposed later.

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