

Half-life of ^{130}Te double- β decay measured with geologically qualified samples

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Tellurobismuthite (Bi_2Te_3) has been analyzed for Xe isotopes to determine the half-life for double- β decay of ^{130}Te . Excess ^{130}Xe amounts to $(6.18 \pm 0.18) \times 10^7$ atom/g with 47.8 wt. % Te, or $(1.24 \pm 0.04) \times 10^{13}$ for the parent/daughter ratio $^{130}\text{Te}/^{130}\text{Xe}$. With $(9.3 \pm 1.1) \times 10^7$ yr for the Xe retention age of Te mineral, this provides $(7.9 \pm 1.0) \times 10^{20}$ yr for the absolute half-life of ^{130}Te double- β decay. With this and a literature ratio $T_{1/2}(130)/T_{1/2}(128)$ of $(3.52 \pm 0.11) \times 10^{-4}$, we have $(2.2 \pm 0.3) \times 10^{24}$ yr for the absolute half-life of ^{128}Te double- β decay.

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I. INTRODUCTION

Isobars ^{130}Te , ^{130}I , and ^{130}Xe belong to a system of nuclides for which successive single- β decays are energetically forbidden but disintegration through double- β decay (DBD) is possible from ^{130}Te to ^{130}Xe . Since the first report on the possibility of DBD by Goeppert-Mayer [1], many experimental as well as theoretical studies have been done. The first experimental result on the ^{130}Te DBD half-life has been reported by Inghram and Reynolds [2]. Since then, many data have been reported on the ^{130}Te half-life determined by a geochemical method. However, there are long-standing problems of discrepancies in the half-life: Experimentally some workers have reported 2 to 3×10^{21} yr [3–7] and others 7 to 9×10^{20} yr [8–10]. These experimental results are significantly larger by one to two orders of magnitude than theoretical estimates [11]. However, this does not imply that the geochemical method is unreliable because it has given the double- β decay rate for ^{82}Se ([10] and references cited therein) in excellent agreement with that determined by a counting method [12]. In order to solve the discrepancies both in experimental results and between experimental and theoretical half-lives, further works are awaited with samples that are well documented on their occurrences and geological relationships. By the geochemical method, geologically old Te minerals are analyzed for the decay product ^{130}Xe by high sensitivity and high precision mass spectrometry. The amount of parent ^{130}Te is determined from the chemical composition of the sample mineral and the isotopic abundance for ^{130}Te . That the parent to daughter ratio $^{130}\text{Te}/^{130}\text{Xe}$ reported by different workers are in good agreement with one another for samples collected at the same locality (i.e., the same mine) indicates that problems are not in analytical techniques for Xe isotopes and Te contents, but in determination of the time interval in which Te minerals accumulated and retained radiogenic ^{130}Xe produced by *in situ* decay of ^{130}Te [7,13].

In this paper, we will report the half-life of ^{130}Te determined from Xe data for Te and associated mineral separates,

along with the Xe retention age deduced from K-Ar age data for rocks in genetically close relation to the Te samples. We will also give a brief description of field relationships between Te ores and country rocks at the Oya mine, because they are of great importance to decipher the chronological sequence in the formation of mineral deposits and to estimate the Xe retention age for the Te mineral.

II. EXPERIMENT

A. Samples

Samples used in this work are pure tellurobismuthite (Bi_2Te_3) grains of typically 1–2 mm across, separated from a specimen of Te-bearing ores collected at the Oya mine. The ore specimen was carefully crushed to isolate the mineral grains by hand picking. The Te minerals are embedded in quartz veins, in close relationships with native gold, arsenopyrite (FeAsS) and other minerals. The quartz veins, filling fissures, are hosted in the Triassic formation composed mainly of sandstone and slate. The deposits at Oya are characterized by abundant arsenopyrite and the zonal distribution of other minerals. Tellurobismuthite is richer in close association with native gold in the deeper portions of veins. From viewpoints of ore genesis [14–16], it seems that silica-Au-Te-As-bearing hydrothermal fluids were introduced along fissures to deposit the quartz veins therein. These elements were extracted in part from magma and/or in part from country rocks through interactions with water heated by quartz-diorite magma. The fissures were formed both in circumference hoods of a quartzdiorite cupola and in roof rocks surrounding it (i.e., the Triassic sandstone and slate). A quartzdiorite mass crops out near the northern part of the mine and is believed to underlie the ore deposits, because bosses regarded as apophyses of the quartzdiorite outcrop at the mine and some veins cut them. The fissures were initiated by doming up of quartzdiorite magma and formed after at least part of outer rims of the quartzdiorite body had solidified. Field observations give no definite evidence that the Te-bearing veins have experienced thermal (igneous) events postdating the quartzdiorite intrusion, implying that radiogenic ^{130}Xe was retained quantitatively after the mineralization.

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TABLE I. K-Ar age determination for biotite and hornblende from Oya mine.

Sample	Weight (g)	K (wt. %)	$^{40}\text{Ar}/^{38}\text{Ar}$	$^{36}\text{Ar}/^{38}\text{Ar}$	^{38}Ar spike ($10^{-6} \text{ cm}^3 \text{ STP}$)	Rad. $^{40}\text{Ar}^a$ ($10^{-6} \text{ cm}^3 \text{ STP/g}$)	Air cont. (%)	Age ^b (10^6 yr)
Biotite	0.136	4.80	1.7124 ± 0.0045	0.000414 ± 0.000037	1.358 ± 0.027	16.40 ± 0.35	5.62	85.9 ± 1.8
Horn- blende	0.581	0.631	1.3025 ± 0.0034	0.00073 ± 0.000032	1.354 ± 0.027	2.636 ± 0.060	14.3	104.4 ± 3.1

^aFor calculation of radiogenic ^{40}Ar , the following isotopic data were used: $^{36}\text{Ar}/^{38}\text{Ar} = (3 \pm 1) \times 10^{-6}$ and $^{40}\text{Ar}/^{38}\text{Ar} = (8.7 \pm 0.8) \times 10^{-4}$ for spike Ar; $^{36}\text{Ar} = (1.01 \pm 0.24) \times 10^{-10}$ and $^{40}\text{Ar} = (1.5 \pm 0.7) \times 10^{-8} \text{ cm}^3 \text{ STP}$ for blank Ar.

^bThe following decay constants were used for calculation: $\lambda_\beta = 4.692 \times 10^{-10} \text{ yr}$; $\lambda_{\text{ec}} = 5.82 \times 10^{-11} \text{ yr}$ [30].

We also separated quartz and arsenopyrite from the same ore specimen to measure the isotopic ratios of Xe trapped in them, because both minerals are regarded to have crystallized from the same hydrothermal fluids and trapped the same isotopic composition of Xe as the Te mineral did. In particular, quartz will provide constraints on the Xe isotopic ratios in the fluids, because it contains usually innumerable fluid inclusions of micron size. Xenon dissolved in the fluids must have been occluded therein together with the fluids.

For K-Ar dating, biotite and hornblende were separated from the quartzdiorite and porphyrite, respectively. The porphyrite is named for a rock cooled rapidly from the same composition of magma that crystallized to quartzdiorite at slow cooling. A geologist has reported that some porphyrite dikes cut and are cut by quartz veins [14]. This is strong evidence for a contemporary formation of the quartz veins and the porphyrite dikes. The geology of this area will be summarized elsewhere [17] on field relationships between the quartz veins, the porphyrite dikes and the quartzdiorite body and their chronological sequence, together with data on other noble gas isotopes which are not discussed here.

K-Ar dating was performed by analyzing the biotite and hornblende separates for radiogenic ^{40}Ar with conventional techniques using ^{38}Ar spike. The K content was determined for the biotite (4.80 wt. %) and hornblende (0.631 wt. %) separates using conventional techniques by flamephotometry. The result is given in Table I.

B. Mass spectrometry

Xenon isotopes were measured by pulse-counting mass spectrometry [18]. Samples rinsed with ethanol were wrapped in Al foil and mounted in a sample holder for a Ta furnace. A gas-extraction line including the furnace and a gas-purification line are thoroughly degassed. During degassing the lines, the samples were heated overnight approximately at 100°C to desorb atmospheric surface contamination. This temperature was adopted to prevent Xe loss in heating because tellurobismuthite melts at temperature as low as approximately 575°C . Xenon blank (i.e., background) was measured at 850°C for the Te samples and at 1750°C for others.

Typical blanks are $^{132}\text{Xe} = 4.4 \times 10^{-15} \text{ cm}^3 \text{ STP}$ at 850°C for 20 min ($1 \text{ cm}^3 \text{ STP} = 2.688 \times 10^{19}$ atoms), and $^{132}\text{Xe} = 2.7 \times 10^{-14} \text{ cm}^3 \text{ STP}$ at 1750°C for 15 min. Isotopic ratios for blank Xe are atmospheric except for ^{124}Xe and ^{126}Xe , for which excesses due to interference and memory

are found. However, since blank correction is less than 0.5% at ^{132}Xe for the Te samples, the effects of blank correction are negligible except for these isotopes. Sensitivity and mass discrimination of the mass spectrometer were determined by measuring known amounts of atmosphere Xe. Variation in the Xe sensitivity is 3% and the mass discrimination for Xe isotopes is 0.38% per mass unit. The Te content (47.8 wt. %; 1 wt. % Te = 1.598×10^{19} atoms $^{130}\text{Te/g}$) was determined grain by grain for arbitrarily selected 23 grains of the Te minerals with an electron probe microanalyzer. It is essentially constant both in a grain and among grains [17].

III. RESULTS AND DISCUSSION

A. Radiogenic ^{130}Xe

Results of Xe analyses are given in Table II. Uncertainties include statistical errors (2σ) for isotopic measurement, and mass discrimination and blank corrections. There are definite excesses at ^{129}Xe , ^{130}Xe , and ^{131}Xe isotopes, compared to atmospheric Xe. Isotope studies performed so far on Xe in Te minerals have shown that old Te minerals contain products from double- β decays as well as neutron captures on ^{128}Te and ^{130}Te isotopes [3,7]. Hence the ^{130}Xe excess in the present Te samples is attributable to the ^{130}Te double- β decay, and both excesses at ^{129}Xe and ^{131}Xe are products of the neutron captures on ^{128}Te and ^{130}Te , respectively. The isotopic ratios at other isotopes, except at ^{128}Xe from Oya Te-2, are in agreement with the atmospheric ones in experimental errors. This indicates that contribution of fissiogenic Xe is negligible. Other nuclear effects by neutrons, cosmic-ray muons, and other particles [7] are minor and obscured by experimental errors, particularly in ores of young ages such as the present case. We will focus discussion only on the half-life of ^{130}Te in this work, and other nuclear effects including the neutron captures on ^{128}Te and ^{130}Te will not be discussed here. They will be given in [17].

The concentration of radiogenic ^{130}Xe is calculated by subtracting trapped Xe from the measured one. As will be given later, the isotopic ratios of trapped Xe is atmospheric. Hence

$$(^{130}\text{Xe})_{\text{DBD}} = [(^{130}\text{Xe}/^{132}\text{Xe})_S - (^{130}\text{Xe}/^{132}\text{Xe})_A] (^{132}\text{Xe})_S, \quad (1)$$

where subscripts *S* and *A* represent sample and atmosphere,

TABLE II. Concentrations and isotopic abundances ($^{132}\text{Xe}=100$) of Xe in tellurium and associated minerals from the Oya mine. Definite excesses at ^{130}Xe for Te 1 and 2 are attributed to ^{130}Te DBD, while those at ^{129}Xe and ^{131}Xe to neutron captures on ^{128}Te and ^{130}Te , respectively. Excess ^{130}Xe , calculated with Eq. (1), is given in the bottom lines. In calculation of the excess ^{130}Xe , the following isotopic ratios were used for atmospheric Xe: $^{124}\text{Xe}/^{126}\text{Xe}/^{128}\text{Xe}/^{129}\text{Xe}/^{130}\text{Xe}/^{131}\text{Xe}/^{132}\text{Xe}/^{134}\text{Xe}/^{136}\text{Xe}=0.354/0.330/7.136/98.32/15.14/78.90/100/38.79/32.94$.

Sample	Te 1				Te 2		Quartz	Arsenopyrite
Weight (g)	0.210				0.310		0.0382	0.140
T ($^{\circ}\text{C}$)	850	1020	Total	200	850	Total	1750	1750
$^{132}\text{Xe}^a$	6.07	0.209	6.28	2.06	3.00	5.06	30.2	4.54
^{124}Xe	0.37	—	0.37	0.40	0.38	0.39	0.35	0.35
	± 0.10	—	± 0.10	± 0.22	± 0.26	± 0.24	± 0.14	± 0.18
^{126}Xe	0.35	—	0.35	0.36	0.35	0.35	0.34	0.36
	± 0.06	—	± 0.06	± 0.22	± 0.12	± 0.16	± 0.08	± 0.18
^{128}Xe	7.19	7.10	7.19	7.35	7.25	7.29	7.24	7.44
	± 0.26	± 1.00	± 0.26	± 0.34	± 0.18	± 0.24	± 0.24	± 0.18
^{129}Xe	117.9	99.5	117.8	102.4	139.2	124.2	99.4	102.4
	± 1.2	± 7.6	± 1.2	± 3.4	± 2.4	± 2.8	± 1.8	± 2.4
^{130}Xe	52.3	15.1	52.2	15.92	92.5	61.3	15.17	15.79
	± 2.0	± 2.2	± 2.0	± 0.68	± 7.6	± 4.6	± 0.34	± 0.56
^{131}Xe	91.1	78.4	91.1	79.8	104.2	94.3	78.8	79.9
	± 2.0	± 3.4	± 2.0	± 1.8	± 3.0	± 1.8	± 1.5	± 1.3
^{134}Xe	39.0	37.7	39.0	39.1	38.93	38.99	38.66	39.01
	± 1.0	± 3.8	± 1.0	± 1.0	± 0.88	± 0.92	± 0.56	± 0.94
^{136}Xe	32.8	32.6	32.8	32.9	33.05	32.98	32.74	32.51
	± 0.6	± 3.8	± 0.6	± 1.2	± 0.82	± 0.92	± 0.54	± 0.80
$^{130}\text{Xe}^a$	2.26	0.00	2.26	0.016	2.32	2.34	0.01	0.030
Excess	± 0.12	± 0.005	± 0.13	± 0.014	± 0.23	± 0.23	± 0.10	± 0.025

^aConcentrations are given in units of $10^{-12} \text{ cm}^3 \text{ STP/g}$ ($1 \text{ cm}^3 \text{ STP}=2.688 \times 10^{19}$ atoms).

respectively. (^{130}Xe)_{DBD} amounts to 6.07 and 6.29×10^7 atom/g for Oya Te 1 and 2, respectively. The atomic ratio of parent ^{130}Te to daughter (^{130}Xe)_{DBD} is calculated from the above result and the Te contents (i.e., 7.63×10^{20} atoms $^{130}\text{Te/g}$ for Oya Te 1 and 2) to be 1.26 and 1.22×10^{13} for samples Oya Te 1 and 2, respectively.

A possibility to give a short half-life for ^{130}Te has been argued based on possible inheritance of radiogenic ^{130}Xe from precursor materials at the time of mineralization [7]. In order to clarify whether the present samples retain such inherited ^{130}Xe or not, we measured Xe isotopes trapped in quartz and arsenopyrite that crystallized from the same hydrothermal fluids as the Te minerals did. As listed in Table II, the isotopic ratios of Xe from the quartz sample are atmospheric. Xenon from arsenopyrite reveals small isotopic excesses at ^{129}Xe and ^{130}Xe . However, the excess at ^{130}Xe is trifling and attributable to a few bits of Te contaminant (i.e., 0.6 wt. \% Te or $1.8 \text{ mg tellurobismuthite}$ in $140 \text{ mg arsenopyrite}$ sample). Actually some arsenopyrite grains occlude tellurobismuthite crystals of 0.1 mm size [17]. These imply that Xe in the hydrothermal fluids was atmospheric and the Te minerals must have trapped the same isotopic ratios of Xe when they were deposited from the fluids.

Another way assumed to inherit radiogenic ^{130}Xe from precursor materials is that the Te minerals did not crystallize *in situ* in the present veins but were transferred by the fluids

from a separate deposit, mineralized much earlier somewhere in quartzdiorite or its vicinity. This is unlikely because it is impossible for the Te minerals, being low in melting temperature (ca. 575°C), to survive igneous conditions and to retain radiogenic ^{130}Xe . Hence we conclude that no evidence for inheritance of excess ^{130}Xe is found in the Oya Te minerals and that the excess ^{130}Xe determined is radiogenic ^{130}Xe produced by the *in situ* decay of ^{130}Te double- β decay.

B. Xe-retention age

Hydrothermal activity is essentially circulation of hot solutions enriched in various cations and anions that were extracted in part from magma and/or in part from country rocks through interactions with water heated by magma activity. In the case of the Oya mine, the hydrothermal fluids enriched in silica, Au, Te, Fe, and As originated in connection with quartzdiorite magma and filled fissures in the upper Triassic rocks and in the circumference hoods of quartzdiorite cupola to deposit the Te-bearing quartz veins [14–16]. Therefore, the formation age of quartzdiorite could approximate the formation age of the quartz veins. Considering slow cooling relative to quartz veins, probably it would postdate the vein formation. K-Ar dating for biotite separated from the quartzdiorite gives $(8.59 \pm 0.18) \times 10^7 \text{ yr}$ (Table I), in agree-

ment with a literature value [19]. Porphyrite cooled rapidly from the same composition of magma that crystallized to quartzdiorite at slow cooling. If so, the porphyrite should give an age older than the quartzdiorite, because of the fast cooling and high closure temperature for Ar in hornblende, as will be mentioned later. The K-Ar age of $(1.04 \pm 0.03) \times 10^8$ yr for hornblende from the porphyrite predates the age for biotite from the quartzdiorite and this is consistent with the presumed sequence.

One problem to be considered is the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio of initially trapped Ar. Isotopic data on Ar from quartz have revealed that the hydrothermal fluids relevant to the mineralization at Oya contained inherited excess ^{40}Ar [17]. This suggests that both biotite and hornblende might have trapped initial Ar of such high $^{40}\text{Ar}/^{36}\text{Ar}$ at crystallization, because the porphyrite dykes and the quartzdiorite are regarded to be products from the essentially same magma activity responsible for the quartz vein mineralization [14–16]. A possible choice of the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for initial Ar is the measured ratio for quartz: that is, 500.6. Because of accumulation of *in situ* decayed ^{40}Ar , this ratio is likely overestimated. Because the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio for biotite is as high as 5260 (Table I), the initial Ar correction does not affect the age largely. With this initial ratio, the K-Ar age for biotite would be decreased by 4.1% to give 8.24×10^7 yr, whereas the age for porphyrite, for which the $^{40}\text{Ar}/^{36}\text{Ar}$ ratio is 2064, would be decreased by 11.5% to give 9.20×10^7 yr [17]. These ages may be regarded as the lower limit of age for the respective sample.

Another important issue concerning the Xe retention is of difference in closure temperature between Ar and Xe in different minerals. Below the closure temperatures of parent and daughter elements, a geochronometer remains a “closed system” for those elements, and can give a reliable age. Comparing in the same mineral, Xe is higher in closure temperature than Ar. However, in the present case, radiogenic ^{130}Xe resides in tellurobismuthite, while radiogenic ^{40}Ar does in K-rich silicate minerals. Tellurobismuthite has cleavage fracture and low melting temperature. Both cleavage fracture and low melting temperature tend to lower the Xe closure temperature. The closure temperature of radiogenic ^{40}Ar is 250–350 °C for biotite [20–22] and 550 °C for hornblende [20,21,23]. The K-Ar ages for biotite and hornblende, therefore, represent when rocks cooled to 300 °C, on average, and 550 °C, respectively. The closure temperature for hornblende is significantly high and near the melting point of tellurobismuthite. Based on this fact and the field observation of mutual cutting between the quartz veins and the porphyrite dikes [14], we can adopt the age for hornblende separated from the porphyrite dike as an upper limit of the Xe retention age: $t < 1.04 \times 10^8$ yr. However, we are not conclusive for a lower limit of the Xe retention age, because of no quantitative data available on the closure of Xe in tellurobismuthite. As mentioned earlier, the formation age of quartzdiorite probably postdates the formation age of the quartz veins because of the slow cooling of the quartzdiorite body relative to the quartz veins. Based on this assumption, we adopt tentatively the formation age of quartzdiorite determined with biotite as a lower limit: $t > 8.24 \times 10^7$ yr. We believe that this is not a wrong estimate. The mineral deposits at Oya are typical mesothermal deposits [14]. A hint to min-

eralization temperature is acquired from decrepitation temperature of vein quartz. The decrepitation temperature for the Oya quartz veins ranges from 270 to 320 °C [16]. Taking an account of a pressure effect on conversion from the decrepitation temperature to the mineralization one [24], the mineralization temperature of Oya quartz should be higher than these figures. This implies the possibility that the temperature of quartzdiorite that was a heat source of the hydrothermal fluids was still high and biotite in the quartzdiorite was not yet closed for radiogenic ^{40}Ar at the formation of quartz veins. Hence the K-Ar age for biotite could approximate the lower limit of the Xe retention for the Te minerals crystallized in the quartz veins. Consequently, we have $8.2 \times 10^7 < t < 1.04 \times 10^8$, or $(9.3 \pm 1.1) \times 10^7$ yr on average, for the Xe retention age of Te mineral.

C. Half-life for double- β decay of ^{130}Te

The half-life for ^{130}Te is calculated by

$$T_{1/2} = 0.692 \{^{130}\text{Te} / (^{130}\text{Xe})_{\text{DBD}}\} t, \quad (2)$$

where t is the Xe retention age of Te mineral. Substituting $(1.24 \pm 0.04) \times 10^{13}$ for $^{130}\text{Te} / (^{130}\text{Xe})_{\text{DBD}}$ and $(9.3 \pm 1.1) \times 10^7$ yr for the Xe retention age into Eq. (2), we have $(7.9 \pm 1.0) \times 10^{20}$ yr for the half-life of ^{130}Te double- β decay. Uncertainty attached to the half-life was calculated according to error propagation. This is in good agreement with results reported by [8] and [10], but distinctly shorter than those by [3–5] and [6,7,13]. The present result agrees well not only with [8] but also with the ^{130}Te half-life of 8.6×10^{20} yr recalculated from data for another Te sample [25] from the Suwa mine [17,26].

Works reported so far are solid and without doubt in their analytical results [13]. However, there are two problems on determination of Xe retention ages. The first question is of dating methods. Dating with Te mineral itself [3,4] is inadequate because of the extremely low content of parent element, which signifies the possible contribution from inherited daughter and sampling heterogeneity, as discussed by [7,10,13,27]. The common Pb method [6,7,13], as the authors stated, is not based on the *in situ* accumulation of radiogenic daughters, but depends on a model of temporal evolution of average isotopic composition for terrestrial Pb. Therefore, the age represents the time for segregation of magma from its source, but it cannot reflect any geological disturbance to the geochronometer system, such as violation of a “closed system” caused by geological metamorphism, because the common Pb age is dependent on the contents of neither parent nor daughter elements but only on the Pb isotopic ratios, which are insusceptible to partial loss of Pb.

The second question is concerning retentivity of radiogenic ^{130}Xe in Te minerals. The fact that radiogenic ^{130}Xe is mobile from Te minerals even at low temperature, as found in Table II as well as in the literature [4,7,28], suggests that the Te-Xe system is sensitive to geothermal conditions. Although we cannot estimate retentivity for radiogenic ^{130}Xe quantitatively, because of no data available on Xe diffusivity in Te minerals, Xe loss should have resulted in an apparent, long half-life. As well known, Colorado is on a famous orogeny belt, in which there was a wide variety of geological activities such as metamorphism and volcanism postdating

the relevant ore deposits. No comments have been given on geothermal effects on Te minerals from Colorado [3,6,7,28]. For the Oya Te mineral, its occurrence and the geological signatures relating with mineralization have been well documented [14–16].

Compared with theoretical computations [11], the present half-life is more than seven times larger, except for the calculation using the quasiparticle random phase approximation (QRPA) [29]. It is impossible to reduce the Xe retention age largely, if quantitative retention of radiogenic ^{130}Xe is assumed, because the area in which the Oya mine is located is geologically quiescent after termination of the igneous activities in the late Cretaceous. This suggests the existence of some nuclear effect suppressing the double- β decay probability, such as a particle-particle interaction effect in the QRPA [29].

Because of the relatively short Xe-retention age and the small sample size, the present result on the ^{128}Xe isotope is obscured by experimental errors: We have no definite excesses at ^{128}Xe , and therefore no meaningful result on the half-life ratio between ^{130}Te and ^{128}Te , beyond experimental errors for the Oya Te 1 and 2 samples. Since the relative half-life between ^{130}Te and ^{128}Te can be determined independently of the Xe retention age, Te samples of greater

antiquity are more preferable to detection of radiogenic ^{128}Xe produced by the ^{128}Te DBD. The best estimate is $(3.52 \pm 0.11) \times 10^{-4}$ [6,7], which was determined with Te ores of the Precambrian era. With this ratio, the present result of the absolute ^{130}Te half-life provides $(2.2 \pm 0.3) \times 10^{24}$ yr for the absolute half-life of ^{128}Te double- β decay. This agrees with a recommended value of 2×10^{24} by [10].

In conclusion, the tellurium mineral of $(9.3 \pm 1.1) \times 10^7$ yr contains radiogenic $^{130}\text{Xe}_{\text{DBD}}$ amounting to $(6.18 \pm 0.18) \times 10^7$ atom/g with 47.8 wt. % Te, or $(1.24 \pm 0.04) \times 10^{13}$ for parent ^{130}Te /daughter ^{130}Xe . These give $(7.9 \pm 1.0) \times 10^{20}$ yr for the absolute half-life of ^{130}Te double- β decay. With the $T_{1/2}(^{130})/T_{1/2}(^{128})$ ratio of $(3.52 \pm 0.11) \times 10^{-4}$ [6], we have $(2.2 \pm 0.3) \times 10^{24}$ yr for the absolute half-life of ^{128}Te double- β decay.

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- [1] M. Goeppert-Mayer, Phys. Rev. **48**, 512 (1935).
 - [2] M. G. Inghram and J. H. Reynolds, Phys. Rev. **78**, 822 (1950).
 - [3] T. Kirsten, O. A. Schaeffer, E. Norton, and R. W. Stoenner, Phys. Rev. Lett. **20**, 1300 (1968).
 - [4] T. Kirsten, H. Richter, and E. K. Jessberger, Z. Phys. C **16**, 189 (1983).
 - [5] T. Kirsten, E. Heusser, D. Kaether, J. Oehm, E. Pernicka, and H. Richter, in *Nuclear Beta Decays and Neutrino*, edited by T. Kotani, H. Ejiri, and E. Takasugi (World Scientific, Singapore, 1986), p. 81.
 - [6] T. Bernatowicz, J. Brannon, R. Brazle, R. Cowsik, C. Hohenberg, and F. Podosek, Phys. Rev. Lett. **69**, 2341 (1992).
 - [7] T. Bernatowicz, J. Brannon, R. Brazle, R. Cowsik, C. Hohenberg, and F. Podosek, Phys. Rev. C **47**, 806 (1993).
 - [8] N. Takaoka and K. Ogata, Z. Naturforsch. **21a**, 84 (1966).
 - [9] W. J. Lin, O. K. Manuel, G. Cumming, D. Krstic, and R. I. Thorpe, Nucl. Phys. A **481**, 477 (1988).
 - [10] O. K. Manuel, J. Phys. G **17**, 221 (1991).
 - [11] H. V. Klapdor, in *Nuclear Week Process and Nuclear Structure*, edited by M. Morita, H. Ejiri, H. Ohtsubo, and T. Sato (World Scientific, Singapore, 1989), p. 297.
 - [12] S. R. Elliott, A. A. Hahn, and M. K. Moe, Phys. Rev. Lett. **59**, 2020 (1987).
 - [13] F. Podosek *et al.*, in *Noble Gas Geochemistry and Cosmochemistry*, edited by J. Matsuda (Terra Sci. Publ. Co., Tokyo, 1994), p. 89.
 - [14] M. Watanabe, *Sci. Report Tohoku Univ., 3rd Series (Mineral. Petrol. Econom. Geol.)* (Tohoku Univ., Sendai, 1951–1953), Vol. 4, p. 45.
 - [15] K. Yamaoka, Kozan Tisitu (Mining Geol.), Spec. Issue **10**, 151 (1981) (in Japanese with English abstract).
 - [16] H. Imai, Kozan Tisitu (Mining Geol.) **11**, 66 (1961) (in Japanese with English abstract).
 - [17] N. Takaoka, Y. Motomura, and K. Nagao, J. Mass Spectr. Soc. Jpn. **44**, 63 (1996).
 - [18] K. Nagao and Y. Miura, in *Proceedings of the NIPR Symposium on Antarctic Meteorites*, edited by K. Yanai *et al.* (Nat. Inst. Polar Res., Tokyo, 1993), Vol. 6, p. 76.
 - [19] Y. Kawano and Y. Ueda, Ganko (J. Petrol. Mineral. Econom. Geol. Soc. Jpn.) **51**, 127 (1964) (in Japanese with English abstract).
 - [20] G. W. Berger, D. York, and D. J. Dunlop, Nature **277**, 46 (1979).
 - [21] T. M. Harrison, R. L. Armstrong, C. W. Naeser, and J. E. Harakal, Can. J. Earth Sci. **16**, 400 (1979).
 - [22] R. D. Dallmeyer, Am. J. Sci. **278**, 124 (1977).
 - [23] T. M. Harrison, Contr. Mineral. Petrol. **78**, 324 (1981).
 - [24] E. Roedder, in *Fluid Inclusions*, edited by P. Ribbe, Reviews in Mineralogy Vol. 12 (Mineral. Soc. Amer., Washington, D.C., 1984), p. 252.
 - [25] N. Takaoka, PhD. thesis, Osaka University, 1967.
 - [26] T. Takeuchi and M. Nambu, *Sci. Report Tohoku Univ. 3rd Series (Mineral. Petrol. Econom. Geol.)* (Tohoku Univ., Sendai, 1958–1960), Vol. 6, p. 487.
 - [27] O. K. Manuel, in *Nuclear Beta Decays and Neutrino*, edited by T. Kotani, H. Ejiri, and E. Takasugi (World Scientific, Singapore, 1986), p. 71.
 - [28] J. F. Richardson, B. Sinha, and O. K. Manuel, Nucl. Phys. A **453**, 26 (1986).
 - [29] P. Vogel and M. R. Zirnbauer, Phys. Rev. Lett. **57**, 3148 (1986).
 - [30] R. Steiger and E. Jäger, Earth Planet. Sci. Lett. **36**, 359 (1977).