Geochronology of Tellurium Ores and the Double-Beta Decay Lifetime of $^{130}\text{Te}$

F. A. PODOSEK$^{1,3}$, J. C. BRANNON$^{1,3}$, T. J. BERNATOWICZ$^{1,2,3}$, R. BRAZZLE$^{2,3}$, R. GRAUCH$^{4}$, R. COWSIK$^{3,5}$, and C. M. HOHENBERG$^{2,3}$

$^1$Department of Earth and Planetary Sciences, Washington University, St. Louis, MO 63130, U.S.A.
$^2$Department of Physics, Washington University, St. Louis, MO 63130, U.S.A.
$^3$McDonnell Center for the Space Sciences, Washington University, St. Louis, MO 63130, U.S.A.
$^4$U.S. Geological Survey, Denver, CO 80225, U.S.A.
$^5$Tata Institute of Fundamental Research, Bombay 400005, India

**Abstract.** $^{130}\text{Te}$ undergoes double-beta decay to $^{130}\text{Xe}$, which is readily observed in Te ores. Its half-life is of interest in fundamental physical theories of the weak interaction and the nature of the neutrino, and also in potential application as a radiometric geochronometer. Although the occurrence of this decay has long been known and studied and several excellent datasets are available, the value of the half-life of $^{130}\text{Te}$ has remained controversial, with different reviewers recommending values differing by more than a factor of three. In this paper we present relevant new data and review prior data and arguments regarding this issue.

We consider that the difficulty in achieving a definitive geochemical estimate of the half-life of $^{130}\text{Te}$ arises not from any limitations in laboratory analysis but rather from insufficient understanding of the geological character and histories of Te ores. We suggest the possible importance of inherited radiogenic $^{130}\text{Xe}$ in Te ores, and that in order to minimize uncertainties in chronological associations, as well as possible inheritance, evaluation of the half-life of $^{130}\text{Te}$ should be based on ores from volcanogenic massive sulfide deposits. In this light we recommend a "best estimate" half-life of $(2.9 \pm 0.3) \times 10^{21}$ yr, at the high end of the range of other recommendations. We do not consider this value, or any other value recommended in prior literature, to be definitive.

1. Introduction

Our concern in this paper is with two problems that are of interest in rather diverse scientific disciplines but which are closely coupled experimentally: the double-beta decay rate of $^{130}\text{Te}$ and the geochronology of Te ores. Double-beta decay of $^{130}\text{Te}$ is readily observed via excesses of daughter $^{130}\text{Xe}$ in Te ores, but its decay rate is uncertain because it rests on poorly constrained ore ages. Te ores in turn are difficult to date directly because they have very low concentrations of the trace elements used in mainstream radiometric dating; double-beta decay of $^{130}\text{Te}$ might
be a useful geochronometer for these ores if its decay rate were reliably and accurately known.

Double-beta decay is a spontaneous transition of the nuclide \((A, Z)\) to \((A, Z \pm 2)\), accompanied by emission of two electrons (or positrons) and two (or possibly zero) electron antineutrinos (or neutrinos). It is of substantial interest in fundamental particle physics because it has “long been recognized as a sensitive test of lepton number conservation, of the mass and charge conjugation properties of the electron neutrino, and of possible right-handed admixtures in the weak leptonic current” (HAXTON and STEPHENSON, 1984). In the standard model of electroweak interactions it is a second-order process and it is consequently very slow, the slowest natural process that is actually known to occur, with theoretically predicted half-lives of order \(10^{18}\) yr and longer. Since it is so slow, the prospect of observation rests on the absence of any competing decay channels. Double beta decay candidates are thus even-even nuclides \((A, Z)\) for which decay to even-even \((A, Z \pm 2)\) is energetically allowed but for which the intervening odd-odd nuclide \((A, Z \pm 1)\) has a higher energy (because of nuclear pairing effects), so that sequential ordinary (single) beta decays do not occur.

Because of the low decay rates, observations of double-beta decay are very limited. For a half-life of order \(10^{21}\) yr, for example, one gram of parent isotope will produce only a few decays per year, so that laboratory observations are quite difficult. Although the issue has long been of interest, successful laboratory observations of double-beta decay have been reported only in the last decade (see MOE and VOGEL, 1994), and are still restricted to only a handful of cases, notably \(^{82}\text{Se} \rightarrow ^{82}\text{Kr}\) (Elliott et al., 1987; MOE and LOWENTHAL, 1980) and \(^{76}\text{Ge} \rightarrow ^{76}\text{Se}\) (AVIGNONE et al., 1991). The alternative “geochemical method” for observation of double beta decay, based on measurement of excess daughter isotope in ores of the parent element, does not provide any information about decay mechanism but allows much longer integration times. Even so, plausible candidates are limited: Again illustrating for a half-life of \(10^{21}\) yr, in a geological age of \(10^9\) yr only a fraction \(10^{-12}\) of the parent decays, and in order for this to make a significant and detectable excess of the daughter isotope the daughter element must be depleted with respect to the parent by about this order of magnitude. In practice, this restricts candidates to those whose daughters are noble gases, the only elements so strongly depleted in natural materials. It has long been recognized that there are three plausible candidates with noble gas daughters: \(^{130}\text{Te} \rightarrow ^{130}\text{Xe}\), \(^{82}\text{Se} \rightarrow ^{82}\text{Kr}\) and \(^{128}\text{Te} \rightarrow ^{128}\text{Xe}\). The transition \(^{130}\text{Te} \rightarrow ^{130}\text{Xe}\), which is the focus of this paper, was the first instance in which double beta decay was known to actually occur, as evidenced by excesses of \(^{130}\text{Xe}\) in Te ores observed by INGHRAM and REYNOLDS (1949, 1950). (“Excess” in this context, and hereafter in this paper, means excess relative to a background atmospheric component, which can be gauged by other isotopes.) Geochemical evidence has also established the decay \(^{82}\text{Se} \rightarrow ^{82}\text{Kr}\) (KIRSTEN and MÜLLER, 1969; KIRSTEN et al., 1967a, 1967b, 1986).

Although it has been more than four decades since the first observations of double-beta decay, during which a substantial number of noble gas measurements have been made on diverse ores, two major issues concerning the geochemical
observations have been controversial. Opposing views on these issues have been expressed most clearly and consistently by two research groups, in Heidelberg and in Rolla (Missouri), which are responsible for much of the data. One of these issues is whether or not double-beta decay of $^{128}\text{Te}$ occurs; we believe that we have presented compelling evidence for the double-beta decay of $^{128}\text{Te}$ (Bernatowicz et al., 1992, 1993), and this result seems now widely accepted, but this issue is outside the scope of this paper. Here we focus on the second issue, that of the precise value of the halflife of $^{130}\text{Te}$. In passing, however, it should be noted that the implications of $^{128}\text{Te}$ decay for weak interaction and neutrino physics depend in part on the absolute value of the halflife of $^{130}\text{Te}$.

Before we discuss the halflife problem in detail below, some prefatory remarks are in order. Table 1 lists previously published reports of Xe data in Te ores and the corresponding inferred halflives for $^{130}\text{Te}$. It is evident that despite substantial improvements in experimental sophistication and instrumental performance over four decades, the data have not converged. Neither has opinion: Current positions of different reviewers advocate “best estimate” halflives which differ by a factor of three (cf. Kirsten, 1983; Manuel, 1991).

The fault does not lie in the Xe analyses, or at least so it presently appears. Early reports of $^{130}\text{Te}$ double-beta decay encountered some skepticism because the $^{130}\text{Xe}$ excesses were characteristically accompanied by large and poorly understood excesses of $^{129}\text{Xe}$ and $^{131}\text{Xe}$; as long as the $^{129}\text{Xe}$ and $^{131}\text{Xe}$ excesses were not quantitatively understood, whatever process produced them must be suspected of producing the $^{130}\text{Xe}$ as well. Kirsten et al. (1968), however, observed large excesses of $^{130}\text{Xe}$ without the large excesses of $^{129}\text{Xe}$ and $^{131}\text{Xe}$, and Browne and Berman (1973) showed that the $^{129}\text{Xe}$ and $^{131}\text{Xe}$ effects could be understood satisfactorily in terms of resonance capture of epithermal neutrons by $^{128}\text{Te}$ and $^{130}\text{Te}$, respectively. Subsequently there has been essentially no challenge to identification of the large excesses of $^{130}\text{Xe}$ in Te ores as the double-beta decay product of $^{130}\text{Te}$, and all known alternative mechanisms for $^{130}\text{Xe}$ production are quantitatively insignificant. Identification of excess $^{130}\text{Xe}$, in many cases, is not a difficult experimental problem, since the excess $^{130}\text{Xe}$ is commonly a major if not the dominant component contributing to total $^{130}\text{Xe}$ (see Table 1). In these cases, quantification of radiogenic (excess) $^{130}\text{Xe}$* should be possible with errors little if any greater than the basic uncertainty in instrumental sensitivity. Sensitivity errors are commonly quoted at around 10% or better, so that even allowing for some degree of unwarranted optimism in analytical error evaluation we would expect to be able to know the halflife of $^{130}\text{Te}$ within, say, 20% or better. In principle, the uncertainty could be much smaller if statistical convergence of the mean were applicable or if gas abundance quantitation by isotope dilution were more widely applied to this problem than it has been in the past.

The problem enters with the ore ages, specifically gas (Xe) retention ages, which are necessary to calculate the halflife from the measured concentrations of $^{130}\text{Xe}$* (and of Te). The ores are difficult to date directly, since they typically have very low concentrations of the radioactive elements commonly employed in isotopic geochronology. Ages can also be estimated by association of the ores with dateable
Table 1. Previously published results for the halflife of $^{130}\text{Te}$.

<table>
<thead>
<tr>
<th>Locality</th>
<th>Mineral</th>
<th>$^{130}\text{Xe}^{*}$/$^{130}\text{Xe}$</th>
<th>$^{130}\text{Xe}^{*}/\text{Te}$ ccSTP/g×10^{-11}</th>
<th>Age (Ma)</th>
<th>$^{130}\text{Te}$ halflife $\times 10^{21}$ yr</th>
<th>Age method</th>
<th>Ref</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boliden, Sweden</td>
<td>tellurobismuthite (Bi$_2$Te$_3$)</td>
<td>0.64</td>
<td>5.0</td>
<td>1500</td>
<td>1.4</td>
<td>host rock</td>
<td>Inghram &amp; Reynolds (1950)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.78</td>
<td>1.7</td>
<td>126</td>
<td>0.30±0.04</td>
<td>$\text{U-Xe in ore}$</td>
<td>Gerling et al. (1967)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.95</td>
<td>3.6</td>
<td>540</td>
<td>0.62</td>
<td>$\text{K-Ar, U-He in ore}$</td>
<td>Kirsten et al. (1967)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.98</td>
<td>2.6</td>
<td>1560</td>
<td>2.51</td>
<td>$\text{U-Xe in ore}$</td>
<td>Srinivasan et al. (1972a)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.98</td>
<td>6.5(a)</td>
<td>1560</td>
<td>1.01±0.29</td>
<td>$\text{U-Xe in ore}$</td>
<td>Richardson et al. (1986)</td>
</tr>
<tr>
<td>Oya, Japan</td>
<td>tetradymite (Bi$_2$Te$_2$S)</td>
<td>0.68</td>
<td>0.45</td>
<td>91</td>
<td>0.82±0.06</td>
<td>host rock</td>
<td>Takaoka and Ogata (1966)</td>
</tr>
<tr>
<td>Good Hope, Colorado</td>
<td>native Te</td>
<td>0.98</td>
<td>2.4</td>
<td>1310</td>
<td>2.19±0.70</td>
<td>$\text{K-Ar in ore}$</td>
<td>Kirsten et al. (1968)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.96</td>
<td>2.1</td>
<td>1310</td>
<td>2.60±0.28</td>
<td>$\text{K-Ar in ore}$</td>
<td>Kirsten et al. (1983)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.98</td>
<td>2.8(a)</td>
<td>1310</td>
<td>2.00±0.25</td>
<td>$\text{K-Ar in ore}$</td>
<td>Richardson et al. (1986)</td>
</tr>
<tr>
<td>Kirkland Lake, Ontario</td>
<td>altaite (PbTe)</td>
<td>0.87</td>
<td>3.4</td>
<td>1826</td>
<td>2.03±0.30</td>
<td></td>
<td>Alexander et al. (1969)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.47</td>
<td>4.1(a)</td>
<td>1826</td>
<td>1.88±0.30</td>
<td></td>
<td>Richardson et al. (1986)</td>
</tr>
<tr>
<td>Kalgoorlie, Australia</td>
<td>mixed tellurides</td>
<td>0.73</td>
<td>3.6</td>
<td>2460</td>
<td>2.83±0.30</td>
<td></td>
<td>Srinivasan et al. (1972b)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.98</td>
<td>10.8(a)</td>
<td>2460</td>
<td>1.0</td>
<td></td>
<td>Hennecke et al. (1975)</td>
</tr>
<tr>
<td>Mattagami Lake, Quebec</td>
<td>krennerite ((Au,Ag)/Te$_2$)</td>
<td>0.99</td>
<td>14.0</td>
<td>2750</td>
<td>0.72(b)</td>
<td></td>
<td>Lee et al. (1991)</td>
</tr>
<tr>
<td>Kuusamo, Finland</td>
<td>altaite (NiTeSe)</td>
<td>0.96</td>
<td>5.1(a)</td>
<td>&gt;0.99</td>
<td>2.27±0.08</td>
<td></td>
<td>Richardson et al. (1986)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.98</td>
<td>5.8</td>
<td>2725</td>
<td>1.93(b)</td>
<td></td>
<td>Lin et al. (1988b)</td>
</tr>
<tr>
<td>Robb-Montbray, Quebec</td>
<td>melonite (NiTe$_2$)</td>
<td>0.93</td>
<td>6.7</td>
<td>2000</td>
<td>1.25±0.08</td>
<td>host rock</td>
<td>Lin et al. (1986)</td>
</tr>
<tr>
<td>&quot;</td>
<td></td>
<td>0.98</td>
<td>6.2</td>
<td>1095</td>
<td>0.75±0.03</td>
<td>associated mineral</td>
<td>Lin et al. (1988a)</td>
</tr>
</tbody>
</table>

(a) $^{130}\text{Xe}^{*}/\text{Te}$ calculated via $^{130}\text{Te}$ (n, γβ) $^{131}\text{Xe}^{*}$ in reactor irradiation, rather than by direct division of $^{130}\text{Xe}^{*}$ by Te concentrations.

(b) Nominal halflife not stated by authors; calculated from tabulated data.
rocks and minerals, but only with an added risk that the association may not be valid.

The view of the Heidelberg group, as expressed by KIRSTEN (1983) for example, is that the halflife of $^{130}\text{Te}$ should be based on samples in which there is a dominant excess of $^{130}\text{Xe}$ and for which there are "internal gas retention ages," i.e. those determined for the ore itself. In practice this restricts the field to the U-Xe age of Boliden tellurobismuthite (KIRSTEN et al., 1967a) and the K-Ar age for Good Hope native Te (KIRSTEN et al., 1968). On the basis of data for these ores, KIRSTEN (1983) recommends a "best value" halflife of $(2.55 \pm 0.20) \times 10^{21}$ yr. There are also several other ore analyses (Table 1), with ages based on association with other rocks and minerals, which yield halflife estimates in the range $(2-3) \times 10^{21}$ yr, close enough to Kirsten's recommended value that the differences might possibly be understood in terms of analytical errors. There are also several analyses (Table 1) which indicate lower halflives, however, around $1 \times 10^{21}$ yr and lower. These have to be dismissed as "age underestimates" (KIRSTEN et al., 1986).

The Rolla group, particularly beginning with RICHARDSON et al. (1986), has advocated a much shorter halflife, $\leq 1 \times 10^{21}$ yr. They note that the Xe retention age of a given sample must be less than or equal to its mineralization age, which in turn must be less than or equal to its host rock age. They therefore argue that if different samples lead to different apparent halflives, the lowest apparent halflife based on a well-constrained upper limit mineralization or host rock age is an upper limit to the true halflife. By this logic, any analysis which leads to an apparent halflife greater than this bound must be in error by assuming too great an age, i.e. that radiogenic $^{130}\text{Xe}^*$ has been lost due to metamorphism, deformation, etc. RICHARDSON et al. (1986) presented an upper bound based on their own analysis of Boliden tellurobismuthite (Table 1) combined with the U-Xe age of KIRSTEN et al. (1967b). Further reinterpretation (MANUEL, 1986) and additional data have reduced this limit somewhat, to a "current best estimate" halflife of $0.8 \times 10^{21}$ yr in the most recent review by MANUEL (1991).

As a parenthetical note, it should be appreciated that prior geochemical studies of Xe in Te ores have been motivated primarily by interest in the fundamental physics of double beta decay, and dissemination of the results via publication and conference presentation has been directed mostly to the physics audience. From the theoretical physics viewpoint, one of the major issues, that of double beta decay of $^{128}\text{Te}$, is now resolved, and even a factor of three uncertainty in the absolute halflife, as long as the true halflife can be confidently taken to lie within that range, is something less than a first-order problem. From a geochemistry or economic geology viewpoint, however, this level of uncertainty is highly unsatisfactory, and consists of a particularly vexing situation in view of laboratory facilities which should afford a much better result. Prior Te-Xe work has only marginally explored geological ramifications, and to first order the economic geology community seems mostly unaware of this work. This is regrettable both because increased geology-based attention might help resolve this problem and because the potential utility of such work to geological studies has been largely unexploited.

We have performed new analyses of a suite of Te ores, using improved noble gas analytical techniques and additional measurements by thermal ionization mass
spectrometry. Aspects most pertinent to basic physical theory, especially concerning evidence for double-beta decay of $^{128}$Te, are discussed elsewhere (BERNATOWICZ et al., 1992, 1993); here we concentrate on the closely linked problems of Te ore ages and the halflife of $^{130}$Te. Our dataset includes Pb isotopic analyses which provide chronological constraints through models for the evolution of common Pb, a source of relevant information which has been very much underexploited in consideration of this problem.

2. Samples and Experimental Procedures

Samples of native Te from three mines in Colorado, USA, were obtained from P. Dunn, USNMH, Smithsonian Institution: the American Mine (#R177), Boulder County; the Vulcan Mine (#C88) and the Good Hope Mine (#85138), Gunnison County. The Good Hope sample is from the same locality as the samples studied by KIRSTEN et al. (1968, 1983).

Samples of altaite (PbTe) from Mattagami Lake, Quebec and krennerite ((Au, Ag)Te$_2$) from Kalgoorlie, Australia, were obtained from O. K. Manuel. These ores were previously analyzed by LEE et al. (1991), who report that the Mattagami altaite “was hand-picked from a telluride-rich seam that is conformable with schistosity in highly altered rock stratigraphically beneath the Mattagami Lake volcanogenic massive sulfide deposit” and that the Kalgoorlie krennerite “was hand-picked from veins in a siliceous lode, collected at level #10, Lake View Shaft, Golden Mile Dolerite.”

Samples from the Cripple Creek District in Colorado were krennerite from the Independence Mine (USNMH #92677) and calaverite (AuTe$_2$) from the 2800 foot level of the Bobtail vein in the Ajax Mine (Colorado School of Mines #9516). These two samples are from two different vein systems separated by about 1500 feet. The mineralogical identity of each was determined by x-ray diffraction.

Te concentrations in all samples were measured by energy-dispersive x-ray spectrometry (EDS). The three native Te samples were all at least 99.9% Te (i.e. no other peaks were detected). We observed 39.0% Te in the Mattagami altaite, 60.0% Te in the Kalgoorlie krennerite, 58.8% Te in the Cripple Creek krennerite and 57.4% Te in the Cripple Creek calaverite; the EDS analyses are believed accurate within 2% or better. The Te concentrations of these samples are in satisfactory agreement with published values for specimens from these locations (SINDEEVA, 1964).

Xe was analyzed by stepwise heating of samples crushed in vacuum to enhance the separation of trapped and in situ gas; procedures for the noble gas analysis are described in detail by BERNATOWICZ et al. (1993). For thermal ionization analysis, fragments adjacent to the samples used for noble gas analysis were briefly washed in 1 N HCl and 1 N HNO$_3$, crushed, washed again in 1 N HCl and 1 N HNO$_3$, and dissolved in aqua regia. The American and Good Hope samples were split and analyzed in replicate (samples A and B). The crushed American, Good Hope and Vulcan samples were first washed in H$_2$O, which was analyzed separately from the dissolution fractions in an (unsuccessful) attempt to obtain chronological information via the Rb-Sr method (cf. BRANNON et al., 1992). Selected elements (K, Ca, Rb,
Sr, Pb, U) were separated by ion exchange chromatography. Concentrations were determined by isotope dilution. Thermal ionization isotopic analyses were performed on a VG354 mass spectrometer using techniques previously described, e.g., by Brannon et al. (1991) and Pier et al. (1989).

3. Results

Raw data for Xe in stepwise heating of the Te ores analyzed in this study are presented by Bernatowicz et al. (1993), who also describe resolution of the measured composite Xe into contributory components: ambient or trapped Xe of atmospheric composition; heavy isotopes (\(^{131}\)Xe, \(^{132}\)Xe, \(^{134}\)Xe, \(^{136}\)Xe) from spontaneous fission of \(^{238}\)U; excesses of \(^{129}\)Xe and \(^{131}\)Xe attributed to \((n,\gamma)\) reactions on \(^{128}\)Te and \(^{130}\)Te, respectively; excess \(^{130}\)Xe attributed to double-beta decay of \(^{130}\)Te; excess \(^{128}\)Xe attributed primarily to double-beta decay of \(^{128}\)Te; excess \(^{126}\)Xe attributed to cosmic-ray secondary muon reactions; and other potential nuclear reaction components evaluated as being quantitatively insignificant. In the present context the quantities of interest are (excess) radiogenic \(^{130}\)Xe\(^*\) (derived from double-beta decay of \(^{130}\)Te) and (excess) fission \(^{136}\)Xe\(^*\) (derived from spontaneous fission of \(^{238}\)U); these are listed in Table 2.

In general terms this deconvolution could be considered a delicate affair, in that various aspects depend sensitively on experimental errors or involve not very well constrained parameters. It must be stressed, however, that this is not the case for the quantities of primary interest here. Evaluation of the fission Xe contribution is based on ratios in the \(^{132}\)Xe, \(^{134}\)Xe, \(^{136}\)Xe system. In general the fission components are relatively small (compared to the trapped components), but there is no ambiguity in that the trapped and fission components are the only known contributors to this three-isotope system and the end-member compositions are well known; uncertainties in the fission component are thus adequately represented by formal error propagation based on analytical errors. For the \(^{130}\)Xe\(^*\) ascribed to double-beta decay of \(^{130}\)Te there is no subtlety at all: No suggestions of quantitatively competitive sources of excess \(^{130}\)Xe, other than double-beta decay of \(^{130}\)Te, have been advanced, and even for the

<table>
<thead>
<tr>
<th>Sample</th>
<th>Mineral</th>
<th>(^{136})Xe(^*) (ccSTP/g×10(^{-14}))</th>
<th>(^{130})Xe(^*) (ccSTP/g×10(^{-11}))</th>
<th>(^{130})Xe(^*)/(^{130})Te</th>
<th>Age(^{(b)}) (Ma)</th>
<th>(^{130})Te half-life(^{(c)}) (×10(^{21})yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>American Mine</td>
<td>native Te</td>
<td>0.82±0.09</td>
<td>0.993</td>
<td>2.12</td>
<td>1660</td>
<td>(3.22)</td>
</tr>
<tr>
<td>Good Hope Mine</td>
<td>native Te</td>
<td>0.18±0.09</td>
<td>0.994</td>
<td>2.34</td>
<td>1700</td>
<td>2.99</td>
</tr>
<tr>
<td>Vulcan Mine</td>
<td>native Te</td>
<td>0.49±0.07</td>
<td>0.997</td>
<td>2.38</td>
<td>1700</td>
<td>2.94</td>
</tr>
<tr>
<td>Mattagami</td>
<td>altaite</td>
<td>0.38±0.12</td>
<td>0.994</td>
<td>4.26</td>
<td>2720</td>
<td>2.63</td>
</tr>
<tr>
<td>Kalgoorlie</td>
<td>krennerite</td>
<td>&lt;0.11</td>
<td>0.998</td>
<td>9.37</td>
<td>(743)</td>
<td>(0.33)</td>
</tr>
<tr>
<td>Cripple Creek</td>
<td>krennerite</td>
<td>2.54±0.15</td>
<td>0.81</td>
<td>0.123</td>
<td>30</td>
<td>(1.00)</td>
</tr>
<tr>
<td>Cripple Creek</td>
<td>calaverite</td>
<td>1.08±0.52</td>
<td>0.77</td>
<td>0.127</td>
<td>30</td>
<td>(0.97)</td>
</tr>
</tbody>
</table>

(a) Detailed presentation and discussion of Xe data by Bernatowicz et al. (1993).
(b) Ages are "best estimates"; see text for discussion.
(c) Values in parentheses not used for overall "best estimate" half-life evaluation; see text for discussion.
young Cripple Creek samples the excess $^{130}\text{Xe}^*$ so dominates total measured $^{130}\text{Xe}$ (Table 2) that uncertainty in inferred concentration of $^{130}\text{Xe}^*$ is essentially the uncertainty in instrumental sensitivity (5%; Bernatowicz et al., 1993).

It is instructive to make to make comparisons, where possible, between our results for $^{130}\text{Xe}/\text{Te}$ (Table 2) and those of earlier studies (Table 1). Our results for Good Hope and Vulcan native Te are in good agreement with the Kirsten et al. (1968, 1983) data for Good Hope. The result of Richardson et al. (1986) is higher by about 17%; some of this difference is likely to reflect their use of reactor irradiation calibration and stated 90% Te concentration in a sample described as “essentially pure tellurium,” and direct division of their $^{130}\text{Xe}$ by sample weight (assuming 100% Te) results in a value 9% lower. Our value for $^{130}\text{Xe}/\text{Te}$ in Kalgoorlie krennerite, the same sample studied by Lee et al. (1991), is 33% lower than theirs; this difference arises primarily because our value for Te concentration in this sample is 60%, which is 37% higher than their reported atomic absorption value of 43.8% (the stoichiometric concentration of Te in pure $\text{AuTe}_2$ is 56.3%, and will be higher if Ag substitutes for Au). Our value for Mattagami altaite is 16–28% lower than the various Rolla measurements (Table 1); part of this difference arises from our Te concentration of 39%, which is 10% higher than the 35.5% atomic absorption value of Lee et al. (1991) and Lin et al. (1988b) (Te in stoichiometric altaite is 38%). Overall, it appears that the differences between our results and prior results arises less from Xe measurement than from Te measurement.

Thermal ionization data are presented in Table 3. As noted, the attempt to obtain chronological information from the Rb-Sr system was unsuccessful, essentially because of the very low concentrations of Rb and Sr. Chronological information is also potentially available through comparison of the concentrations of U and of fission Xe from spontaneous fission of $^{238}\text{U}$. For the Good Hope sample, the U-Xe age, within rather large error limits, is consistent with its geological age (see below). The Cripple Creek krennerite, with its much higher U concentration, yields a U-Xe age of 23 ± 2 Ma, roughly consistent with an inferred geological age of 30 Ma (Thompson et al., 1985). The American and Vulcan mine samples, however, have nominal U-Xe ages in excess of the age of the earth. It must be concluded that at least these samples exhibit excess or “inherited” fission Xe, i.e. that fission Xe, presumably extracted from the country rock, was present in the fluids from which these ores formed. Such an effect is well known for the radiogenic gases $^4\text{He}$ and $^{40}\text{Ar}$ (see e.g., Ozima and Podosek, 1983) and it is not implausible in the present case, especially for materials which have such low parent (U) concentrations in comparison with most crustal rocks.

Common Pb isotopic compositions from Table 3 are displayed in Fig. 1. These provide another possible approach to ore geochronology through comparison with a global Pb evolution model such as that of Stacey and Kramers (1975). Formal common Pb ages, calculated from the Stacey-Kramers model with second-stage $\mu$ ($^{238}\text{U}/^{204}\text{Pb}$) as a free parameter, are included in Table 3. It is not realistic to assign formal statistical errors to such ages, since the uncertainties arise not from analytical errors but from how well the histories of the samples in question conform to the assumptions of the model. For samples which derive their Pb from the mantle or from
Table 3. Thermal ionization data for tellurium ores.

<table>
<thead>
<tr>
<th>Sample(a)</th>
<th>Mineral</th>
<th>[K] (ppm)</th>
<th>[Ca] (ppm)</th>
<th>[Rb] (ppb)</th>
<th>[Sr] (ppb)</th>
<th>[U] (ppb)</th>
<th>[Pb] (ppm)</th>
<th>206Pb (b)</th>
<th>207Pb (b)</th>
<th>208Pb (b)</th>
<th>204Pb</th>
<th>Model age(c)</th>
<th>Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td>American Mine A</td>
<td>native Te</td>
<td>0.14</td>
<td>0.35</td>
<td>0.14</td>
<td>1.7</td>
<td>0.8</td>
<td>0.5</td>
<td>15.58</td>
<td>15.20</td>
<td>34.99</td>
<td></td>
<td>1648</td>
<td></td>
</tr>
<tr>
<td>American Mine B</td>
<td>native Te</td>
<td>0.40</td>
<td>0.36</td>
<td>0.34</td>
<td>3.4</td>
<td>1.0</td>
<td>0.6</td>
<td>15.60</td>
<td>15.22</td>
<td>35.03</td>
<td></td>
<td>1675</td>
<td></td>
</tr>
<tr>
<td>Good Hope Mine A</td>
<td>native Te</td>
<td>2.6</td>
<td>3.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>15.63</td>
<td>15.20</td>
<td>34.94</td>
<td></td>
<td>1603</td>
<td></td>
</tr>
<tr>
<td>Good Hope Mine B</td>
<td>native Te</td>
<td>0.20</td>
<td>≤0.04</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3.3</td>
<td>2.3</td>
<td>15.67</td>
<td>15.22</td>
<td>35.04</td>
<td>1596</td>
</tr>
<tr>
<td>Vulcan Mine</td>
<td>native Te</td>
<td>≤0.04</td>
<td>≤0.03</td>
<td>0.36</td>
<td>3.1</td>
<td>0.8</td>
<td>0.8</td>
<td>15.70</td>
<td>15.24</td>
<td>35.13</td>
<td></td>
<td>1612</td>
<td></td>
</tr>
<tr>
<td>Mattagami</td>
<td>altaite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(61%)</td>
<td>13.11</td>
<td>14.31</td>
<td>32.94</td>
<td>2657</td>
<td></td>
</tr>
<tr>
<td>Kalgoorlie</td>
<td>krennerite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>16.75</td>
<td>15.25</td>
<td>36.14</td>
<td></td>
<td>743</td>
<td></td>
</tr>
<tr>
<td>Cripple Creek</td>
<td>krennerite</td>
<td>2200</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>18.16</td>
<td>15.33</td>
<td>38.81</td>
<td></td>
<td>210</td>
<td></td>
</tr>
</tbody>
</table>

(a) A and B designate separate analyses of the American and Good Hope samples.
(b) Two-sigma errors 2.0 permil in 206/204 and 207/204, 2.4 permil in 208/204; error correlation coefficients 0.908 for 207/204-206/204 and 0.928 for 208/204-206/204. Data for 206/204 and 207/204 include (very minor) corrections for U decay.
(c) Calculated from model of Stacey and Kramers (1975) with second-stage $\mu$ (238U/204Pb) as a free parameter.
(d) Pb concentration determined by energy-dispersive x-ray analysis.
Fig. 1. Common Pb isotopic evolution diagram for Te ores. Panel (a) shows data from Table 3 and second-stage evolution and isochrons for the two-stage model of Stacey and Kramers (1975). Panel (b) shows detail for Colorado native Te samples (two-sigma error ellipses) and Stacey-Kramers second stage isochrons and evolution trajectories for different values of $\mu^{238}$ (U/204Pb).

representative crustal rocks, common Pb ages are typically considered reasonably reliable, but even so there is no evident consensus about exactly how reliable. A reasonable estimate is perhaps one to two hundred Ma; this makes the method useless for discriminating ages for relatively young samples (e.g., Cripple Creek) but still quite useful on a $10^9$ yr timescale.

4. Geological Considerations

The Vulcan and Good Hope mines, in west-central Colorado, are developed through two shafts, separated by about 550 feet, into the same mineral deposit. As described by Hartley (1976, 1983), this is a zoned deposit with a basal stratabound massive Zn-Cu-Fe-rich sulfide lens overlain by a telluride-bearing precious-metal vein system, which in turn interingers with and is partially overlain by a selenosulfur lens. This is interpreted to be a primary exhalative massive sulfide deposit, with hot spring activity which redistributed some of the sulfur and metals to form the
precious-metal veins and the selenosulfur lens at about the same time as the formation of the primary deposit and of the host rock. This deposit is hosted by the Dubois greenstone, a metamorphosed suite of felsic tuffs and mafic flows, probably formed no earlier than 1790 Ma, the age of stratigraphically equivalent rocks (E. deWitt, private communication), but metamorphosed before 1700 ± 70 Ma, the age of the oldest intrusive unit in the area (HANSEN and PETERMAN, 1968).

KIRSTEN et al. (1968) report a K-Ar age of 1310 Ma for Good Hope native Te, but as discussed below we do not consider this result definitive. As far as we know, the only other chronological data for the ore minerals are our common Pb data (Table 3). With due allowance for uncertainty in the host rock age and systematic uncertainty in common Pb ages, there is no reason to challenge the interpretation that the ores formed at essentially the same time as their host rocks. We thus adopt 1700 Ma as a “best estimate” age for this deposit.

The American Mine is in the Sunshine District of the Boulder Telluride Belt in east-central Colorado, which has been extensively studied and described (e.g., see GELLER et al., 1987; LOVEERING and GODDARD, 1950). Ores occur in epithermal veins generally thought to be related to Tertiary intrusions. Deposits are hosted by the Boulder Creek Granodiorite, of approximate age 1700 Ma (cf. PREMO and VAN
Paragenetic studies of hydrothermal minerals in two nearby deposits (B. A. Geller, private communication) indicate that native Te formed after adularia which has been dated at 55 Ma (Walter and Geller, 1992).

As at Vulcan-Good Hope, our Pb results for the American Mine sample are consistent with the host rock age, but they certainly conflict with the inferred ≤55 Ma age for native Te. The Xe results also conflict: If the observed $^{130}$Xe were supposed to have been generated in only 55 Ma, the half-life of $^{130}$Te would have to be $0.11 \times 10^{21}$, far shorter than any value advocated and completely incompatible with the $^{130}$Te-$^{82}$Se relationship discussed below. The similarity of both the Pb and Xe results for the American Mine sample to those for the Vulcan and Good Hope samples suggests that much older and previously unrecognized tellurides are present in the Boulder Telluride Belt, or that our American Mine sample is misidentified. In any case, this sample should not be used for $^{130}$Te half-life evaluation.

Ores in the Cripple Creek District, approximately 20 miles west of Colorado Springs, Colorado, occur in veins hosted by a Tertiary diatreme-intrusive complex; some of the veins extend into surrounding Precambrian metamorphites and igneous rocks. The geology of the area, and an extensive literature describing the district, are reviewed by Thompson et al. (1985). The igneous rocks of the diatreme-intrusive complex were emplaced in several discrete pulses between 32.5 and 28.7 Ma (Kelly et al., 1993); $^{40}$Ar-$^{39}$Ar ages for alteration minerals and veins thought to be associated with telluride-bearing veins indicate that the age of mineralization is between 31.3 and 28.2 Ma (Kelly et al., 1993; K. D. Kelley, private communication). With reasonable allowance for systematic uncertainties, our Pb model age for Cripple Creek (Table 3) is consistent with such a low age. Our “best estimate” age for the Cripple Creek deposit is 30 Ma.

The Mattagami Lake ore deposit is in the northern volcanic zone of the large Archean Abitibi Greenstone Belt in Quebec and Ontario, Canada. It is a volcanogenic massive sulfide deposit thought to have formed in a seafloor depression on top of rhyolitic volcanic rocks and considered to be coeval with its host rocks (Costa et al., 1983). We take this age to be 2720 Ma, as reviewed by Luden and Hubert (1986); this is consistent with the age adopted by Lee et al. (1991) (citing private communication from J. K. Mortensen). With reasonable allowance for systematic uncertainty, our Pb data (Table 3) are quite consistent with this age. We thus adopt a “best estimate” age of 2720 Ma for the Mattagami Lake deposit.

The Kirkland Lake district (Macassa Gold Mine), samples from which were studied by Alexander et al. (1969) (see Table 1), is in the southern portion of the Abitibi Greenstone Belt, about 150 miles southwest of the Mattagami Lake deposit. The intrusive igneous host rocks are about 2690 Ma, but the tellurides (and gold-bearing minerals) occur in veins which evidently formed over a long period of time in response to a variety of intrusive events, with gold-bearing veins dated down to 2400 Ma (Hattori, 1993; Tefsaya, 1992). Pb isotopic compositions in altaite from nearby mines show significant variations between mines and within individual samples, indicating multiple sources for metals in the veins (Hattori, 1993). For their altaite samples Alexander et al. (1969) adopt a mineralization age of 1826 Ma, citing Reesman et al. (1968) and R. H. Reesman (private communication).
LIN et al. (1988b) analyzed melonite from the Robb-Montbray property, which is also in the Abitibi Greenstone Belt. The tellurides are thought to occur in vein systems similar to other gold-bearing veins in this region, but the age of mineralization is not well known. LIN et al. (1988b) suggest that the age is probably between 2720 Ma (host rock age) and 2404 Ma (unreferenced common Pb age).

N. Takaoka (private communication of unpublished results) has analyzed tellurides from the Suwa mine, obtaining an apparent halflife equivalent to that obtained by TAKAOKA and OGATA (1966) for the Oya mine (Table 1). These mines are in northeastern Honshu (Japan); tellurides occur in hydrothermal quartz veins believed to be associated with granodioritic intrusives (TAKEUCHI and NAMBU, 1960; WATANABE, 1923, 1953) of ages 90–120 Ma (KAWANO and VEDA, 1967; N. Takaoka, private communication).

The Kalgoorlie mine is in the Archean Norseman-Wiluna greenstone belt within the Yilgarn Block of western Australia. Gold and telluride mineralization occur in veins commonly attributed to devolatilization of underlying greenstones. Prior Te-Xe studies by HENNECKE et al. (1975) and SRINIVASAN et al. (1972b) have cited an estimated mineralization age of 2460 Ma due to TUREK (1966), but more recent data (e.g., COMPSTON et al., 1986; GROVES, 1993; MCNAUGHTON and DAHL, 1988) indicate that the volcanic and sedimentary host rocks in this area were deposited and metamorphosed between 2715 and 2630 Ma. Major gold-bearing veins in this greenstone belt are thought to be contemporaneous with metamorphism at about 2630 Ma (GROVES, 1993). This age range is also well supported by common Pb model ages in this region, including some for altaite from nearby mines (BROWNING et al., 1987).

Our Pb analysis of Kalgoorlie krennerite is much more radiogenic, indicating a much shorter model age (Table 3). This is certainly unusual, but not necessarily unreasonable. While major (gold) mineralization in this area was evidently emplaced during the ductile phase of shear zone evolution (e.g., MUELLER et al., 1988), there are at least some telluride ores which are undeformed and therefore stratigraphically later (PHILLIPS, 1986), indicating fluid evolution or remobilization of earlier mineralization. BROWNING et al. (1987) also report (rare) occurrences of common Pb in this region which are even more radiogenic than our analysis (Table 3), indicating minor mineralization at later times, possibly as recently as ca. 400 Ma. Evidently the time of telluride mineralization around Kalgoorlie is uncertain and widely variable. In the absence of detailed independent chronological control it would seem that Kalgoorlie tellurides are not useful for $^{130}$Te halflife estimation. It is worth noting, however, that if an apparent common Pb age is seriously in error it is not implausible that the age appears too great but it is less plausible that the age appears too small, since it is easy to understand mineralizing fluids acquiring Pb from an older source with low U/Pb but not easy to understand acquisition of more modern-appearing Pb.

If we stipulate that this sample of Kalgoorlie krennerite is actually not much older than its nominal common Pb age (Table 3), then pursuit of the Rolla argument advanced by RICHARDSON et al. (1986) et seq., leads to a $^{130}$Te halflife (Table 2) at least twofold lower than any recent investigations have recommended.

The Boliden mine, samples of which have been studied in several investigations
(see Table 1), is in the Skellefte district in northern Sweden. This deposit, hosted by metamorphosed volcanic rocks and marine sediments, is interpreted as an exhalative massive sulfide deposit, formed at essentially the same time as its host rocks (RICKARD and ZWEIFEL, 1975). The age is rather uncertain, however. The age of some of the volcanics is about 1950 Ma, but it is not clear whether this is a primary or a metamorphic age. The deposits are thus at least 1950 Ma, but the area has also been affected by younger intrusive events and tectonism within the range 1890 to 1750 Ma (RICKARD and ZWEIFEL, 1975; VIVALLO, 1987). Prior estimates of the $^{130}$Te half life have used lesser ages (Table 1), particularly the 1560 Ma U-Xe age of KIRSTEN et al. (1967a). As discussed below, we consider that such ages may be unreliable for technical reasons.

For subsequent discussion it is important to distinguish different modes of formation for the various telluride-bearing ore deposits whose samples have been used to address the issue of $^{130}$Te half life determination. All are hydrothermal deposits for which the heat source is basically igneous activity. One class of particular relevance in this context is the so-called volcanogenic or exhalative massive sulfide (VMS) deposit, which is generally believed to form in a marine environment (oceanic spreading center or back arc basin). VMS deposits are generally believed to form at essentially the same time as their host igneous and sedimentary rocks, which helps constrain the time of mineralization, and it is also generally believed that the source of the ore constituents is the mantle, which is a relatively uniform reservoir (in contrast to continental crust). The latter feature is presumably responsible for the well-known tendency of VMS common Pb isotopic compositions to agree relatively well with simple global models for Pb evolution (e.g., see DOE and STACEY, 1974).

The contrast is with cases in which hydrothermal fluids deposit minerals in veins in host rock which may be significantly older than the ore deposits. Independent constraints for the time of mineralization can be more difficult for such deposits, particularly if there are multiple generations of hydrothermal activity. Also, the source of the ore constituents may be older and more highly differentiated continental crustal materials. This feature evidently accounts for more anomalous and more variable common Pb isotopic compositions (e.g., BROWNING et al., 1987; DOE and STACEY, 1974; HATTORI, 1993); in present context it may also be relevant that the hydrothermal fluids have opportunity to come into contact with preexisting older Te ores.

5. Discussion

Figure 2 illustrates all available data (Tables 1 and 2) for the apparent half life of $^{130}$Te, calculated directly from $^{130}$Xe*/Te and an adopted age. For consistency, a single age (shown at left) is used for each locality, with the apparent half life recalculated if the original authors used a different age. It is evident that the data disagree well beyond stated analytical errors and that there is no convergence as analytical techniques improve. As stated previously, this cannot be attributed to subtleties in isotopic resolution of radiogenic $^{130}$Xe*. It also appears that this disagreement cannot be attributed to an “interlaboratory bias” effect in measuring the
concentration of Xe, since individual laboratories (including ours; see Table 2) report quite diverse apparent half-lives. We conclude that resolution of this problem must be sought not in laboratory technique but in the geology of the ores.

We first consider the argument (espoused by the Heidelberg group) that the half-life of $^{130}$Te should be based on measurement of $^{130}$Xe* ($^{130}$Xe derived from double-beta decay of $^{130}$Te) in samples whose age is determined independently by another chronometer based on noble gas retention. This argument explicitly recognizes the possibility that an ore could form later than and independently of its host rock, and also that final quantitative retention of $^{130}$Xe* might not begin until some significant time after the formation of the ore itself, e.g., because of gas loss during metamorphism, recrystallization, deformation, etc. The commonly cited candidate...
chronologies are U-He, K-Ar and U-(fission) Xe (or Kr). Actually, the U-He system is widely considered suspect because of the relative ease of diffusive loss of daughter $^4$He, so in practice the usual candidates are K-Ar and U-Xe (Kr). Of these, U-Xe seems especially appropriate because the fission Xe daughter is the same element as $^{130}$Xe.

While this approach is sound in principle, in practice its application is limited and can be severely flawed. The difficulty is chiefly that the radioactive parent elements commonly occur only in rather low abundance in Te ores. Besides any analytical difficulties that may arise in measurement of very low concentrations, this generates two potential problems in data interpretation. One is that the inventory of the radioactive element may be dominated by some accessory or trace phase other than the Te mineral of interest. In such cases sampling heterogeneity problems could be severe. For Good Hope native Te, for example, KIRSTEN et al. (1968) report $[K] = 5.6$ ppm and $[U] = 21$ ppb, while our isotope dilution analyses of Good Hope native Te yield $[K] = 0.2$ ppm and $[U] \geq 3$ ppb. Also, even if a valid age is obtained, if the age is for some trace phase accidentally included but not congeneric with the ores it will not be relevant to Xe accumulation in the Te phase of interest. The second problem is that if the concentration of the parent is low compared to source materials or country rock, a significant fraction of the radiogenic daughter inventory may be an inherited component trapped at the time of mineralization or isotopic closure. If an inherited component is present and not distinguished from the in situ component, an erroneous age will be inferred. Such an excess daughter effect is well known for Ar and He (cf. OZIMA and PODOSEK, 1983). On the basis of isotopic and stepwise release patterns from Good Hope native Te, MANUEL (1986) and LEE and MANUEL (1991) conclude that excess radiogenic $^{40}$Ar is indeed present and interferes with age determination. LEE and MANUEL (1991) also compared $^{40}$Ar and $^{130}$Xe release patterns in Robbr-Montbray melonite, Kalgoorlie krennerite and Boliden tellurobismuthite, concluding that the K-Ar system is essentially useless for determining Xe retention times in these ores. As noted earlier, we also conclude that there must be an analogous excess or inherited fission Xe component in some of the ores studied here (or severe sampling heterogeneity), in that there is more fission Xe present than could be produced by the measured U concentration.

We conclude that application of normal isotopic geochronometers to Te ores, as a means of estimating Xe retention ages, is compromised by potentially severe complications and should be done only with careful attention to the distribution of both parent and daughter and to the possibility of inherited excesses of the daughter isotope. In practice, information on the mineralogical distribution of the relevant elements, particularly on whether they are distributed in the ore mineral or concentrated in trace phases, is essentially completely lacking. Inheritance is possibly less of a problem in samples for which parent element abundance is not so extremely low, e.g., U in Boliden tellurobismuthite or Cripple Creek krennerite, but even in such cases distribution may be a problem. GERLING et al. (1967), for example, noted that U in Boliden tellurobismuthite was distributed inhomogeneously, so sampling heterogeneity may be responsible for the quite disparate U-Xe ages (see Table 1) found by GERLING et al. (1967) and KIRSTEN et al. (1967b).
We next consider the argument (espoused by the Rolla group) that a well-established mineralization or host rock age is an upper limit to the $^{130}$Xe retention age of a given ore, whence it will provide an upper limit to the half-life of $^{130}$Te. The logic seems unimpeachable, but we will later suggest a loophole. Logical loopholes aside, however, this argument is also not without its difficulties. If the argument is correct, and the half-life of $^{130}$Te is therefore $\leq 0.8 \times 10^{21}$ yr, most of the Te ores so far studied (Fig. 2) must have lost most of their radiogenic $^{130}$Xe$^*$, at least if their otherwise estimated formation ages are even approximately correct. In some cases where several samples from a given locality have been analyzed—Boliden and Kalgoorlie—the inferred degree of loss is highly variable. In others—Mattagami and Good Hope/Vulcan—it is not; it would have to be supposed that the examined samples, from these two different localities and of different ages, only coincidentally all lost about two thirds of their $^{130}$Xe$^*$. The Rolla group frequently asserts that tellurides are “soft” and “low temperature” minerals (e.g., MANUEL, 1991) in support of the apparent pervasiveness of Xe loss; we consider such characterization to be of little merit in the absence of any serious attempt to quantify the diffusive properties of Xe in the relevant minerals or to estimate closure temperatures.

Perhaps the most serious shortcoming of the Rolla argument as presented is that there has been no attempt to relate the inferred gas loss to anything else known about the geological history of the ore deposits. For Mattagami and Good Hope/Vulcan, for example, it would have to be supposed that isotopic closure occurred at around 840 Ma and 460 Ma, respectively. Proponents of the shorter half-life have not sought independent evidence for geological events at these times; we have, and we know of no such evidence. In addition, in comparing stepwise heating release of $^{130}$Xe$^*$ with reactor-induced $^{131}$Xe, RICHARDSON et al. (1986) find no evidence for significant partial diffusive loss in several samples, i.e. no evidence for greater loss in more easily degassed sites. For cases like Boliden (Table 1, Fig. 2), then, such all-or-none loss would suggest several gas loss episodes, essentially throughout the age of the deposit. From a geological perspective, such a scenario seems rather implausible. These considerations do not constitute a refutation of the argument that the half-life of $^{130}$Te is $\leq 0.8 \times 10^{21}$ yr, but they should at least inspire caution.

There is a third approach, that of tying evaluation of the half-life of $^{130}$Te to the half-life of $^{82}$Se (see Table 4). This is actually a variant of the first approach, in that it amounts to using double-beta decay of $^{82}$Se to $^{82}$Kr as a gas retention chronometer for calculation of the half-life of $^{130}$Te. This approach predicates the existence of ores in which both Te and Se are sufficiently enriched that excesses of the daughters $^{130}$Xe and $^{82}$Kr are prominent, and assumes that if Xe and Kr are lost at all they are lost in the same proportion. This approach does not suffer from some of the potential liabilities discussed above: By hypothesis, the parent (Se) is not an extremely low-abundance element likely to be concentrated in trace phases and so heterogeneously distributed and/or possibly irrelevant to gas retention in the ore mineral of interest. We note that if there actually is any diffusive loss, Kr would presumably be lost faster than Xe, so that the observed $^{130}$Xe/$^{82}$Kr ratio, if in error because of this, would be too high, i.e. the apparent half-life would be too short; allowance for possible differential diffusive loss would thus render the apparent half-life a lower limit.
Table 4. Data for the ratio of halflives of $^{130}$Te and $^{82}$Se.

<table>
<thead>
<tr>
<th>Location</th>
<th>Mineral</th>
<th>$^{130}$Xe*/$^{130}$Xe</th>
<th>$^{82}$Kr*/$^{82}$Kr</th>
<th>$^{130}$Xe*/Te</th>
<th>$^{82}$Kr*/Se</th>
<th>$T_{1/2}(130)$</th>
<th>$T_{1/2}(82)$</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boliden</td>
<td>tellurobismuthite</td>
<td>1.56±0.41</td>
<td>17±5</td>
<td>1.02±0.14</td>
<td>9.7±2</td>
<td></td>
<td></td>
<td>Srinivasan et al. (1973)</td>
</tr>
<tr>
<td>Boliden</td>
<td>selenokobellite</td>
<td>3.14±0.44</td>
<td>13.7±2.7</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Kirsten (1983)</td>
</tr>
<tr>
<td></td>
<td>(Pb$_2$(Bi,Sb)$_2$(S,Se)$_5$)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Kirsten et al. (1986)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.78</td>
<td>2.61±0.11</td>
<td>13.1±1.2</td>
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<td></td>
<td></td>
<td>0.82</td>
<td>2.96±0.11</td>
<td>11.6±1.2</td>
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<tr>
<td>Kuusamo</td>
<td>kitkaite</td>
<td>0.93</td>
<td>0.82</td>
<td>6.74±0.46</td>
<td>2.10±0.23</td>
<td>7.3±0.9</td>
<td></td>
<td>Lin et al. (1986)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.64</td>
<td>0.35</td>
<td>8.40±0.98</td>
<td>2.36±0.30</td>
<td>6.5±1.1</td>
<td></td>
<td>&quot;</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.53</td>
<td>0.33</td>
<td>6.15±0.25</td>
<td>1.65±0.08</td>
<td>6.3±0.4</td>
<td></td>
<td>Lin et al. (1988a)</td>
</tr>
</tbody>
</table>
Differential diffusive loss of Kr and Xe, however, is usually considered not a problem in general, and it should again be noted that Richardson et al. (1986) interpret their data for several ores as indicating that Xe loss is more nearly an all-or-none phenomenon rather than one of variable degrees of loss from various parent sites. If there is no differential loss between Kr and Xe, the ratio of the half-lives of $^{130}$Xe and $^{82}$Se follows directly from the ratio of $^{82}$Kr*/$^{82}$Se to $^{130}$Xe*/$^{130}$Te, independently of the sample age. The halflife of $^{82}$Se is apparently relatively well known, at least in comparison with that of $^{130}$Te: The range based on geochemical measurements is $(1.0–1.3) \times 10^{20}$ yr, in good agreement with the $^{82}$Se halflife of $(1.1^{+0.8}_{-0.3}) \times 10^{20}$ yr based on laboratory detection of double-beta decay events (Elliot et al., 1987).

Kirsten et al. (1986) advocate a “best estimate” $^{82}$Se halflife of $1.3 \times 10^{20}$ yr, based on independently dated Se ores. Kirsten et al. (1986) also adopt 12.5 as the halflife ratio of $^{130}$Te to $^{82}$Se, on the basis of Heidelberg analyses of Boliden selenokobellite (Table 4). These two figures lead to a $^{130}$Te halflife of $1.6 \times 10^{21}$ yr, significantly higher than the $0.8 \times 10^{21}$ yr halflife recommended by the Rolla group (Manuel, 1986, 1991; Richardson et al., 1986) but also significantly lower than the $2.55 \times 10^{21}$ yr value recommended by Kirsten (1983). Kirsten et al. (1986) regard this as a yet unexplained discrepancy, and so qualify their recommended $^{130}$Te halflife as in the range $(1.5–2.8) \times 10^{21}$ yr. The Rolla group recommends a lower value, around 6.3, for the ratio of half-lives, based on their Kuusamo kitkaite analyses (Table 4), and a lower value for the halflife of $^{82}$Se, around $1.0 \times 10^{20}$ yr, roughly consistent with their lower preferred value for the halflife of $^{130}$Te, around $0.8 \times 10^{21}$ yr.

Rather than try to choose between the alternatives recommended by different reviewers, as described above, we note that even the approach of comparing $^{130}$Te and $^{82}$Se yields inconsistent results. Independently of absolute halflives or mineral gas retention ages, different mineral analyses should yield the same inferred ratio of halflives. They manifestly do not (Table 4): The discrepancy is at least a factor of two, well beyond stated analytical errors. Some basic assumption is violated.

Confronted with these discrepant data and conflicting arguments, we conclude that the halflife of $^{130}$Te cannot be considered well established with any great confidence, some reviewers’ assertions to the contrary notwithstanding. Other factors which might distort interpretation must be considered, and other criteria, specifically geological criteria, must be found and applied to distinguish which data are most relevant to the halflife problem.

We suggest that one potentially relevant effect, already alluded to in connection with other radiogenic gases, is the possible inheritance of radiogenic $^{130}$Xe: Specifically, we conjecture that the fluids from which some Te ores formed may have carried excess $^{130}$Xe, which was incorporated into the ores as they formed. If so, and if laboratory analysis does not distinguish between such trapped $^{130}$Xe* and subsequently produced in situ $^{130}$Xe*, the ore will have too much $^{130}$Xe* for its age, i.e. the halflife will appear to be shorter than it actually is. This is a well known effect for the more abundant radiogenic gases $^4$He and $^{40}$Ar. For some of the ores considered here, inherited $^{40}$Ar (Lee and Manuel, 1991) and fission Xe (see above) have been invoked explicitly, so this effect is at least qualitatively plausible for
$^{130}$Xe. The possibility of inherited $^{130}$Xe is a logical loophole in the argument that the lowest apparent half-life calculated from a constraining age sets an upper limit on the true half-life. Once this possibility is acknowledged, the difficulties described above for this interpretation may assume greater weight. It should be noted, however, that this hypothesized phenomenon requires rather special circumstances. Isotopically identifiable excesses of radiogenic $^4\text{He}$ and $^{40}\text{Ar}$ or fission Xe could reasonably be acquired from essentially any kind of precursor material or through interaction of hydrothermal fluids with essentially any kind of country rock, but this is unlikely to be so for $^{130}$Xe. We are unaware of any reports of isotopically prominent excesses of $^{130}$Te in any terrestrial rocks other than Te ores, and it seems unlikely that such excesses would be generated in common rocks in which Te is a distributed trace element. It seems likely, then, that significant excess $^{130}$Xe could be inherited only from sources in which Te had previously been concentrated, i.e. a prior generation of Te ores, and only in circulation systems which did not subsequently undergo significant exchange with atmospheric gases.

At present we have no compelling evidence either for or against the notion of significant inherited excess $^{130}$Xe, and so must treat it as a possibility and an “alternative working hypothesis” to be evaluated in light of other evidence. One plausibility argument in favor of inheritance is that it is otherwise difficult to understand how a sample such as Kalgoorlie krennerite (see above), which appears to be a relatively young sample formed by remobilization of older ores, can have so much $^{130}$Xe. One plausibility argument against inheritance, the converse of the argument made above, is that appeal to coincidence must be made for cases like Oya, Kuusamo and Cripple Creek accidentally having just the right amount of inherited $^{130}$Xe to yield apparent halflives around $1 \times 10^{21}$ yr. Also, if a given sample contains two or more generations of Te-derived Xe the ratio of, say, excess $^{131}$Xe (a neutron capture product) to $^{130}$Xe might be different in the two generations. This would be manifested as failure of three-isotope stepwise heating variations to define a linear correlation (i.e. more than two isotopically distinct components present) or as a linear correlation (two component mixing) in which both end-members are non-atmospheric. We find no such evidence. We further note that in some cases, such as at Oya and Suwa, the hypotheses of inheritance is only conjectural in that there is no evidence for the presence of older telluride deposits.

Approaching the problem through geological criteria, we consider that, because of the characteristics noted above, volcanogenic massive sulfide (VMS) deposits should be the most suitable for estimation of the halflife of $^{130}$Te. Of the various deposits described above, three—Mattagami Lake, Boliden and Good Hope/Vulcan—are VMS deposits. For all three deposits there are available several analyses by different laboratories. It is noteworthy that for both Mattagami Lake and Good Hope/Vulcan, the various analyses for each deposit yield results (i.e. $^{130}\text{Xe}/\text{Te}$) which can be considered to be consistent within experimental error, particularly if possible errors in Te measurement as well as Xe measurement are considered. Moreover, our own data for these two deposits yield essentially the same (differing by about 12%) apparent halflife for $^{130}$Te, about $2.9 \times 10^{21}$ yr. For the same adopted age, the Heidelberg results for Good Hope (KIRSTEN et al., 1968, 1983) are consistent with ours.
For both Good Hope and Mattagami Lake, the Rolla data differ mostly in the Te measurements; they are otherwise reasonably consistent with both the Heidelberg and St. Louis data. We find this difficult to dismiss as coincidence.

Nevertheless, the data for Boliden, which is also a VMS deposit, do not conform to this generalization. If the age is taken to be around 1950 Ma, the results of Srinivasan et al. (1972a) are reasonably consistent with the half-life estimated from the Mattagami Lake and Good Hope/Vulcan data. Independently of age, however, the available results \( ^{130}\text{Xe}/\text{Te} \) scatter widely. Without invoking hypotheses which cannot be supported by independent geological considerations, we cannot advance any compelling rationalization of these data.

Summing up all the observations and arguments, the net result is not very satisfactory in that we find no definitive resolution. There is no substantial evidence that the half-life of \(^{130}\text{Te} \) is outside the range \( 0.8 \times 10^{21} \text{ yr} \) to \( 3 \times 10^{21} \text{ yr} \). Arguments can be made for either end of this range, arguments which would appear persuasive except that they are contradictory. Faced with a situation involving complications we do not quantitatively understand, in our judgement the best course is to base a half-life estimate on the ores which are geologically least complicated, i.e. the VMS deposits, but recognizing that even these cases are not exempt from significant complication. We thus arrive at a “best estimate” \(^{130}\text{Te} \) half-life of \( (2.9 \pm 0.3) \times 10^{21} \text{ yr} \). This is based on our own results for the Good Hope/Vulcan and Mattagami Lake samples. The Heidelberg Good Hope measurements are quite consistent with this value. We consider that the Rolla measurements on Good Hope and Mattagami Lake samples are also consistent with this figure within likely analytical errors. The stated uncertainty in the half-life reflects our appraisal of analytical errors; it obviously cannot be considered to encompass interpretational uncertainties.

As a concluding remark, we note that the difficulty of dating Te ores by traditional geochronometers enhances the potential appeal of using \(^{130}\text{Te} \) itself as a geochronometer—provided, of course, that the half-life were reliably established. In the context of economic geology the use of such an unusual chronometer, particularly with such a long half-life, may seem exotic, but it need not be so, since identification of radiogenic \(^{130}\text{Xe} \) in ores with a substantial content of Te is well within the present state of the art. Nevertheless, the preceding discussion does not provide grounds for optimism in this respect. Whatever the true half-life, it is evident that when the best available independent age evaluations are made for test cases of ores already studied, some ores have too much \(^{130}\text{Xe} \), or too little, or both, or there is some other complicating factor not yet thought of, and that we cannot confidently say why this is so. Until these factors are quantitatively understood, the potential utility of \(^{130}\text{Te} \) as a geochronometer is likely to remain seriously compromised.

6. Summary and Conclusions

Experimental observations of Xe in Te ores, of the kind needed to determine the double-beta decay half-life of \(^{130}\text{Te} \), are available in seemingly ample supply. Several laboratories have analyzed several telluride minerals from several ore deposits of diverse ages, and in many cases identification of radiogenic excess \(^{130}\text{Xe} \) is unam-
biguous. If noble gas analytical technique were the limiting factor, we might optimistically expect that the half-life should be known within a few percent or, pessimistically, within some 10–20%. Instead, arguments are made for values of the half-life differing by more than a factor of three. This spread does not appear to be attributable to analytical uncertainties.

The difficulty evidently lies in estimating how long a given sample of telluride ore has been quantitatively accumulating Xe. In principle, this question could be addressed by radiometric ages based on accumulation of other radiogenic noble gas daughters such as He, Ar and (fission) Xe; in practice, however, the typically very low abundance of the relevant parent elements U and K leads to technical difficulties which, in our judgment, render this approach inapplicable. In several cases, seemingly robust mineralization ages can be derived through radiometric dating of host rocks or of associated mineralization, or by models for the evolution of common Pb contained in the Te ores themselves. On balance, calculation of the apparent half-life for such samples leads to values at the high end of the plausible range, around \(3 \times 10^{21}\) yr, but not consistently so.

It can be argued that this lack of consistency reflects common and pervasive loss of radiogenic \(^{130}\text{Xe}\) from telluride minerals. Following this reasoning, it can be further argued that the sample with the highest well-determined ratio \(^{130}\text{Te}^*/\text{Te/age}\) is least affected by loss and therefore sets an upper limit to the true half-life. By this logic, the half-life is necessarily at the low end of the plausible range, around \(0.8 \times 10^{21}\) yr. While this argument has considerable logical appeal, it is also not without difficulties, including the circumstances that the diffusion parameters of Xe in the relevant minerals are not actually known, and that no attempt has been made to relate the inferred gas loss to geologic history.

There are some minerals in which both Te and Se are major constituents. This allows the possibility of using double-beta decay of \(^{82}\text{Se}\) to \(^{82}\text{Kr}\) as a noble gas retention chronometer for \(^{130}\text{Xe}\) accumulation. This approach should be immune to most of the systematic problems of the other approaches noted above, but nevertheless it also fails to yield consistent results, so that appropriate data selection can lead to either a long or short half-life of \(^{130}\text{Te}\).

We are thus led to the conclusion that it is not presently possible to reach a definitive assessment of the half-life of \(^{130}\text{Te}\) within a plausible range approximately \(0.8 \times 10^{21}\) yr to \(3 \times 10^{21}\) yr, and correspondingly that (unless analytical problems are much greater than they appear to be) some significant geological factors have not been properly considered.

As one possibility we conjecture that inheritance of excess radiogenic \(^{130}\text{Xe}\) may be significant, i.e. that at their formation some Te ores incorporate excess \(^{130}\text{Xe}\) present in their parent hydrothermal fluids. Such an effect is well known for \(^{4}\text{He}\) and \(^{40}\text{Ar}\), and there is a strong case for inherited fission Xe in tellurides. It must be appreciated, however, that for \(^{130}\text{Xe}\) this is a rather radical suggestion; it would require that for the affected ores the mineralizing fluids acquired (and did not subsequently lose) the \(^{130}\text{Xe}\) from some prior generation of Te ores. At present, we know of no strong evidence which would support or refute this hypothesis. If the possibility of inheritance is admitted, however, the logic underlying the argument that the lowest
apparent half-life is an upper limit to the true half-life is no longer compelling.

Under these circumstances, assessment of the true half-life of $^{130}$Te rests on a variety of plausibility arguments, which can be made in favor of either a long or a short half-life. We do not consider any of the available arguments definitive. In our judgment, the “best estimate” of the half-life of $^{130}$Te should be based on ores from a specific type of deposit, the so-called volcanicogenic massive sulfide (VMS) deposits, on the grounds that they have the least complicated geological histories, the best constrained mineralization chronology, and the least complicated sources of ore constituents. Available data for two VMS deposits, the Good Hope/Vulcan and Mattagami Lake deposits, are consistent within limits understandable as analytical uncertainties and comprise the basis for our “best estimate” half-life of $2.9 \times 10^{21}$ yr. Data for a third VMS deposit, Boliden, are not consistent with this interpretation, however. In the end, we consider that definitive evaluation requires more refined analytical work and/or a better understanding of relevant geological factors.

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