Excess Helium-4 in Teggau Lake: Possibilities for a Uranium Ore Body

Abstract. *Excess ^4He (more than five times the solubility) has been measured in Teggau Lake in northwestern Ontario. A model suggests that an adjacent mass of greater than 10^4 kilograms of uranium is responsible for the observed ^4He excess. The area is favorable for pegmatitic uranium deposits, and the release of trapped ^4He from uraninite dikes larger than 30 cubic meters (1 percent U_3O_8) provides the best explanation for the excess ^4He in the lake.*

In 1972 Torgersen and Clarke began a study of dissolved helium isotopes and ^3H in lakes. These investigations (*1*, *2*) have demonstrated that the dynamic equilibrium established between ^3He production by β^- decay of in situ ^3H and the eventual escape of ^3He to the atmosphere can be used to examine a number of lake processes. Measurements of ^3H and ^3He made in this study and in other studies (*1*, *2*) have been described elsewhere (*3*, *4*).

The initial survey was conducted at the Experimental Lakes Area (*5*) in the Kenora District of northwestern Ontario, an area of many diversified and interconnected lakes set in Precambrian shield rock. Most of the lakes sampled gave ^4He concentrations within 5 percent of the solubility and the expected profile. These lakes were characterized by relatively constant but small excesses of ^3He ($\delta^3He \simeq 0$) in the surface water due to rapid mixing and exchange with the atmosphere, and gradually increasing δ^3He through the thermocline (stratification barrier) to some larger but relatively constant value in the deep water. This deep water acts as a closed system except during ventilation, which occurs with spring and fall overturn (*2*).

However, Teggau Lake (area, 13.3 km^2; mean depth, 77 m; maximum depth, 167 m; total volume, 1.4×10^8 m^3) exhibited up to 26×10^{-8} cm^3 of excess ^4He per gram of water [at standard temperature and pressure (STP)] (see Table 1), and a total of 2×10^7 cm^3 (STP) of excess ^4He (over the solubility) is present in the lake. Teggau Lake was the only one of five lakes studied in the area that shows significant ^4He excess.

This excess ^4He is the result of a combination of a solubility component and an added crustal helium component ($^3He/^4He = R_c$). The total ^3He can be corrected to give a tritiugenic ^3He component, $[^3He]_{tri}$, from ^3H decay according to the equation:

$$[^3He]_{tri} = [^3He]_{tot} - [^3He]_{sol} - ([^4He]_{tot} - [^4He]_{sol})R_c \quad (1)$$

where the subscripts tot and sol represent the total and solubility components, respectively. Although the R_c of natural gases and groundwater varies from 1.4×10^{-6}, the atmospheric value (*6*), to 10^{-9} for some uranium-bearing minerals (*7*), a plot of $[^3He]_{tot}$ versus $[^4He]_{tot}$ for the Teggau Lake data (Fig. 1) indicates a value at maximum depth of 2.1×10^{-7} (that is, the improbable case of 167-m water containing only solubility and crustal helium components and no ^3He from in situ ^3H decay). A more reasonable ratio of R_c is 1.4×10^{-7}, based on the average granitic concentrations of uranium and lithium (*8*). Using this ratio, the ^3H-^3He age (*9*) can be calculated from $[^3He]_{tri}$ (Table 1). An average deepwater ^3H-^3He age of 302 ± 100 days is calculated. (A maximum average age of 460 days is obtained if $R_c = 0$ is assumed, and the calculated age therefore has a maximum error of a factor of 2.) If we assume that the crustal ^4He residence time in the lake is the same as the tritiugenic ^3He age, a crustal ^4He supply rate of $\sim 2 \times 10^7$ cm^3 (STP) per year is obtained.

The α-decay of elements in the uranium and thorium series presents the best possible natural source of the excess ^4He. Steady-state α-decay of these elements from shield rock [the average concentrations of uranium and thorium in the Canadian shield are, respectively, 2.45 and 10.3 parts per million (ppm) (*10*)] could be the source of this excess ^4He. Considering only the uranium source of ^4He, the approximate ^4He production rate of 1.2×10^{-7} cm^3 (STP) of ^4He per gram of uranium per year (*11*) would require a body of $\sim 3 \times 10^{14}$ g of uranium to be feeding into Teggau Lake at a rate of 2×10^7 cm^3 (STP) per year (the inclusion of thorium decay increases the ^4He production rate by less than a factor of 2). This mass represents a volume of 10^{13} m^3 of ordinary shield rock. Such a volume of rock funneling its entire production of ^4He at steady state into this lake is unlikely. It is probable that a more spatially concentrated source of ^4He is present.

We will hypothesize the exponential release of trapped ^4He (from uranium decay during the aging of the shield) by some leaching mechanism. Let M_t represent the present mass of uranium (in kilograms) and let M_0 be the original mass; then

$$M_t = M_0 e^{-\lambda t} \quad (2)$$

$$-\frac{dM}{dt} = \lambda M_t \quad (3)$$

where λ is the leaching constant and t is the time over which the process has been active. Let M_tC represent the present-

Table 1. Tritium-helium data for Teggau Lake, 6 August 1973. Although the presence of excess neon resulted in larger than normal analytical errors, especially in the deep samples, the measurements of ^3He and ^4He show ^4He rapidly increasing with depth as seen in column 7. An average ^3H concentration of 258 T.U. was used with the computed $[^3He]_{tri}$ to calculate a model age (*9*). An average hypolimnion age of 302 ± 100 days is calculated. The significantly older water at 150 m is probably a relic of incomplete turnover mixing in the restricted deep water.

Depth (m)	Temperature (°C)	$[^4He]$ [cm^3 (STP)/g × 10^8]	$[^3He]$ [cm^3 (STP)/g × 10^{14}]	$[^3H]$ (T.U.)	δ^3He (%)	^4He excess [cm^3 (STP)/g × 10^8]	$[^3He]_{tri}$ [cm^3 (STP)/g × 10^{14}]	Model age (days)
0	20.4	4.71 ± 0.14	6.50 ± 0.2		− 1.4 ± 0.3	0.24	0.34	34 ± 20
7	20.3	4.95 ± 0.15	6.87 ± 0.2	252	− 0.8 ± 0.3	0.48	0.68	67 ± 20
15	15.3	5.39 ± 0.16	7.57 ± 0.2		+ 0.3 ± 0.3	0.84	1.22	121 ± 20
20	8.6	6.48 ± 0.19	8.67 ± 0.2	272*	− 4.4 ± 0.3	1.81	2.01	198 ± 20
70	4.6	8.13 ± 0.41	9.15 ± 0.6	276	−19.6 ± 0.3	3.36	2.15	212 ± 55
90	4.4	9.88 ± 0.30	10.42 ± 0.3		−24.7 ± 0.3	5.11	3.17	309 ± 30
110	4.2	13.2 ± 0.40	10.55 ± 0.6	253	−42.9 ± 0.3	8.43	2.83	278 ± 55
130	4.2	13.0 ± 0.40	9.30 ± 0.6		−48.9 ± 0.3	8.23	1.60	160 ± 55
150	4.1	24.9 ± 3.7	16.59 ± 2.2	235	−52.4 ± 1.0	20.12	7.23	688 ± 200
167	4.1	30.6 ± 6.1	11.82 ± 2.2		−72.4 ± 1.0	25.82	1.67	166 ± 200

*At 30 m.

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day mass of ^{4}He, where C is the ^{4}He concentration in the rock; C can be approximated as 1.2×10^{-4} A cm^3 (STP) per kilogram of uranium (*11*), where A is the uranium-helium age of the rock. Setting this input equal to the input of ^{4}He to Teggau Lake, 2×10^7 cm^3 (STP) per year, we have

$$2 \times 10^7 = \lambda M_t C = \lambda M_t (1.2 \times 10^{-4} A)$$

To estimate M_t, we assume an upper limit of 2.4×10^9 years for the age of Canadian shield rock. Therefore,

$$2 \times 10^7 = (1.2 \times 10^{-4})(2.4 \times 10^9) M_t \lambda$$

Rearranging, we have

$$\frac{69.4}{\lambda} = M_t \qquad (4)$$

Since $\lambda = 0.693/t_{1/2}$ (where $t_{1/2}$ is the half-time of the leaching process), Eq. 4 can be substituted into Eq. 2 and rearranged to give

$$M_0 = 100 t_{1/2} e^{t_{event}/\lambda} \qquad (5)$$

where t_{event} marks the beginning of the present leaching process. A reasonable estimate for t_{event} might be 10,000 years, the time since the last glacial retreat (*12*) and the beginning of the present hydrologic mode. Table 2 shows values of M_t and M_0 for $1 \leq t_{1/2} \leq 10^8$ years and t_{event} = 10,000 years and 2.4×10^9 years. Eliminating unreasonable values of M_0, we conclude that M_t is most likely of the order of 10^4 kg of uranium or more.

At a uranium concentration of 2.45 ppm for ordinary shield rock, the above limit requires ^{4}He release to occur from $>10^6$ m^3 of shield rock. Considering the very low porosity of shield rock [~ 1 percent (*13*)], this would appear to be an improbably large volume for shield hydrology. Two possibilities remain that might account for the observed ^{4}He flux into Teggau Lake: (i) leaching of trapped ^{4}He from a concentrated source, that is, uranium minerals, and (ii) a special funneling mechanism to gather the release from $>10^6$ m^3 of ordinary shield.

The Department of Mines and Technical Surveys of Canada has classified the area around Teggau Lake as one favorable for the occurrence of pegmatitic uraninite (*14*). Several reports (*14–16*) also indicate the occurrence of uraninite in the Kenora area, most of it occurring as a constituent part of numerous narrow stringers and dikes with large amounts of biotite. These intrusives into Keewatin volcanics are also known to thicken south of Kenora (*15*). Several claims were staked in the 1950's, and, although assays of up to 0.7 percent were reported (0.1 percent is considered ore grade), no deposit large enough for economic recovery has been found. General interest in the Kenora District as a uranium area consequently ceased in the 1960's.

Table 2. Uranium source size for Teggau Lake based on an estimate of M_t and M_0. The most likely t_{event} = 10,000 years suggests an unreasonable M_0 for $t_{1/2} < 10^2$ years. The present mass of uranium, M_t, is therefore estimated to be greater than 10^4 kg.

$t_{1/2}$ (years)	M_t (kg)	M_0 for t_{event} = 10,000 years (kg)	M_0 for t_{event} = 2.4×10^9 years (kg)
1	10^2	$>10^{99}$	
10	10^3	$>10^{99}$	
10^2	10^4	10^{34}	
10^3	10^5	10^8	
10^4	10^6	10^6	
10^5	10^7	10^7	
10^6	10^8	10^8	$>10^{99}$
10^7	10^9	10^9	10^{81}
10^8	10^{10}	10^{10}	10^{17}

If the leached mass of uranium outlined above ($>10^4$ kg) consists of uranium minerals containing 1 percent U_3O_8, the ^{4}He input to the lake can be accounted for in terms of a rock volume of >30 m^3, certainly a more feasible rock volume than $>10^6$ m^3. Although a special funneling mechanism may be present by way of a possible fault line intersecting the lake (precise geologic mapping of the area has not been done), this fault line would probably intersect several dikes and stringers which would supply the major portion of the ^{4}He release. We therefore conclude that the release of trapped ^{4}He from >30 m^3 of uranium-bearing (~ 1 percent U_3O_8) dikes and stringers provides the best explanation for the ^{4}He input to Teggau Lake.

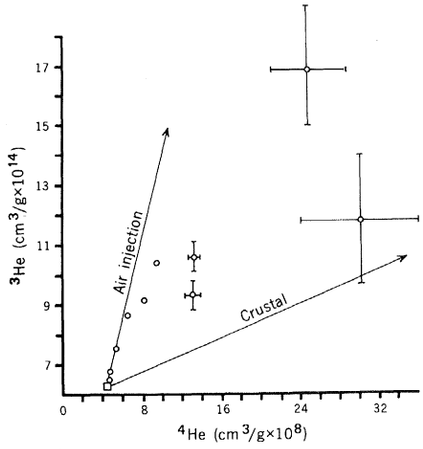

Fig. 1. A plot of ^{3}He versus ^{4}He [in cubic centimeters (STP) per gram of water] for the Teggau Lake data. The data are consistent with a binary mixing model between lake water and an additional crustal component. The R_c of this crustal component (indicated in the figure) is that of average granitic rock (*8*). The air injection line demonstrates that this phenomenon is not a simple case of supersaturation.

We have demonstrated the distinct possibility that a uranium ore body is supplying excess ^{4}He to Teggau Lake. If this possibility is confirmed by other geophysical measurements, this would be the first location of a uranium deposit on the basis of its ^{4}He release, the method outlined by Clarke and Kugler (*17*). We believe that this ^{4}He method is a very sensitive tool in lakes where deep water can act as an integrator of the ^{4}He released over several months. Further study, however, is needed to elucidate the structural controls maintaining ^{4}He release from the crust.

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References and Notes

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$$\delta^3\text{He} = \left[\frac{(^3\text{He}/^4\text{He})_{\text{measured}}}{(^3\text{He}/^4\text{He})_{\text{air}}} - 1\right] \times 100$$
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$$t = 6457 \ell n\left[\left(\frac{4.01\,[^3\text{He}]_{\text{trl}}}{[^3\text{H}]} \times 10^4\right) + 1\right]$$
where t is the age (in days), $[^3\text{He}]_{\text{trl}}$ is the radiogenic ^{3}He concentration (that produced directly by ^{3}H decay) [in cubic centimeters (STP) per gram], and $[^3\text{H}]$ is the ^{3}H concentration in tritium units, T.U. (1 T.U. = 1 ^{3}H per 10^8 ^{1}H).
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$$2.52 \times 10^{21} \times 8 \times 1.54 \times 10^{-10}\left(\frac{22{,}400}{6 \times 10^{23}}\right) \simeq 1.2 \times 10^{-7}$$

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18. We thank W. S. Broecker, H. J. Simpson, and Y. H. Li for helpful discussions. The assistance of Dr. D. W. Schindler and the staff of the ELA are gratefully acknowledged. This work was supported by National Science Foundation grants GB 36348 and GA 33124 and Canadian National Research Council grants A2951 and E2068. Lamont-Doherty Geological Observatory contribution 2599.

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The Ancient Lunar Core Dynamo

Abstract. *Lunar paleomagnetism provides evidence for the existence of an ancient lunar magnetic field generated in an iron core. Paleointensity experiments give a surface field of 1.3 gauss, 4.0 × 10⁹ years ago, subsequently decreasing exponentially. Thermodynamic arguments give a minimum value of the heat source in the core at that time: known sources, radioactive and other, are quantitatively implausible, and it is suggested that superheavy elements were present in the early moon.*

Igneous rocks and high-grade breccia returned from the moon by the Apollo project possess natural remanent magnetization (NRM) with a stability similar to that of terrestrial rocks (*1*). This paleomagnetism has been interpreted as a thermoremanent magnetization (TRM) acquired at the original extrusion and crystallization of the lavas between 4.0 and 3.2 × 10^9 years ago (*2*) and at the cooling of the breccias during their formation. Lunar magnetic anomalies have been mapped at an altitude of 100 km by subsatellite magnetometers (*3*) and at the surface both by magnetometers and by low-energy electron reflections (*4*). These anomalies have magnitudes and scales such that they evidently arise from this remanent magnetization of crustal rocks. Surprisingly, the present dipole moment arising from the magnetized outer shell is negligible (*5*), but from this null result I conclude that the moon once possessed a magnetic field that was generated in its interior and was responsible for magnetizing the rocks (*6*). This field has now disappeared. Paleointensity determinations have been made on Apollo rocks by two methods. In the Thellier-Thellier method, the rock is heated to successively higher temperatures in a zero field so as to remove the NRM stepwise, and a comparison is made with a similar stepwise acquisition of a TRM as the rock is cooled from above the Curie point in a small laboratory magnetic field. In an analogous method, strong alternating magnetic fields replace temperature in its effect on domains. The results have been interpreted as showing that the magnetizing field in which the rocks acquired their TRM decreased exponentially from 1.3 gauss at 4.0 × 10^9 years ago to 5000 γ (1 γ = 10^{-5} gauss) 3.2 × 10^9 years ago (*7*). Thus, although explanations of lunar paleomagnetism caused by the impact of meteorites or comets are being examined (*8*), this relation between paleointensity and age argues in favor of the existence of an ancient lunar field.

The source of this lunar field of internal origin is a key question with respect to the structure and thermal history of the early moon. An origin consistent with a uniformly accreted and initially cold moon, the deep interior remaining below the Curie point of iron for the first 1 to 1.5 × 10^9 years, was suggested (*9*). As the moon accreted, it acquired a uniform permanent magnetization from a solar system magnetic field retained in the solar nebula after its formation: its deep interior remained magnetized, although diminishingly so because of radioactive heating, until after the youngest lava samples examined had been extruded. This theory is no longer tenable on quantitative grounds as a result of the paleointensity determinations (*10*). According to the alternative theory, the moon has an iron core which was fluid in its early history and generated a field by a dynamo process. The absence of a field today is explained if the core either solidified or ceased dynamo action because the magnetic Reynolds number became subcritical in the time since 3.2 × 10^9 years ago (*1*). I inferred the existence of an iron core (*11*) on entirely different grounds: if creep occurs, the moon's nonhydrostatic figure must result from a second-degree harmonic convection current in its solid silicate mantle. But direct evidence for the existence of the core based on seismic signals (*12*) or electrical conductivity determinations (*13*) is, at present, only suggestive. However, the latest value (*14*) of the moment of inertia factor C/Ma^2, where C is the polar moment of inertia, M is the mass, and a is the radius, is 0.391 ± 0.002. This value is not consistent with differentiated outer shells over an otherwise uniformly dense moon, for I have shown (*15*) that these outer shells would reduce C/Ma^2 by only about 0.002 below the value of 0.4 for a uniform moon. In fact, the mean value of C/Ma^2 given above would allow the existence of a lunar core of radius b = 500 km. I will use this radius.

Suppose that an iron core exists within the moon and that it was molten (that is, at 1900°K) at least between 4.0 and 3.2 × 10^9 years ago. Thermodynamic arguments may now be applied to the core (if we think of it as a heat engine) in which a yet unknown heat source (E) within it drives convective motions which generate the dipole field. I use an argument applied by Gubbins (*16*) and Backus (*17*) to the geomagnetic dynamo, although for the case of the earth's core no critical conclusions emerge. The dipole field (H_0 at the core surface on the lunar equator) arises from a toroidal, second-degree harmonic, electric current distribution over concentric spherical surfaces within the core with an unknown radial distribution. The latter, however, can be chosen, to give a minimum ohmic dissipation P in the core, which equals 30 $b\ H_0^2/4\ \pi\sigma$, where σ is the electrical conductivity of molten iron (7.4 × 10^3 ohm^{-1} cm^{-1}) (*16*). The ratio of "useful" work done by the dynamo (that which is essential to the field generation) to the heat source driving it, P/E, is not greater than a quantity of similar form to the Carnot efficiency, that is, $k\Delta T/T$, where T is the temperature at the outside of the core and ΔT is the actual difference in temperature between the center and the surface of the core. The difference from the Carnot cycle arises as the work done in this heat engine is fed into the convecting system as heat, but no case where $k > 1$ has been found. I assume that $k = 1$ if the heat source is at the center and $k = 2/5$ if the heat source is uniformly distributed through the core (*18*). On such a scale, the superadiabatic gradient necessary for convection is very much less than the adiabatic gradient and thus makes a negligible contribution to ΔT. The adiabatic gradient is $\alpha g T/C_p$ where α is the volume coefficient of expansion of molten iron (12.2 × 10^{-5} per degree Celsius), g is the gravitational acceleration (98 cm sec^{-2} at the core surface), and C_p is the specific heat of iron per unit mass. Thus $P/E \leq (2/5)(75/1900) = 0.016$ for a uniformly distributed heat source, which is the most plausible case.

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