Helium accumulation in groundwater. II: A model for the accumulation of the crustal $^4$He degassing flux

T. TORGersen* and G. N. IVEY
Research School of Earth Sciences, Australian National University, G.P.O. Box 4, Canberra, A.C.T. Australia 2601

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Abstract—A model of $^4$He accumulation in a confined aquifer is presented which includes in situ production and a crustal degassing flux of $^4$He into the bottom of a confined aquifer. The model is consistent with the measurements of $^4$He concentration in the Great Artesian Basin, Australia (TORGersen and CLARKE, 1985) and also with the measurements of $^4$He concentration in the Auob Sandstone of Namibia (HEATON, 1981). The model presented gives a framework in which to interpret these observations and to identify and evaluate the factors which control the accumulation of $^4$He in confined aquifers.

INTRODUCTION

In Part One of this series (TORGersen and CLARKE, 1985) it was shown that the rate of accumulation of $^4$He along a hydrologic flow line in the Great Artesian Basin (GAB), Australia, varies as a function of time and distance. For groundwater ages less than 50 Kyr the rate of $^4$He accumulation is $4.6 \pm 1.2 \times 10^{-12}$ He AU (as in TORGersen and CLARKE, 1985). 1 He AU = 1 cm$^2$ STP $^4$He $\cdot$ cm$^{-3}$ H$_2$O $\cdot$ yr$^{-1}$ which is in close agreement with the calculated in situ production rate (3.95 $\times$ 10$^{-12}$ He AU) based on measured U and Th concentrations in the aquifer rock. However, for groundwater ages greater than 100 Kyr, the measured rate of $^4$He accumulation is 2.91 $\pm$ 0.54 $\times$ 10$^{-10}$ He AU; a value that cannot be accounted for by in situ production or by the "weathering" release of helium trapped in mineral grains over geologic time. TORGersen and CLARKE (1985) calculated that a steady-state $^4$He flux from the whole crust degassing would accumulate in the GAB at a rate of 3.02 $\times$ 10$^{-10}$ He AU and concluded that for long times the rate of $^4$He accumulation in groundwater aquifers may be controlled by the whole crust production of $^4$He and its flux to the earth's surface.

The purpose of this work is to provide a mathematical framework in which to interpret these observations and to identify and evaluate the factors which control the accumulation of $^4$He in confined aquifers. The in situ production, crustal degassing flux, vertical diffusion coefficient, aquifer thickness, groundwater velocity and an "impermeable" upper aquiclude are identified as the primary parameters determining the concentration of $^4$He in confined aquifers as a function of vertical and horizontal position (age). The model presented provides a framework in which (1) to explore the relative importance and the variability of these parameters and (2) to interpret the observations of TORGersen and CLARKE, (1985) and others.

TORGersen (1980) has discussed the in situ production of $^4$He in groundwater systems as a function of (1) the U- and Th-series element concentrations and their distribution, (2) the fractional release of $^4$He from the rock phase and (3) the water/rock ratio. All of these factors may vary in space and time due to the local composition of the aquifer rock, dissolution-precipitation-adsorption of radioactive elements, the presence of surface coatings (i.e. CaCO$_3$) and variations in porosity. In any aquifer, one would therefore expect the in situ production rate of $^4$He to vary but it is most likely limited to a factor of 2-3. The crustal degassing source is usually quite large compared to in situ production and thus variability in this latter term is only significant when in situ production determines the accumulation rate.

Downstream in the aquifer where TORGersen and CLARKE (1985) showed the crustal degassing flux becomes the more important source term, there is no a priori reason to believe that the crustal $^4$He degassing flux is constant. The rate of $^4$He accumulation due to a crustal degassing "source" will be a function of porosity and the crustal degassing flux should be controlled by the local advection/diffusion and degassing properties of the local crust which should also vary in space and time. The tensile or compressive stresses on the crust as well as tectonic (vertical) movement should also affect the variability of the degassing flux. An alternative flux source to the whole crustal production term (TORGersen and CLARKE, 1985) is the degassing of a part of the crust which has retained a portion of its helium production over significant geologic time and is therefore "enriched in $^4$He". The release of helium from this rock would also create a high $^4$He flux into the aquifer but is unlikely to be as constant or as
widespread as the observations. A clear differentiation of these alternatives on a local scale, however, can only be judged by the agreement of the model and the data, considerably more data, and in particular data from deeper in the crust. However, as will be shown later in this work, the long-time, large-space scale average degassing flux does match the predicted whole crust production.

Whatever the depth or transfer mechanism of the \(^{4}\)He source, the conclusion of a crustal degassing flux into confined groundwater systems requires an explanation of the delay time before the degassing flux is apparent. TORGERSEN and CLARKE (1985) discussed this “crossover” delay as the diffusion (dispersion) time required to transport the degassing flux (which enters the bottom of the aquifer) through the total thickness of the aquifer and into the top fraction that can be sampled by existing bores. The diffusion (dispersion) rate is usually rather inhomogeneous (both vertically and horizontally) due to the presence of (i.e.) interbedded clay layers and changes in porosity and velocity with distance. The “crossover” is also a function of aquifer thickness which can vary considerably. The diffusion (dispersion) coefficient and aquifer thickness chosen for the model should therefore represent the average thickness and the average effective vertical transport coefficient of the aquifer system and it must be realized that individual bores may reflect more local conditions. For example, an interbedded clay layer just below “our well” may isolate it from the bottom flux. Conversely it must also be remembered that if such a system has operated for a number of cycles (residence times), steady-state transport may exist across this interbedded clay layer and the degassing flux may be fully transmitted. In this work, it is assumed that vertical diffusivity (dispersivity) and aquifer thickness are not functions of vertical or horizontal position.

If the crustal degassing flux controls the rate of \(^{4}\)He accumulation in the aquifer, the question also arises as to why “degassing” does not pass through the aquiclude at the top of the aquifer. TORGERSEN and CLARKE (1985) have quantified this loss via an advective loss. \(W_C\), where \(W\) is the vertical water (leakage) velocity of the aquiclude and \(C\) is the \(^{4}\)He concentration of the aquifer waters. For “fast” shales and the helium concentrations observed at 600 Kys in the GAB, the advective loss is less than one third of the crustal degassing flux and for “average shale permeability,” the loss is insignificant. It is anticipated that at some longer time (distance), \(W_C\) might be significant and at some even later time, the rate of accumulation of \(^{4}\)He in the aquifer may approach zero because loss terms equal source terms. However, the observation that \(^{4}\)He continues to increase implies that loss terms are less than the source terms.

In this study, a consistent simple model of a confined aquifer is proposed that quantifies the transfer and accumulation of \(^{4}\)He. The model uses simplifying assumptions and neglects the possible variability of some parameters which must exist in the “real world” as discussed above. Nevertheless, testing the model against data from the G.A.B (Australia; TORGERSEN and CLARKE, 1985) and from the Auob Sandstone (Namibia, HEATON, 1981) reveals good agreement between model predictions and observations. In addition, other studies of \(^{4}\)He in groundwater are reviewed from the literature and while these are limited in extent, they also lend (quantitative) support to the model. This agreement supports the hypothesis that vertical diffusive transport of a crustally-derived \(^{4}\)He degassing flux is primarily responsible for the delayed “crossover” and the increased rate of \(^{4}\)He accumulation. The comparison of observed degassing fluxes entering confined aquifers and the degassing fluxes calculated from the atmospheric budget also support the concept of (large-scale, long-time) whole crustal degassing.

A MODEL OF \(^{4}\)He ACCUMULATION IN A CONFINED AQUIFER

The rate of accumulation of \(^{4}\)He in a confined aquifer has been shown (TORGERSEN and CLARKE, 1985) to have two obvious components. For short distances (time), the rate of \(^{4}\)He accumulation is controlled by the \textit{in situ} production of \(^{4}\)He \textit{via} the alpha-decay of U- and Th-series elements. Since the U- and Th-series elements reside predominantly in the solid phase, the rate of accumulation of \(^{4}\)He in the liquid phase, \(P\), can be expressed as (TORGERSEN, 1980):

\[
P = J_{\text{He},\text{in}} \cdot \lambda_{\text{He}} \left(\frac{1 - \phi}{\phi}\right)
\]  

(1)

where \(J_{\text{He},\text{in}}\) is the rate of \(^{4}\)He production per gram rock (TORGERSEN, 1980, Eqn. 3; or CRAIG and LUPTON, 1981), \(\rho\) is the density of the rock, \(\phi\) is the porosity (cm\(^{-3}\)), \(\lambda_{\text{He}}\) is the release factor. TORGERSEN and CLARKE (1985) have shown, in agreement with numerous other authors, that \(\lambda_{\text{He}} \approx 1\) is most likely. They have also shown that \textit{in situ} production (Eqn. 1) greatly exceeds the “weathering” release of \(^{4}\)He (which could be considered as an additional “production” term) in agreement with the calculations of HEATON (1984).

For long distances (time), the rate of accumulation of \(^{4}\)He is controlled by the sum of the crustal \(^{4}\)He degassing flux \(F\) entering the bottom of the aquifer and the \textit{in situ} production. It is assumed that the crustal flux is uniform over the area of the basin and enters only the fluid phase of the confined aquifer. Hence,

\[
F = F_{\text{in situ}}
\]  

(2)

where \(F\) is the flux of \(^{4}\)He out of the crust and into the bottom of the aquifer and \(F_{\text{in situ}}\) is the flux of \(^{4}\)He into the water contained in the aquifer.

For the simple model configuration shown in Fig. 1, and assuming the steady-state, the species conservation equation for the concentration of \(^{4}\)He in the water of the aquifer as a
function of vertical (z) and horizontal (x) position, \( [^4\text{He}]_x \), may be written as

\[
U \frac{\partial [^4\text{He}]}{\partial x} - K_a \frac{\partial [^4\text{He}]}{\partial z^2} + P
\]  

(3)

where \( U \) is the advective velocity, assumed to be constant in this simple model and governed by Darcy’s Law, and \( K_a \) is the effective vertical diffusion (transverse dispersion) coefficient for helium in the aquifer (assumed constant) and \( P \) is the internal production in the aquifer (Eqn. 1). The appropriate boundary conditions are

(i) the initial concentration of \( ^4\text{He} \) for all depths (z) is zero

\[
[^4\text{He}]_{z,0} = 0
\]  

(4)

(ii) the flux of \( ^4\text{He} \) out the top of the aquifer is assumed to be zero (see the calculations of TORGersen and CLARKE (1985) and HEATON (1984) for the validity of this assumption)

\[
\left. \frac{\partial [^4\text{He}]}{\partial z} \right|_{z=0} = 0
\]  

(5)

(iii) the flux of \( ^4\text{He} \) into the aquifer across the bottom boundary at \( z = h \) is a constant, \( F \).  

\[
K_a \frac{[^4\text{He}](x, h)}{\partial z} \bigg|_{z=h} = F
\]  

(6)

The general solution to this problem is

\[
[^4\text{He}]_{x,z}= \left( \frac{F}{U} \right)x + \left( \frac{F}{U} \right) \int_0^x \left( \frac{hU}{K_a} \right) \left( \frac{3z^2 - h^2}{6h^2} \right)
\]

\[
- \frac{2}{x^2} \sum_{n=1}^{\infty} \frac{(-1)^n}{n^2} e^{-K_a(h+nx)/hu} \cos\left( \frac{\pi nx}{h} \right)
\]

\[= A + B[R_1 + R_2(D_z)]
\]  

(7)

The \( A \) term in Eqn. (7) represents the linear increase of \( ^4\text{He} \) concentration with distance (time) along the aquifer due to constant \( ^4\text{He} \) production. The other term, with both \( x \) and \( z \) dependance represents the increase in concentration due to the constant flux, \( F \), through the lower boundary that is redistributed by the effective vertical diffusion (dispersion) in the aquifer capped by an impermeable layer (see CARSLAw and JAeger, 1959; p. 112). The \( R_1 D_z \) term is a distance (time) correction to the decaying flux contribution which in very old groundwater (\( x = \infty \); \( x/U = t_H - \infty \) ) is vanishingly small. Only then does the \( ^4\text{He} \) concentration increase linearly with \( x/U \). The “crossover” point for depth \( z \) as the position \( x \) where the contribution to the \([ ^4\text{He} ] \) from the bottom flux becomes as large as the contribution due to the in situ production, i.e.

\[
\left( \frac{F}{U} \right) x + \left( \frac{F}{U} \right) \int_0^x \left( \frac{hU}{K_a} \right) \left( \frac{3z^2 - h^2}{6h^2} \right)
\]

\[
- \frac{2}{x^2} \sum_{n=1}^{\infty} \frac{(-1)^n}{n^2} e^{-K_a(h+nx)/hu} \cos\left( \frac{\pi nx}{h} \right)
\]

\[= A + B[R_1 + R_2(D_z)]
\]  

(8)

“Crossover” is thus strongly controlled by the aspect ratio, \( R_1 = x/h \), the ratio \( K_a h /U = 1/R_1 \), and their product, \( R_1 R_2 \). Further inspection of Eqn. (8) shows that “crossover” is proportional to \( Ph/F \), which reflects the relative strength of each source.

**RESULTS AND DISCUSSION**

The Great Artesian Basin, Australia—The data of HABERMEHL (1980) infer an aquifer thickness of 600 m (h), a flow velocity of 1 m/yr (U) and a porosity (\( \phi \)) of 20% for the Jurassic aquifer of the Great Artesian Basin. TORGersen and CLARKE (1985) showed that the in situ production (\( P \)) is \( 3.95 \times 10^{-13} \text{HeAU} \) and the whole crustal \( ^4\text{He} \) flux (\( F \)) is \( 3.62 \times 10^{-6} \text{cm}^3\text{STPHe} \cdot \text{cm}_{\text{mex}} \cdot \text{yr}^{-1} \). From FREEZE and CHERRY (1979, p. 396), the effective vertical dispersion (transverse dispersion) coefficient (\( K_a \)) is estimated as \( 4 \times 10^{-4} \text{m}^2 \cdot \text{s}^{-1} \). With these best estimates, the concentration distributions predicted by Eqn. (7) were solved numerically (ten terms in the series for values of \( x \), expressed non-dimensionally as \( x/h \) and values of \( x \) which can be converted to a hydrologic age \( t_H = x/U \). The results are shown in Figs. 2 and 3.

Figure 2 shows that in the upper part of the aquifer (\( z/h < 0.4 \)), the total \( ^4\text{He} \) concentration is only minimally affected by the bottom flux term for periods up to 40-80 Kyrs. Figure 3 shows that at 1000 Kyrs, the rate of accumulation of \( ^4\text{He} \) at all depths is nearly equal and within <10% of the \( t_H = \infty \) rate of \( ^4\text{He} \) accumulation. (The steady-state \( t_H = \infty \) situation is reached at \( K_a h /U = K_a h /h^2 = 1 \) (CARSLAw and JAeger, 1959), for the parameters indicated, this corresponds to \( t_H = 2852 \text{Kyrs} \).) For \( t_H = 2000 \text{Kyrs} \), the rate of accumulation of \( ^4\text{He} \) is within 2% of the \( t_H = \infty \) value. As a general rule (not applicable to very small \( x \) and \( t_H \)), the rate of accumulation of \( ^4\text{He} \) will be less than the \( t_H = \infty \) value for all \( z/h < 0.5 \) and the rate of accumulation of \( ^4\text{He} \) will increase to the \( t_H = \infty \) rate. For values of \( z/h > 0.5 \), the initial rate of accumulation of \( ^4\text{He} \) will be greater than the \( t_H = \infty \) rate and the rates will decrease to the \( t_H = \infty \) rate. (For small values of \( x \) (\( t_H \)), this generality can be slightly misleading but not confusing.) The approach to the \( x = \infty \) (\( t_H = \infty \)) rate of accumulation of \( ^4\text{He} \) is governed by the summation in Eqn. (7) and in particular the exponential term.

In Fig. 2, the range of \( ^4\text{He} \) concentration in the Great Artesian Basin samples with ages of less than 120 Kyrs (TORGersen and CLARKE, 1985) is indicated, and in Fig. 3, the \( ^4\text{He} \) concentrations for all wells with ages of less than 120 Kyrs are shown. The model was calculated for \( K_a = 4 \times 10^{-9} \text{m}^2 \cdot \text{s}^{-1} \), \( h = 600 \text{m} \), \( \phi = 0.2 \text{cm}_{\text{mex}} \cdot \text{cm}_{\text{mex}} \cdot \text{yr}^{-1} \), \( P = 3.95 \times 10^{-13} \text{HeAU} \) and \( F = 3.62 \times 10^{-6} \text{cm}^3\text{STPHe} \cdot \text{cm}_{\text{mex}} \cdot \text{yr}^{-1} \).
FIG. 3. The evolution curves for $^4\text{He}$ concentration based on Eqn. (7) are shown here with the $[^4\text{He}]$ vs. age data of Torgersen and Clarke (1985). The curves are labelled according to their relative depth in the aquifer expressed as $z/h$. The stippled area in the lower left shows the area covered by Fig. 2. Most of the observations fall between the $z/h = 0$ and $z/h = 0.4$, i.e. in the upper 40% of the aquifer as would be expected. The Innamincka-1 well, I, is an old oil exploration well and may be leaking $^4\text{He}$ from deeper in the aquifer. The Gap, G, and Eromanga, E, wells may also be tapping "deeper" water as explained in the text. The Mt. Margaret well, M, is most likely mixing with younger water of the overlying Cretaceous aquifer.

greater than 120 Kyrs is plotted. Since these wells were drilled to provide water for stock and agriculture, it would be expected that wells would only be drilled deep enough to obtain an adequate supply of water. Indeed all the analyses presented in Figs. 2 and 3 appear between $z/h = 0.0$ and $z/h = 0.4$ and cluster near $z/h = 0.2$. The one big exception, Innamincka-1 (I), is an old oil exploration well that was "sealed" to produce water from shallower depths. It is quite possible that leakage is occurring from a greater depth.

Figure 4 compares the reference model results (Fig. 2 and 3) at a depth $z/h = 0.2$ to model evolution curves (at $z/h = 0.2$) for various values of $K_x$, $\phi$, and $h$. It can be seen that higher values of $K_x$ result in "crossover" occurring at smaller values of $x$ (and $t_H = x/U$) due to the more rapid vertical distribution of crustal $^4\text{He}$ through the aquifer. For a lower value of $K_x$, "crossover" occurs later and it takes more time and distance to attain the $t_H = \infty$ rate of $^4\text{He}$ accumulation.

A change in porosity, $\phi$, will undoubtedly affect $K_x$ and $U$. However, the independent effect is to alter the initial rate and the final rate of $^4\text{He}$ accumulation (see Eqns. 1 and 2). Changes (or errors) in $h$ also have an effect that is most easily recognized as a change in the $t_H = \infty$ rate of $^4\text{He}$ accumulation.

In general, the model appears to be most sensitive to the value chosen for $K_x$, which affects the distance (time) of "crossover", and is less sensitive to $h$ and $\phi$. The gross deviation of the Innamincka-1 well (I) is probably explained by leakage of waters from deeper in the aquifer (CH4 measurements would support this conclusion. HABERMEL, pers. commun. 1984) as explained previously. The Gap (G) and Eromanga (E) wells may be drilled deeper than $z/h = 0.4$ or they may be much older (100-200 Kyrs) than they were originally calculated due to stagnation in the flow line (inhomogeneity in aquifer permeability). The most likely explanation, however, is that The Gap and the Eromanga wells occur nearer the edge of the Great Artesian Basin where the primary aquifers begin to thin. The "excess" $^4\text{He}$ at the Gap and Eromanga may therefore be due to a decrease in the local value of $h$, the aquifer thickness. Therefore, in these areas, the bottom flux is felt more quickly and the $^4\text{He}$ concentration can increase more rapidly.

The Mt. Margaret (M) data appear to have too little helium according to the model. This may be the result of the loss of $^4\text{He}$, although mechanisms would be difficult to imagine, or the isolation of the well from the bottom flux of $^4\text{He}$ due to (for example) a large and
The helium concentration data of Torgersen and Clarke (1985) are shown with the base model evolution curves (Figs. 2 and 3) at the depth $z/h = 0.2$. To test the sensitivity of the model, curves (at $z/h = 0.2$) are also plotted for higher and lower values of $K_a$, $m^*$, $s^{-1}$, $\phi$, $cm^2$-$cm^3$-$s^{-1}$, and $h$, $m$. The labelled data points are as in Fig. 3.

From the Great Artesian Basin data of Torgersen and Clarke (1985) and the model developed here, it is apparent that there is agreement between the theory presented and the observations.

The Auob Sandstone, Namibia—Another data source of interest is the Auob Sandstone in the semi-arid farming district of the Auob River between Stampriet and Gochas (Namibia). An extensive network of groundwater wells used for agriculture and by Heaton (1981) in his study show the principal flow direction to be “parallel” to the river. Tredoux and Kirchner (1981) describe the geologic cross-section along the river and the sampling line of Heaton (1981) and indicate an aquifer thickness varying between 25 and 55 $m$. For this study a mean aquifer thickness, $h$, of 35 $m$ will be used.

Frommuzre (1931) crudely estimates a maximum flow velocity “of the order of a few meters per year”. The flow velocities as deduced from the $^{14}$C groundwater ages (Voegel et al., 1981; Heaton, 1981), vary from $\sim 20$ $m$-$yr^{-1}$ for groundwaters less than 3000 yrs old to $<3$ $m$-$yr^{-1}$ for samples at the limit of $^{14}$C dating. Smoothing these velocities would create a difficulty in the interpretation, and adherence to the velocities would create computational difficulties in the model. A constant value of $U$ was therefore chosen which gives reasonable values of $f_t$ and the model results are reported only as $t_h = x/U$ for direct comparison with the $^{14}$C-dated groundwaters. This can be done because Eqn. (7) can be reduced to three terms which are a function of $x/U$ and not $x$ independent of $U$.

The porosity of the Auob Sandstone is given as 0.24 (Heaton, 1981) and 0.14 (Frommuzre, 1931). The more recent determination was chosen and an in situ production rate of $4.16 \times 10^{-12}$ HeAU calculated (Heaton, 1981). The whole crustal $^4$He degassing flux is calculated as $3.12 \times 10^{-8}$ cm$^2$-HeAU $cm^{-2}$-$yr^{-1}$ from Torgersen and Clarke (1985) based on a 0–16 km upper crust ($2.8$ ppm U) and 16–36 km lower crust ($0.28$ ppm U) ($U$ concentrations from Taylor and McNellran, 1985).

Figure 5 shows the data of Heaton (1981) and the evolution curves for different $K_a$ values and the model parameters presented above. Although the $K_a = 0.4 \times 10^{-9}$ $cm^2$-$s^{-1}$ evolution curves reasonably account for the data between $z/h = 0$ and $z/h = 0.5$, this is a factor of 10 less than expected for $K_a$ (see Freeze and Cherry, 1979). Tredoux and Kirchner (1981) and Frommuzre (1931) both refer to thin coal seams and carbonaceous shales (which under anaerobic conditions lead to the observed bacterial denitrification of water in the Auob Sandstone) but there is no indication of their extent. Interbedded fine grained layers in the Auob Sandstone could reduce the effective $K_a$ to this low value, but little information is available to these authors.

Tredoux and Kirchner (1981) present additional data on the chemical zonation and piezometric head contours for the Auob Sandstone and, while the principal flow vectors are along the sampling line of Heaton (1981), there are two recharge areas: (1) upstream of the Heaton (1981) sampling line and (2) orthogonal
The helium concentration data of HEATON (1981) are shown with the model evolution curves for depth $z/h = 0$ and $z/h = 0.5$. Three sets of evolution curves are shown for various values of $K_s$. The model is based on $h = 35$ m, $P = 4.16 \times 10^{-12}$ HeAU and $F = 3.12 \times 10^{-6}$ cm$^3$STP/He·cm$^{-2}$·yr$^{-1}$. From data supplied by HEATON (1981) and TREDoux and KIRCHNER (1981) and the summary of FREEZE and CHERRY (1979), a transverse dispersion coefficient of $4 \times 10^{-9}$ m$^2$·s$^{-1}$ might be predicted. However, the data are better fitted to the value $0.4 \times 10^{-9}$ m$^2$·s$^{-1}$.

The water inflowing from the "orthogonal" recharge area is more saline, chemically dissimilar, and from a thinner section of the Auob Sandstone. This "orthogonal" recharge water appears to become increasingly important in the HEATON sampling line near and after the "crossover" point at ~16 Krys. All of these effects could produce a mixture of waters that cause the sampled wells to have $^{14}$C ages that are "too old" or to have less enrichment by crustal $^4$He flux when compared to the idealized model and higher values of $K_s$. However, in spite of these complications, the model is still able to account for the two major observations in the HEATON (1981) data (also HEATON, 1984): (1) the initial rate of $^4$He accumulation is too high to be accounted for by in situ production, and (2) there is a downstream increase in the rate of $^4$He accumulation to a much higher and seemingly constant rate.

Figure 6 shows the effect of increasing the thickness $h$ of the model aquifer. As expected, an unreasonably thick aquifer would respond "later" and provide a better fit to the measured values. However, increasing the aquifer thickness also reduces the $t_{4He} = \infty$ rate of $^4$He accumulation to values significantly less than observed.

The other parameter that must be tested is the bottom flux of $^4$He, $F$. Although this work advocates a whole crustal degassing of $^4$He in the Great Artesian Basin and in Namibia, we are aware that there may be special (and interesting) geophysical, geological, geochemical reasons that the crustal flux of $^4$He in Namibia (or elsewhere) does not include the whole crust as was previously calculated. Therefore, Fig. 7 shows the effect of a reduced (by $1/2$) bottom flux. Although the reduction in bottom flux does increase the response distance (time) of the Auob Sandstone aquifer, the $t_{4He} = \infty$ rate of $^4$He accumulation is half the measured value (HEATON, 1981) for the same reasons that an increase in $h$ changes the $t_{4He} = \infty$ rate. The overall fit of the data and the model is not good and it is again concluded that whole crustal degassing provides the best fit to the data.

HEATON (1981) realized that (1) "weathering" and the release of helium "which has been accumulating in the sandstone minerals since their formation", or (2) helium derived from the "upward diffusion of helium produced in deeper crustal rocks" could account for his observations. His latter study (HEATON, 1984) showed that "weathering" reactions are not an important source of $^4$He. The model presented here shows that the effect of crustal $^4$He degassing flux occurs very quickly in this relatively thin aquifer and that the final rate of $^4$He accumulation can be predicted from the whole crustal $^4$He flux. Further investigations of the Auob Sandstone on an areal (rather than linear) basis should determine whether the differences between the
model and the data are due to (1) exceptionally low values of \( K_s \), (2) mixing of other groundwater components from the "orthogonal" recharge area having a different \(^{14}\text{C}\) and \(^{4}\text{He}\) history, or (3) errors in the \(^{14}\text{C}\) ages (up to 6 Kys too old).

Other studies—BATH et al. (1979) report on several aquifer systems where \(^{4}\text{He}\) concentration is in excess of that predicted by \textit{in situ} production and \(^{14}\text{C}\) ages. HEATON (1984) presents contours of \(^{4}\text{He}\) concentration in the Aubob Sandstone near Aranos (~75 km NW of Stampriet) which (from HEATON (1984). Fig. 5 only) indicate a rate of \(^{4}\text{He}\) accumulation significantly in excess of the predicted \textit{in situ} production. HEATON and VOGEL (1979) conclude that \(^{4}\text{He}\) accumulation rates in excess of \textit{in situ} production may be the result of \(^{4}\text{He}\) migration associated with known CH\(_4\) migration. DATTA et al. (1980) concluded that excess \(^{4}\text{He}\) in their unconfined aquifer is the result of preferential basaltic basement degassing along fault zones.

HEATON (1981, 1984) present \(^{4}\text{He}\) data from the Uitenhage Artesian Aquifer (South Africa), where the "weathering" release of \(^{4}\text{He}\) is insignificant and the rate of \(^{4}\text{He}\) accumulation is more than 5 times the maximum rate of \textit{in situ} production over the \(^{14}\text{C}\) time scale of 20 Kys. HEATON draws particular attention to samples having identical \(^{14}\text{C}\) ages (boreholes only 700 m apart) that differ in \(^{4}\text{He}\) concentration by a factor of two. The results of this study suggest that the samples may be from different depths.

The \(^{14}\text{C}\)-dated waters of the Bunter Sandstone (England, \( h = 100-300 \) m, ANDREWS and LEE, 1979) accumulate \(^{4}\text{He}\) at a rate about twice that predicted. Although \( t_{\text{in}} \) for "crossover" is a complex function of \( Ph/F \), \( R_1 \), \( R_2 \), etc. if the Bunter Sandstone can be considered "similar" to the Great Artesian Basin, "crossover" would be expected in the range of 16-32 Kys. ANDREWS and LEE (1979, Fig. 5) show a distinct departure from the predicted \textit{in situ} production rate of \(^{4}\text{He}\) accumulation in this age range.

All of these studies support the concept of a large additional flux of \(^{4}\text{He}\) into groundwater aquifers. However, the auxillary data required by the model presented are not available in the "accessible" literature. We cannot therefore make any more definitive statements about the applicability of the model or the crustal degassing flux in these locations.

In a different type of environment, ANDREWS et al. (1982), MARINE (1979) and BOTTOMLEY et al. (1984) have studied the \(^{4}\text{He}\) concentration in fractured rock media and found \(^{4}\text{He}\) at a large excess over that predicted from \(^{14}\text{C}\) ages (see also DAVIS and BENTLEY, 1982, with regard to MARINE, 1979). ANDREWS et al. (1982) conclude that in addition to an \textit{in situ} production component, there is either a mixing of groundwaters or a component of \(^{4}\text{He}\) due to diffusion from deeper layers and BOND ANDREWS et al. (1982) and BOTTOMLEY et al. (1984) show \(^{4}\text{He}\) profiles increasing with depth. Since this work has shown the presence of a \(^{4}\text{He}\) flux from deep crustal sources, \(^{14}\text{C}\) derived transport coefficients \( \left( K_s \right) \) for these areas and \(^{4}\text{He}\) profiles may provide a measure of the \(^{4}\text{He}\) flux from these Precambrian Shield areas.

The atmospheric \(^{4}\text{He}\) budget—The concept of a large crustal degassing source term for \(^{4}\text{He}\) which dominates the \textit{in situ} production term by several orders of magnitude can be checked against the atmospheric budget of helium. WASSERBURG et al. (1963) calculated the terrestrial \(^{4}\text{He}\) degassing flux based on the atmospheric inventory of \(^{36}\text{Ar}\), the age of the earth, and the \(^{4}\text{He}/^{36}\text{Ar}\) ratio of gas samples from various earth sources. The value obtained for the degassing rate from continental crust was \( 4.4 \times 10^{27} \) to \( 2.2 \times 10^{27} \) atoms \(^{4}\text{He} \cdot \text{cm}^2 \cdot \text{yr}^{-1} \) with a best estimate of \( 7.3 \times 10^{27} \) atoms \(^{4}\text{He} \cdot \text{cm}^2 \cdot \text{yr}^{-1} \). The measured value in the Great Artesian Basin is \( 2.9 \times 10^{27} \) HeAU (TORGESER and CLARKE, 1985) or \( 2.8 \times 10^{27} \) atoms \(^{4}\text{He} \cdot \text{cm}^2 \cdot \text{yr}^{-1} \) (h = 600 m, \( \phi = 0.2 \), 30% continental area). The measured degassing flux for Namibia is \( 3.7 \times 10^{27} \) HeAU or \( 2.5 \times 10^{27} \) atoms \(^{4}\text{He} \cdot \text{cm}^2 \cdot \text{yr}^{-1} \) (h = 35 m, \( \phi = 0.24 \), 30% continental area). The predicted whole crust production is \( 2.3 \times 10^{27} \) to \( 3.3 \times 10^{27} \) atoms \(^{4}\text{He} \cdot \text{cm}^2 \cdot \text{yr}^{-1} \) with a best estimate of \( 2.9 \times 10^{27} \) atoms \(^{4}\text{He} \cdot \text{cm}^2 \cdot \text{yr}^{-1} \) (TORGESER and CLARKE, 1985).

The measured continental \(^{4}\text{He}\) degassing flux in the Great Artesian Basin (TORGESER and CLARKE, 1985) and in Namibia (HEATON, 1981, 1984) is therefore in agreement with the atmospheric budget of \(^{4}\text{He}\) (WASSERBURG et al., 1963; OZIMA and PODOSEK, 1983) and the predicted whole crust production. Future studies should use groundwater systems to better define crustal degassing in other regimes having a variety of geophysical, geological and geochemical characteristics. A better understanding of the tectonic controls on the crustal degassing flux will clarify the relationship of the crustal degassing flux to the whole crust production and the mechanisms of fluid transport in the crust.
CONCLUSIONS

(1) The predictions of a simple analytical model for the accumulation of $^4$He in a confined aquifer are consistent with the data and the interpretation of TORGersen and CLARKE (1985) indicating that the Great Artesian Basin is accumulating a whole crustal degassing flux of $^4$He.

(2) The predictions of the simple analytical model have also been shown to be consistent with the data of HEATON (1981) and the interpretation of TORGersen and CLARKE (1985) and indicate that the Auob Sandstone of Namibia is also accumulating a whole crustal degassing flux of $^4$He. In this case, an extremely low vertical diffusivity, mixing of groundwaters, and/or dating errors may account for minor discrepancies between the model and the observations.

(3) A literature review of $^4$He studies in groundwater shows that $^4$He concentrations are often in (considerable) excess of that predicted by in situ production. This suggests that crustal degassing of $^4$He may be a common phenomena.

(4) The most critical parameter for modelling the $^4$He concentration of groundwater appears to be the value of the vertical diffusion (dispersion) coefficient. Series of piezometer nests that sample vertically and horizontally in an aquifer should yield important information about horizontal velocities and vertical dispersion. Further studies will require better knowledge of the penetration depth of the well in the aquifer ($z/h$) and the mixing of waters having different histories of in situ production and vertical dispersion.

(5) The measurement and quantitative modelling of the accumulation of the crustal degassing flux of $^4$He is in agreement with the continental degassing rate calculated from the atmospheric budget of $^4$He. This budget, however, is insufficiently constrained to say that whole crustal degassing occurs in every region.

(6) Groundwater systems may provide a means of measuring the continental degassing rate in a variety of geologic and geophysical domains.

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