THE U-TH-Pb SYSTEM

U and Th are, strictly speaking, rare earth elements, although they belong to the actinide series instead of the lanthanide series. The other rare earths we have met so far, Nd and Sm, are lanthanides. As in the lanthanide rare earths, an inner electron shell is being filled as atomic number increases in the actinides. Both U and Th generally have a valence of +4, but under oxidizing conditions, such as at the surface of the Earth, U has a valence of +6. In six-fold coordination, U\(^{4+}\) has an ionic radius of 89 pm\(^1\) (100 pico meters = 1 Å); U\(^{6+}\) has an ionic radius of 73 pm in 6-fold and 86 pm in 8-fold coordination. Th\(^{4+}\) has an ionic radius of 94 pm. These radii are not particularly large, but the combination of somewhat large radius and high-charge is not readily accommodated in crystal lattices of most common rock-forming minerals, so both U and Th are highly incompatible elements. Th is relatively immobile under most circumstances. In its reduced form, U\(^{4+}\) is insoluble and therefore fairly immobile, but in the U\(^{6+}\) form, which is stable under a wide range of conditions at the surface of the Earth, U forms the soluble oxygen complex, UO\(_2^+\). As a result, U can be quite mobile. U and Th can form their own phases in sedimentary rocks, uranite and thorite, but they are quite rare. In igneous and metamorphic rocks, U and Th are either dispersed as trace elements in major phases, or concentrated in accessory minerals (when they are present) such as zircon (ZrSiO\(_4\)), which concentrates U more than Th, and monazite ([La,Ce,Th]PO\(_4\)) which concentrations Th more than U. These elements may be also concentrated in other accessory phases such as apatite (Ca\(_5\)(PO\(_4\))\(_3\)(OH)) and sphene (CaTi(SiO\(_4\))OH). However, zircon is by far and away the most important from a geochronological perspective.

U and Th are refractory elements, and we can therefore expect the Th/U ratio of the Earth to be the same as chondrites or nearly so. There is, however, some debate about the exact terrestrial Th/U ratio, and we can be no more precise than to say it is 4±0.2\(^2\). This ratio is 3.8 in the CI chondrite Orgueil, but may be low due to mobility of U in hydrous fluid in the CI parent body.

The geochemical behavior of Pb is more complex than that of the elements we have discussed so far and consequently, less well understood. It is a relatively volatile element, so its concentration in the Earth is certainly much lower than in chondrites. It is also a chalcophile element. If the core contains, as some believe, S as the light element, it is possible that some of the Earth’s Pb is in the core (it is, however, difficult to distinguish loss of Pb from the Earth due to its volatility from loss of Pb from the silicate portion of the Earth due to extraction into the core). Pb can exist in two valence states, Pb\(^{2+}\) and Pb\(^{4+}\). Pb\(^{2+}\) is by far the most common state; the Pb\(^{4+}\) state is rare and restricted to highly alkaline or oxidizing solutions. The ionic radius of Pb\(^{2+}\) is 119 pm in 6-fold coordination and 129 pm in 8-fold coordination. As a result of its large ionic size, Pb is an incompatible element, though not as incompatible as U and Th (incompatibility seems to be comparable to the light rare earths). The most common Pb mineral is galena (PbS). In silicates, Pb substitutes readily for K (ionic radius 133 pm) particularly in potassium feldspar, but less so in other K minerals such as biotite. Most naturally occurring compounds of Pb are highly insoluble under most conditions. As a result, Pb is usually reasonably immobile. However, under conditions of low pH and high temperature, Pb forms stable and somewhat soluble chloride and sulfide complexes, so that Pb can sometimes be readily transported in hydrothermal solutions.

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\(^1\)In eight-fold coordination, the effective ionic radius of U\(^{4+}\) is 1.00Å. In zircon, a mineral which highly concentrations U, U is in 8-fold coordination. This is probably a pretty good indication that 8-fold coordination is the preferred configuration. The figure for 6-fold coordination is given for comparison to other radii, which have been for 6-fold coordination. Th has a radius of 1.05Å in 8-fold coordination.

\(^2\)The uncertainty results from the mobility of U. The CI carbonaceous chondrites experienced mild alteration in hydrous conditions on the parent body. U was mobilized under these conditions and thus the U/Th ratio varies in these meteorites. For this reason, they cannot be used to precisely determine the U/Th ratio of the Solar System and the Earth.
Although Pb is clearly less incompatible that U and Th, these 3 elements have been extracted from the mantle and concentrated in the crust to approximately the same degree. The reason for this is not yet completely understood, and we will discuss the problem later in the course.

The U-Th-Pb system is certainly the most powerful tool in the geochronologist’s tool chest. Table 6.1 summarizes this decay system. The reason for the power is simply that there are three parents decaying to 3 isotopes of Pb, and in particular, there are two isotopes of U which decay to Pb with very different half lives. This is important because chemical processes will not change the ratio of the two U isotopes to each other and will not change the ratio of the two Pb daughter isotopes to each other. The point is best illustrated as follows. First we write the decay equation for each of the two U decay systems:

\[
\begin{align*}
207 \text{Pb}^* &= 235 \text{U}(e^{\lambda_t t} - 1) \quad 6.01 \\
206 \text{Pb}^* &= 238 \text{U}(e^{\lambda_t t} - 1) \quad 6.02
\end{align*}
\]

where the asterisk designates radiogenic Pb and \(207\text{Pb}\), and \(\lambda_t\) are the decay constants for \(235\text{U}\) and \(238\text{U}\) respectively. If we divide 6.01 by 6.02, we have:

\[
\frac{207 \text{Pb}^*}{206 \text{Pb}^*} = \frac{235 \text{U}(e^{\lambda_t t} - 1)}{238 \text{U}(e^{\lambda_t t} - 1)} = 6.03
\]

Now if the ratio of the U isotopes is everywhere the same (as it is at the present day), 6.03 can be written as

\[
\frac{207 \text{Pb}^*}{206 \text{Pb}^*} = \frac{(e^{\lambda_t t} - 1)}{137.88(e^{\lambda_t t} - 1)} = 6.04
\]

The nice thing about equation 6.04 is that the only variable on the right hand side is time; in other words the \(\frac{207\text{Pb}^*}{206\text{Pb}^*}\) is a function only of time.

**TABLE 6.1. Parameters of the U-Th-Pb System**

<table>
<thead>
<tr>
<th>Parent</th>
<th>Decay Mode</th>
<th>(\lambda)</th>
<th>Half-life</th>
<th>Daughter</th>
<th>Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>(232\text{Th})</td>
<td>(\alpha, \beta)</td>
<td>(4.948 \times 10^{11}\text{y}^{-1})</td>
<td>(1.4 \times 10^{10}\text{y})</td>
<td>(208\text{Pb}, 8\text{He})</td>
<td>(208\text{Pb}/204\text{Pb}, 9\text{He}/5\text{He})</td>
</tr>
<tr>
<td>(235\text{U})</td>
<td>(\alpha, \beta)</td>
<td>(9.849 \times 10^{11}\text{y}^{-1})</td>
<td>(7.07 \times 10^{8}\text{y})</td>
<td>(207\text{Pb}, 7\text{He})</td>
<td>(207\text{Pb}/204\text{Pb}, 9\text{He}/5\text{He})</td>
</tr>
<tr>
<td>(238\text{U})</td>
<td>(\alpha, \beta)</td>
<td>(1.551 \times 10^{11}\text{y}^{-1})</td>
<td>(4.47 \times 10^{9}\text{y})</td>
<td>(206\text{Pb}, 6\text{He})</td>
<td>(206\text{Pb}/204\text{Pb}, 9\text{He}/5\text{He})</td>
</tr>
</tbody>
</table>

In practice what this means is that the age is independent of the parent/daughter ratio; i.e., we can relax somewhat the requirement that the system remain closed. We can also see that although we could write an equation similar to 80.3 using \(232\text{Th}\) and \(208\text{Pb}\) instead of \(235\text{U}\) and \(207\text{Pb}\), there would be little advantage to doing so because Th and U are different elements and could well be lost or gained in different proportions.

If Madison Avenue were given the task of selling the U-Th-Pb system, they would probably say that you get 4 dating methods for the price of one: \(238\text{U}/206\text{Pb}, 235\text{U}/207\text{Pb}, 232\text{Th}/208\text{Pb}, \) and \(207\text{Pb}/204\text{Pb}\). In a certain sense, this is true. However, if you bought the package, you would probably quickly discover that the first three above, applied independently, were not particularly powerful, at least in comparison to either the Pb-Pb technique or simultaneous use of a combination of several techniques (an exception might be the \(232\text{Th}/208\text{Pb}\) system, which might prove useful on separated Th-bearing minerals). The Pb-Pb method, as it is called, can be quite useful when applied independently, particularly where there is reason to believe that there has been some recent change in the parent/daughter ratio. We have mentioned in an earlier lecture that the slope on a plot of \(208\text{Pb}/204\text{Pb}\) vs. \(206\text{Pb}/204\text{Pb}\) is proportional to age since:
Equation 6.05 is very similar to equation 6.04. We would use 6.04 when either there is no significant initial Pb, or the amount of initial Pb is sufficiently small that we can make a reasonable estimate of its isotopic composition and make a correction for it. We would use 6.05 when initial Pb is present in significant quantities and has an unknown composition. Figure 6.1 shows an example of a Pb-Pb isochron that yielded a reasonably precise age. Unlike a conventional isochron, the intercept in the Pb-Pb isochron has no significance, and the initial isotopic composition cannot be determined without some additional information about parent/daughter ratios.

There are a couple of reasons why we might suspect parent/daughter ratios have changed, and hence might prefer the Pb-Pb approach to one involving parent-daughter ratios. First, the solubility of U under oxidizing conditions often leads to mobility (open-system behavior) in the zone of weathering. It has often been found that U-Pb ages are spurious, yet Pb-Pb ages seem correct. This circumstance appears to result from recent U mobility as erosion brings a rock into the weathering zone. A second situation where parent/daughter ratios would have experienced recent change is in magma generation. When melting occurs, the U and Pb isotope ratios in the magma will be identical to those in the source (because the isotopes of an element are chemically identical), but the U/Pb ratio (and Th/Pb) ratio will change, as the chemical behaviors of U and Pb differ. So conventional dating schemes cannot generally provide useful geochronological information about sources of magmas. However, the Pb-Pb dating method can, at least in principle, provide useful information, because the Pb isotope ratios of a magma are representative of the source and the method does not depend on parent/daughter ratios. Essentially, what we are doing is allowing volcanism to ‘sample’ the source, generally the mantle, but sometimes the lower continental crust. The sample is representative of the isotopic composition of the source, but not representative of the elemental chemistry of the source. The relationship between Pb isotope ratios in mantle-derived magmas has lead to the conclusion that heterogeneities in the mantle must have existed for times on the order of 1-2 Ga. This is an extremely important constraint not only on the chemical evolution of the mantle, but also on its dynamics.

The U-Pb system achieves its greatest power when we use the $^{238}\text{U}-^{206}\text{Pb}$, $^{235}\text{U}-^{207}\text{Pb}$, and $^{207}\text{Pb}-^{206}\text{Pb}$ methods in combination. In many instances, it can actually be used to ‘see through’ open system behavior and obtain an age of initial crystallization. We shall examine this in the next lecture.

### Th/U Ratios

Provided Th/U ratios are constant and known in a set of samples we wish to date, we can calculate ages from $^{208}\text{Pb}/^{206}\text{Pb}$, $^{207}\text{Pb}/^{206}\text{Pb}$ isochrons just as we can using $^{207}\text{Pb}$ and $^{208}\text{Pb}$. However, although U and Th are geo-
chemically similar and the Th/U ratio is not likely to vary much, it would not be prudent to assume the ratio is actually constant for geochronological purposes. Furthermore, there is little reason to do so, since we can already compute the age using $^{207}$Pb and $^{206}$Pb. But it may be useful in some circumstances to turn the problem around and compute the Th/U ratio from the age and the slope of the data on a plot of $^{206}$Pb/$^{204}$Pb vs. $^{208}$Pb/$^{204}$Pb. The basis of this is as follows. We write the usual growth equations for $^{206}$Pb and $^{208}$Pb:

$$\frac{^{206}\text{Pb}}{^{204}\text{Pb}} = \left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_0 + \frac{^{235}\text{U}}{^{204}\text{Pb}}(e^{\lambda_{^{235}U}t} - 1)$$  \hspace{1cm}  6.06

$$\frac{^{208}\text{Pb}}{^{204}\text{Pb}} = \left(\frac{^{208}\text{Pb}}{^{204}\text{Pb}}\right)_0 + \frac{^{232}\text{Th}}{^{204}\text{Pb}}(e^{\lambda_{^{232}Th}t} - 1)$$  \hspace{1cm}  6.07

Subtracting the initial ratio from each side of each equation and dividing 6.7 by 6.6 we have

$$\frac{\Delta\left(\frac{^{208}\text{Pb}}{^{204}\text{Pb}}\right)}{\Delta\left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)} = \frac{^{232}\text{Th}}{^{238}\text{U}} \left(\frac{^{204}\text{Pb}}{^{204}\text{Pb}}\right)(e^{\lambda_{^{232}Th}t} - 1)$$  \hspace{1cm}  6.08

or

$$\frac{\Delta\left(\frac{^{208}\text{Pb}}{^{204}\text{Pb}}\right)}{\Delta\left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)} = \kappa(e^{\lambda_{^{232}Th}t} - 1)$$  \hspace{1cm}  6.09

where $\kappa$ is used to designate the $^{232}\text{Th}/^{238}\text{U}$ ratio. Using $\mu$ to designate the $^{238}\text{U}/^{204}\text{Pb}$ ratio, the parent-daughter ratio of the Th-Pb system is the product $\mu \kappa$.

Equation 6.09 tells us that the slope of a line on a plot of $^{208}$Pb/$^{204}$Pb vs. $^{206}$Pb/$^{204}$Pb is proportional to time and $\kappa$, provided that $\kappa$ does not vary. If we can calculate $t$ from the corresponding $^{206}$Pb/$^{204}$Pb–$^{208}$Pb/$^{204}$Pb slope, we can solve 6.09 for $\kappa$. If, however, $\kappa$ varies linearly with $\mu$, a straight line will still result on the $^{208}$Pb/$^{204}$Pb vs. $^{206}$Pb/$^{204}$Pb plot and our estimate of $\kappa$ will be incorrect.

**THE U-TH-PB SYSTEM: ZIRCON DATING**

Zircon (ZrSiO$_4$) is a mineral with a number of properties that make it extremely useful for geochronologists (Figure 6.2). First of all, it is very hard (hardness 7 to 8), which means it is extremely resistant to mechanical weathering. Second, it is extremely resistant to chemical weathering and metamorphism. For geochronological purposes, these properties mean it is likely to remain a closed system. Third, it concentrates U (and Th to a lesser extent) and excludes Pb, resulting in typically very high $^{238}\text{U}/^{204}\text{Pb}$ ratios. It is quite possibly nature’s best clock. Finally, it is reasonably common as an accessory phase in a variety of igneous and metamorphic rocks.

The very high $^{238}\text{U}/^{204}\text{Pb}$ ratios in zircon (and similar high $\mu$ minerals such as sphere and apatite) provide some special geochronological opportunities and a special diagram, the concordia diagram, has been developed to take advantage of them. The discussion that follows can be applied to any other system with extremely high $^{238}\text{U}/^{204}\text{Pb}$ ratios, but in practice, zircons constitute the principle target for Pb geochronologists.

A concordia diagram is simply a plot of $^{206}\text{Pb}*/^{238}\text{U}$ vs. $^{207}\text{Pb}*/^{235}\text{U}$. You should satisfy yourself that both of these ratios are proportional to time. In essence, the concordia diagram is a plot of the $^{238}\text{U}–^{207}\text{Pb}$ age against the $^{235}\text{U}–^{206}\text{Pb}$ age. The ‘concordia’ curve on such a diagram that is the locus of points where the $^{238}\text{U}–^{207}\text{Pb}$ age equals the $^{235}\text{U}–^{206}\text{Pb}$ age. Such ages are said to be concordant. Figure 6.3 is an example of a concordia diagram.

The best way to think about evolution of Pb/U ratios is to imagine that the diagram itself evolves with time, along with its axes, while the actual data point stays fixed. Let’s take a 4.0 Ga old zircon as an example. When it first formed, or “closed”, it would have plotted at the origin, because had anyone been around to analyze it, they would have found the $^{207}\text{Pb}*/^{235}\text{U}$ and $^{206}\text{Pb}*/^{238}\text{U}$ ratios to be 0. Initially, $^{207}\text{Pb}*/^{235}\text{U}$ would have increased rapidly, while the $^{206}\text{Pb}*/^{238}\text{U}$ would have been increasing only slowly. This is because 4.0 Ga ago there was a lot of $^{235}\text{U}$ around (recall that $^{235}\text{U}$ has a short half-life). As time passed, the increase in $^{207}\text{Pb}*/^{235}\text{U}$ would have slowed as the $^{235}\text{U}$ was ‘used up’. So imagine...
that the diagram initially 'grows' or 'expands' to the left, expanding downward only slowly. Had someone been around 3.0 Ga ago to determine 'zircon' ages, he would have drawn it as it appears in Figure 6.4 (of course, he would have labeled the 3.0 Ga point as 0, the 4.0 Ga point as 1.0, etc.).

Any zircon that has remained as a completely closed system since its crystallization must plot on the concordia line. What happens when a zircon gains or loses U or Pb? Let's take the case of Pb loss, since that is the most common type of open-system behavior in zircons. The zircon must lose $^{207}\text{Pb}$ and $^{206}\text{Pb}$ in exactly the proportions they exist in the zircon because the two are chemically identical. In other words, a zircon will not lose $^{206}\text{Pb}$ in preference to $^{207}\text{Pb}$ or visa versa.

Let's take the specific case of a 4.0 Ga zircon that experienced some Pb loss during a metamorphic event at 3.0 Ga. If the loss was complete, the zircon would have been reset and would have plotted at the origin in Figure 6.4. We could not distinguish it from one that formed 3.0 Ga. Suppose now that that zircon had lost only half its Pb at 3.0 Ga. During the Pb loss, the $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ would have both decreased by half. Consequently, the point would have migrated half way to the origin. At 3.0 Ga, therefore, it would have plotted on a 'cord', i.e., a straight line, between its initial position on the concordia curve, the 4.0 Ga point, and the origin (Figure 6.5a) at 3.0 Ga. Had it lost some other amount of Pb, say 30\% or 80\%, it would have plotted on the same cord, but further or nearer the origin. The line is straight because the loss of $^{207}\text{Pb}$ is always directly proportional to the loss of $^{206}\text{Pb}$. The origin in Figure 9.3a corresponds to the 3.0 Ga point on the concordia in Figure 6.5b. So, in Figure 6.5b, the zircon would lie on a cord between the 4.0 Ga and the 3.0 Ga point. We would say this is a 'discordant' zircon.

Figure 6.2 Upper. Separated Zircon crystals. Notice the zoning. Lower. Strongly zoned zircon showing differing ages of spots analyzed by ion probe.
The intercepts of this cord with the concordia give the ages of initial crystallization (4.0 Ga) and metamorphism (3.0 Ga). So if we can determine the cord on which this discordant zircon lies, we can determine the ages of both events from the intercepts of that cord with the concordia. Unfortunately, if our only data point is this single zircon, we can draw an infinite number of cords passing through this point, so the ages of crystallization and metamorphism are indeterminate. However, we can draw only 1 line through 2 points. So by measuring two zircons (or populations of zircons) that have the same crystallization ages and metamorphism ages, but have lost different amounts of Pb, and hence plot on different points on the same cord, the cord can be determined. The closure age and partial resetting ages can then be determined from the intercepts. (as usual in geochronology, however, we are reluctant to draw a line through only two points since any two points define some line; so at least three measurements are generally made). In practice, different zircon populations are selected based on size, appearance, magnetic properties, color, etc. While zircon is generally a trace mineral, only very small quantities, a few milligrams, are needed for a measurement. Indeed, it is possible to analyze single zircons and even parts of zircons.

U gain would affect the position of zircons on the concordia diagram in the same manner as Pb loss; the two processes are essentially indistinguishable on the concordia diagram. U loss, on the other hand, moves the points away from the origin at the time of the loss (Figure 6.6). In this case, the zircons lie on an extension of a cord above the concordia. As is the case for Pb loss, the upper intercept of the cord

Figure 6.3. The concordia diagram.

Figure 6.4. A concordia diagram as it would have been drawn at 3.0 Ga.
gives the initial age and the lower intercept gives the age of U loss. U loss in less common than Pb loss. This is true for two reasons. First, U is happy in the zircon, Pb is not. Second, Pb will occupy a site damaged by the alpha decay, making diffusion out of this site easier. Radiation damage is a significant problem in zircon geochronology, and one of the main reasons ages can be imprecise. U-rich zircons are particularly subject to radiation damage. Heavily damaged crystals are easily recognized under the microscope and are termed metamict.

Pb gain in zircons is not predictable because the isotopic composition of the Pb gained need not be the same as the composition of the Pb in the zircon. Thus Pb gain would destroy any age relationships. However, Pb gain is much less likely than other open system behaviors.

Zircons that have suffered multiple episodes of open system behavior will have U-Pb systematics that are difficult to interpret and could be incorrectly interpreted. For example, zircons lying on a cord between 4.0 and 3.0 Ga that subsequently lose Pb and move on a second cord toward the 2.0 Ga could be interpreted as having a metamorphic age of 2.0 Ga and a crystallization age of between 4.0 and 3.0 Ga.

Continuous Pb loss from zircons can also complicate the task of interpretation. The reason is that in continuous Pb loss, zircons do not define a straight line cord, but rather a slightly curved one. Again imagining that the concordia diagram grows with time, a zircon loosing Pb will always move toward the origin. However, the position of the origin relative to the position of the zircon moves with time in a non-linear fashion. The result is a non-linear evolution of the isotopic composition of the zircon.

Given the mechanical and chemical stability of zircon, it should not be surprising that the oldest terrestrial material yet identified is zircon. Until a decade ago, the oldest dated terrestrial rocks were the Isua gneisses in Greenland. These are roughly 3850 Ma old. Work published in 1989, revealed that the Acasta gneisses of the Slave Province (Northwest Territories, Canada) are 3.96 Ga old. These ages were determined using an ion probe to date the cores of zircon crystals extracted from
these gneisses. Concordia diagrams for these gneisses are shown in Figure 6.7.

Zircons having ages in the range of 4100-4260 Ma have been identified in quartzites at Mt. Narryer and the Jack Hills in western Australia (e.g., Compston and Pidgeon, 1986). The quartzites themselves are metamorphosed sandstones that were probably deposited about 3100-3300 Ma. They contain zircons derived from a number of sources. A small fraction of these zircons has cores that are in the range of 4100-4200 Ma. The zircons were analyzed by a specially built high-resolution ion probe at the Australian National University nicknamed ‘SHRIMP’ (subsequently, the instrument has been commercialized and similar instruments are available from competing companies). The great advantage of these instrument over conventional analysis of zircons is not only that individual zircons can be analyzed, but individual parts of the zircons can be analyzed. As can be seen in Figure 6.2, zoning in zircon is not uncommon and this zoning often reflects multiple episodes of growth. The was the case with the Mt. Narryer zircons, which had complex histories suffering multiple metamorphic events between 4260 and 2600 Ma. The principal effect was the growth of rims of new material on the older cores around 3500 Ma. Conventional analysis of these zircons would not have recognized the older ages. The cores of these zircons, however, proved to be nearly concordant at the older ages. These ages determined by ion probe were initially highly controversial. By and large, however, the community has come to accept them as reliable, when performed carefully.

Figure 6.7. Concordia diagrams showing ion probe Pb-U analyses of Acasta gneiss zircons. Size of point is proportional to 1 σ analytical uncertainty. Triangles are zircon analyses done by conventional mass spectrometry. From Bowring, et al, 1989.
Subsequently even older zircons (would be more correct to say parts of zircons) were discovered in the Jack Hills of Australia (Wilde, et al., 2001). An ion probe date on one part of one of these zircons (Figure 6.8) is shown 4.404 Ga ±8 Ma. Thus the oldest known terrestrial materials are approaching the oldest ages from other planetary bodies, including the Moon, Mars, and asteroids (as represented by meteorites). They remain, however, significantly younger than the age of the Solar System, which is 4.556 Ga. Nevertheless, these very old ages seem to demonstrate that it is zircons, not diamonds, that “are forever”.

Figure 6.8. Photomicrograph of the oldest zircon known. Light area are regions of ion probe analysis. Ages determined on these areas are shown. From Wilde et al. (2001).

REFERENCES AND SUGGESTIONS FOR FURTHER READING