ISOTOPIC GEOCHEMISTRY OF THE CONTINENTAL CRUST

Isotope systems have particular value in studies of mantle geochemistry because of the difficulty of obtaining direct, representative samples of mantle. Because isotope ratios are unchanged during the magma generation process, we can use mantle-derived magmas as samples of the isotopic composition of the mantle even though they are unrepresentative of the elemental chemistry of the mantle. The upper crust is largely accessible to direct sampling, so we are less dependent on isotopic composition in geochemical studies of the continental crust. While isotope ratios of crustal melts are sometimes used to make inferences about the composition of the lower crust, the primary value of radiogenic isotope studies of the crust is the time information they contain. Thus the question of the rate of crust evolution will be of primary interest to us in this section of the course. We will nevertheless begin by attempting to assess the average isotopic composition of the crust.

![Figure 16.1. Sr and Nd isotope ratios in modern marine sediments (from Ben Othman et al., 1989). Also shown are data on the isotopic composition of suspended loads of rivers (from Goldstein and Jacobsen, 1988), and their estimated global river average ($^{87}\text{Sr}/^{86}\text{Sr} = 0.716; \epsilon_{\text{Nd}} = -10.4$).](image-url)
SEDIMENTS AS SAMPLES OF THE CRUST

The Earth is a big place and obtaining a representative sample of the continental crust is therefore a big job. Just as we let nature do some of the work of sampling the mantle for us (by bringing magmas to the surface), we take advantage of nature's sampling of the crust. Weathering and erosion are constantly removing material from the continents and depositing it in the oceans as sediment. Thus to the isotope geochemist, sediment is a sort of premixed (homogenized) and pre-powdered sample of the continental crust. Figures 16.1 and 16.2 show Sr, Nd and Pb isotopic compositions of marine sediments from the world oceans. The data provide some indication of the composition of the crust. As may be seen, $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are much higher and $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratios much lower than those of the mantle. This is precisely what we expect since we know that Rb and Nd are enriched in the crust relative to Sr and Sm.

U, Th and Pb should also be enriched in the crust in the order Th>U>Pb. Somewhat surprisingly, we find that $^{206}\text{Pb}/^{204}\text{Pb}$ ratios in the crust overlap the MORB range considerably. $^{207}\text{Pb}/^{204}\text{Pb}$ ratios are, however, systematically higher. For a given value of $^{206}\text{Pb}/^{204}\text{Pb}$, $^{208}\text{Pb}/^{204}\text{Pb}$ ratios in the crust are also systematically higher than in MORB (which are shown in these figures primarily to represent the isotopic composition of the upper mantle), indicating a higher Th/U ratio in the crust as expected. The $^{207}\text{Pb}/^{204}\text{Pb}$ ratio is an indicator of U/Pb ratios in the early part of earth's history: the $^{235}\text{U}$ present in the Earth today is only about 2% of the $^{238}\text{U}$ the Earth started out with; i.e., 98% of $^{235}\text{U}$ has already decayed to $^{207}\text{Pb}$. Half of the $^{235}\text{U}$ had already decayed by 3.8 Ga. So the high $^{207}\text{Pb}/^{204}\text{Pb}$ of the crust relative to $^{208}\text{Pb}/^{204}\text{Pb}$ tells us that in the early part of Earth's history, crustal rocks, or their precursors, had a higher U/Pb ratio than the mantle. The half-life of $^{238}\text{U}$ is about the same as the age of the Earth, so $^{208}\text{Pb}/^{204}\text{Pb}$ has grown more linearly over Earth's history. The similarity of $^{208}\text{Pb}/^{204}\text{Pb}$ ratios in crust and upper mantle suggests the average U/Pb ratios of the two have been roughly similar over all of Earth's history.

The Pb data on the crust poses another problem: one of mass balance. We have previously noted that the Pb isotopic composition of the Earth as a whole must lie on the geochron, but both the crust, as sampled by sediments, and the upper mantle lie to the right of the geochron. If the Earth has not lost Pb, mass balance requires the presence of an additional reservoir lying to the left of the geochron. There are three obvious possibilities: the core, the lower mantle, or the lower crust. The core could only plot to the left of the geochron if Pb has been added to it at times substantially after the formation of the Earth. This idea was at one time fairly popular, but studies designed to detect depletion of the mantle in siderophile or chalcophile elements have failed to find evidence of loss of these elements from the mantle after the formation of the Earth, so the idea has largely been abandoned. The other possibilities remain. There is some evidence that this reservoir is the lower crust. We will return to the question of the isotopic composition of the lower crust in later in this lecture.

It is appropriate at this point to consider just how representative the isotopic composition of marine sediments is of the continents. Sediments are only representative of those parts of the crust undergoing erosion. This excludes almost the entirety of the lower crust. So it is therefore proper to consider sediments as representative of only the upper crust. Furthermore, it is likely to be a biased sample of the upper crust. Elevated regions erode faster and therefore generate more sediment than low plains. Tectonically active areas are typically elevated relative to stable areas. By and large, new additions to crust occur exclusively in tectonically active areas. In essence, this means sediments will be biased toward younger crust, and will have lower Sr and higher Nd isotope ratios. This biased sampling is to some degree apparent when the data are considered ocean by ocean. The Pacific Ocean is surrounded by tectonically active continental margins, and as we might expect, Sr, Nd, and Pb isotope ratios are lower in Pacific sediments than in those from the Atlantic and Pacific. Finally, we need to assure ourselves that when a rock weathers and erodes, the erosion products carried to the sea have isotopic compositions of the rock as a whole. This is probably the case for Nd because it ends up primarily in the clay fraction. A larger fraction of Sr may be carried to the sea as dissolved load; this eventually precipitates in car-
bonates. However, Sr in seawater is derived in part from the oceanic crust (entering seawater through hydrothermal activity); furthermore carbonates weather easily and much of the riverine and seawater Sr is derived from earlier marine carbonates. Thus the total composition of marine sediments, including both carbonates and detrital fractions is probably not entirely representative of the continental crust.

Figure 16.2. Pb isotope ratios in modern marine sediments. Asmeron and Jacobsen’s (1993) estimate of the average riverine suspended load is also shown.
For the Lu-Hf system, however, it is possible, and perhaps probable, that the Hf carried to the deep ocean is not representative of the composition of the eroding rock (Patchett et al. 1984; White et al. 1986). This is because much of the Hf in crustal rocks is contained in zircon, which is extremely resistant to weathering. As a result, it is not readily transported great distances from its source. It will typically be retained on the continents in sandstones. That which does reach the sea is mostly deposited on the continental shelf. The Lu/Hf ratio of zircon is lower than that of the bulk rock, so that Hf that does reach the sea may have higher $^{176}$Hf/$^{177}$Hf ratios than the eroding rock. Similarly, while much of the $^{206}$Pb is probably in phases such as feldspars that break down readily to form clays, the radiogenic isotopes will, to some degree, be retained in zircon. Thus Pb isotope ratios of marine sediment may underestimate $^{206}$Pb/$^{204}$Pb, $^{207}$Pb/$^{204}$Pb, and $^{208}$Pb/$^{204}$Pb (less so the latter) ratios of the eroding rock. Finally, some of the Pb in marine sediments is also derived from the oceanic crust through hydrothermal activity. In summary, sediments do provide a sample of the continental crust, but it is a biased one. They can nevertheless provide useful information on the composition of the crust today, but, as we shall see later in the lecture, a perspective on the evolution of the crust through time.

A slightly different, but closely related way of estimating crustal composition is to measure the isotopic composition of dissolved or suspended loads in rivers. Rivers carry most of the weathering products from the continents to the oceans (other material is carried by winds and glaciers), hence this strategy is similar to that of sampling oceanic sediments. By using the river samples, we avoid the problem of hydrothermal contributions to sediment. Furthermore, we can calculate weighed averages, based either on the flux of the rivers or the area they drain, to come up with a more accurate estimate of crustal composition than using marine sediments. An attempt to measure the riverine Sr and Nd isotopic fluxes have been made by Goldstein and Jacobsen (1988), who measured isotopic compositions of suspended load in a subset of rivers (mainly North American) and attempted to extrapolate their results to obtain a global average (shown in Figure 16.1). They estimated the $^{87}$Sr/$^{86}$Sr and $\epsilon_{\text{Nd}}$ of the continental crust exposed to weathering as 0.716 and -10.6 respectively. However, they had no data on a number of major rivers, notably the Brahmaputra, Ganges, and Yangtze. In a related study Goldstein and Jacobsen (1987) also attempted to estimate the global average Sr and Nd isotopic composition of the dissolved load of rivers and estimated these as $^{87}$Sr/$^{86}$Sr = 0.7101 and $\epsilon_{\text{Nd}}$ = -8.4 respectively. The much lower Sr isotope ratios in the dissolved load reflects dissolving carbonate sediments. Palmer and Edmond (1989) did a more thorough job of measuring the Sr isotopic compositions of the dissolved load of rivers and obtained an average $^{87}$Sr/$^{86}$Sr of 0.7119, but did not measure the Nd isotopic composition. Given the more thorough sampling done by Palmer and Edmond, their estimate is probably more accurate. Since Goldstein and Jacobsen’s estimate of average and suspended loads are based on the same rivers sampled at the same locations, it is possible, and perhaps likely, that their estimate of the isotopic composition of the suspended load is also low. It is nevertheless the best estimate available.

As we noted above, a small but significant fraction of the Sr in rivers is in dissolved form, whereas the amount of dissolved Nd is insignificant compared to that in the suspended load. Goldstein and Jacobsen (1988) also calculated the bulk load (dissolved plus suspended) carried by rivers. Their estimate of the $^{87}$Sr/$^{86}$Sr of the bulk load was 0.7133. The lower $^{87}$Sr/$^{86}$Sr in the dissolved fraction reflects the influence of dissolving carbonates, which have lower $^{87}$Sr/$^{86}$Sr than silicate rocks because their Rb/Sr is low and seawater, from which they precipitate, is influenced by hydrothermal activity at mid-ocean ridges.

Asmeron and Jacobsen (1993) estimated the Pb isotopic composition of the crust by measuring Pb isotope ratios in the suspended load of sediments, and then estimating the global average from the correlation between Pb isotope ratios and $\epsilon_{\text{Nd}}$ in suspended loads. Their estimated composition of the upper crust exposed to weathering is $^{206}$Pb/$^{204}$Pb = 19.32, $^{207}$Pb/$^{204}$Pb = 15.76, and $^{208}$Pb/$^{204}$Pb = 39.33. This mean value is shown in Figure 16.2.

Esser and Turekian (1993) measured the Os isotopic composition of river sediments and from this estimated the average $^{187}$Os/$^{186}$Os of the continental crust exposed to weathering at 10.5 ($\gamma_{\text{Os}} = 895$). Pe-
gram et al. (1994) measured the isotopic composition of leachable Os of river sediments. The isotopic composition of the leachable fraction presumably reflects isotopic composition of dissolved Os (which is in too low a concentration to measure directly). $^{187}$Os/$^{186}$Os ranged from 10.1 to 21.5.

Using riverine suspended load eliminates the influence of hydrothermal activity on marine sediments, but the other problems with using sediments to estimate continental material remain: at best we can only estimate the composition of the crust exposed to weathering.

**EVALUATION AND GROWTH OF THE CONTINENTAL CRUST: AN EARLY PERSPECTIVE**

We now want to consider how the continental crust has evolved through time. Perhaps the first quantitative attempt to determine continental evolution rate was by Hurley and others (1962). They compiled radiometric ages (mainly Rb-Sr) of rocks in the North American continent to produce a map that looked similar to that shown in Figure 16.3. Hurley et al. recognized that age provinces in Figure 16.3 could be produced by either tectonic reworking (melting, metamorphism, etc.), of preexisting crust or new additions to crust. They also recognized they could distinguish reworked crust from new crust by initial Sr isotope ratios. They argued that because the crust has a higher Rb/Sr ratio than the mantle, new additions to crust should have lower initial $^{87}$Sr/$^{86}$Sr ratios than material produced by reworking old crust. From consideration of both age and initial isotopic composition, they argued that the crust had grown at a more or less steady rate through geological time. Their results seemed to rule out the possibility that the crust had formed at about the same time as the Earth itself: clearly new crust is created out of the mantle.

An alternative hypothesis was formulated by Armstrong in 1968. He argued that although there had been continuing additions to continental crust through geologic time, these additions were balanced by destruction and recycling of continental crust into the mantle so that there had been no net growth of the crust since very early in Earth’s history. Armstrong’s idea and that of Hurley and Rand frame the range of estimates of the rate of growth of continental crust illustrated in Figure 15.6. To this day, it has not been possible to distinguish unequivocally between these alternatives. The most likely answer, however, lies somewhere between the two extremes.

In either hypothesis, new crust has been created throughout geologic time from the mantle. To a first approximation, the crust can be considered a partial melt of the mantle. In detail, there are problems with the apparent composition of the crust, though there is universal agreement the crust has been created by magmatism. We can identify a number

![Figure 16.3. Age provinces of the North American continent (after Hurley and Rand, 1969).](image-url)
of possible mechanisms that would result in the creation of new continental crust. Most of these mechanisms suffer from the problem that they result in a more mafic crust than that observed.

- **Accretion of oceanic crust and oceanic plateaus.** The oceanic crust is generally subducted and returned to the mantle. It might in unusual situations be thrust upon or under continental crust. Subsequent melting of the basalt could produce granite. Thick oceanic plateaus produced by mantle plumes such as Iceland would be subducted less readily than normal oceanic crust. Its isotopic composition would be less depleted.

- **Underplating.** Because of the low density of the continental crust, magmas may sometimes have difficulty rising through it and may become trapped at the crust mantle boundary. This produces new basaltic lower crust, which upon melting would produce a granitic upper crust.

- **Continental volcanism.** For example, flood basalts have occasionally been erupted in tremendous volumes. Volcanism and underplating may occur simultaneously.

- **Subduction-related volcanism.** Volcanism is usually present along active continental margins. Most of the magma is of mantle derivation. Accretion of intra-oceanic island arcs to continents is a closely related mechanism.

It is clear that at present, and almost certainly throughout the Phanerozoic, the last mechanism has produced the greatest additions to continental crust. It is tempting to assume this has been the case throughout geologic time, but this has not been demonstrated unequivocally. Because subduction zones appear to play such a key role in the evolution of the crust and mantle, we will discuss them in more detail in a subsequent lecture.

**Nd Isotopic Studies of Crustal Evolution**

Nd isotope systematics provide a wonderful tool for examining the evolution of the continental crust. We have already discussed the concept of the Nd model ages, or crustal residence times. This tool will be used extensively in the subsequent discussion. We begin by examining the Western US, a region that has been studied extensively by DePaolo and coworkers.

![Figure 16.4. Isotopic provinces, based on crustal residence times (T_{DM}) of the Western U.S. (Bennett and DePaolo, 1987).](image)
Figure 16.4 is a map of the Western U.S. showing contours of crustal residence times ($\tau_{DM}$). The data define 3 distinct provinces and suggest the existence of several others. There is a general similarity to Hurley’s map, but the Nd work shows greater detail, and the ages are often older. Figure 16.5 shows the initial $\varepsilon_{Nd}$ values of the granites from the three number provinces plotted as a function of crystallization age. Only in province 3 do we find rocks, tholeiitic and calc-alkaline greenstones, whose crustal residence time is equal to their crystallization ages. In the other regions, the oldest rocks have initial $\varepsilon_{Nd}$ values that plot below the depleted mantle evolution curve. This suggests they contain significant amounts of pre-existing crust. We should emphasize at this point that the crustal residence time gives the average crustal residence time of Nd in the material. Thus if a continental rock formed at 1.0 Ga contained Nd derived in equal proportions from the mantle and 2.0 Ga crust, its crustal residence time would be 1.5 Ga. In each province there have been subsequent episodes of magmatism, but in those subsequent episodes there have been no new additions of crust (they plot along the same evolutionary array as the older material in the province).

All three provinces apparently formed between 1.8 and 1.65 Ga, though rocks from province 1 may be slightly older. A scenario suggested by DePaolo that is consistent with the observations is successive accretion or growth of island arcs to the preexisting Archean craton to the north. The earliest formed arcs, or at least those closest to the craton received a substantial component of older crust from the craton. This could have occurred through erosion and subduction, or, if the arc was built directly on the continent, through assimilation of crust. As new Proterozoic crust was built outward from the continent, it screened subsequent arcs from the contribution of material from the Archean crust. A similar effect has been observed in the Proterozoic provinces of Canada.

**ISOTOPIC COMPOSITION OF THE LOWER CRUST**

Like the mantle, the lower continental crust is not generally available for sampling. While much can be learned about the lower crust through remote geophysical means (seismic waves, gravity, heat flow, etc.), defining its composition and history depends on being able to obtain samples of it. As with the mantle, three kinds of samples are available: terranes or massifs that have been tectonically emplaced in the upper crust, xenoliths in igneous rocks, and magmas produced by partial melting of the lower crust. All these kinds of samples have been used and each has advantages and disadvantages similar to mantle samples. We will concentrate here on xenoliths and terrains.
Figure 16.6 summarizes Sr and Nd isotopic compositions of lower crustal xenoliths. Initial Sr and Nd isotopic studies of the lower crust indicated it had similar $\varepsilon_{\text{Nd}}$ to the upper crust, but low $^{87}\text{Sr}/^{86}\text{Sr}$. It is clear from Figure 16.6 that while this may be true in some instances, the lower crust is quite heterogeneous in its isotopic composition and is not easily characterized by a single isotopic composition. Some lower crustal xenoliths have very radiogenic Sr.

The Pb isotopic composition of the lower crust is a particularly important question because of the mass balance problem we discussed in the previous lectures. The upper crust, the upper mantle, and mantle plumes all have Pb isotopic compositions lying to the right of the Geochron. Mass balance requires a significant reservoir of unradiogenic Pb, i.e., Pb that plots to the left of the geochron somewhere in the Earth. Some early studies of granulite terrains, such as the Scourian in Scotland, suggested the lower crust might be characterized by unradiogenic Pb. Furthermore, the lower crust is known to have a low heat production, implying low concentrations of U and Th.

Rudnick and Goldstein (1990) found that while most Archean lower crustal terrains did indeed have very unradiogenic Pb, post-Archean ones did not. This is summarized in Figure 16.7. Furthermore, many lower crustal xenoliths have radiogenic Pb (Figure 16.8). Rudnick and Goldstein concluded that unradiogenic Pb can only develop in regions that have remained stable for long time periods, i.e., only in cratons. In areas where orogenies have occurred subsequent to crust formation, the Pb isotopic composition of the lower crust is rejuvenated through mixing with radiogenic Pb from upper crust and mantle-
Rudnick and Goldstein (1990) attempted to estimated the average Pb isotopic composition of the lower crust based on this orogenic age—Pb isotopic composition relationship. Their estimate is compared with other estimates for the Pb isotopic composition of the upper and lower crust in Figure 16.9. Rudnick and Goldstein concluded that while the Pb of the lower crust does lie to the left of the geochron, it is not sufficiently unradiogenic to balance the unradiogenic Pb of the upper crust and upper mantle. Thus the mystery of the unradiogenic Pb reservoir in the Earth remains.

**Pb Isotope Ratios and the Th/U Ratio of the Crust**

There have been a number of attempts to estimate the bulk composition of the continental crust. Isotope ratios can be used to check and refine estimates of these. Estimates of crustal composition are generally based on compilations of compositional data over wide, but nevertheless limited, areas of the continental crust. Model-based assumptions are often required to transform these limited data sets into estimates of the composition of the entire crust. For example, the well-known work of Taylor and McLennan (1985, 1995) is based largely on sampling of the Canadian Precambrian Shield and the assumption that the crust is ultimately andesitic in composition. Taylor and McLennan (1995) estimated the $^{232}$Th/$^{238}$U ratio ($\kappa$) of the crust to be 3.87, a value that is surprisingly low given the value for the bulk Earth is 4.0±0.2 and Th isotope ratios indicate that $\kappa$ in the depleted mantle is ~2.5.

Rudnick and Fountian (1995) estimated a value of 3.97, also surprisingly low, but perhaps not surprising since it was based in part of the work of Taylor and McLennan. If the mantle has lower $\kappa$ than the bulk Earth, then it stands to reason that the crust should have a high value than the bulk Earth. Furthermore, crustal rocks tend to have higher $^{208}$Pb*/$^{206}$Pb*, suggesting $\kappa$ has been lower in the mantle than in the crust for geologically long times.

Paul et al. (2003) used Asmeron and Jacobsen's (1993) estimate of the Pb composition of the riverine flux to estimate the Pb isotopic composition of upper crust and data from Rudnick and Goldstein (1990) to estimate the Pb isotope composition the lower crust. From this they calculated a $\kappa_{\text{Pb}}$ for the crust of 4.26-4.30. This clearly implies the $\kappa$ value of the crust must be higher than that estimated by...
Taylor and McLennan and Rudnick and Fountain. Indeed, when one considers that the crust is derived from the mantle, and the mantle source of crustal rocks have a \( \kappa \) equal to or lower than the bulk Earth value of 4, this implies a \( \kappa \) substantially higher than 4. Paul et al. used a mathematical model of the isotopic evolution of the Earth to estimate just how much higher. As is illustrated in Figure 16.10, the estimated that \( \kappa \) in the crust is 5.17. That value actually agrees well with an estimate of Wedepohl (1995). Notice that the model also correctly predicts a large difference between the \( \kappa_{\text{Th}} \) and \( \kappa_{\text{Pb}} \) in the depleted mantle, consistent with the observation of Galer and O’Nions (1985). The increase in \( \kappa_{\text{Pb}} \) in the depleted mantle in the model results from recycling crust Pb into the mantle through subduction.

OTHER APPROACHES TO CRUSTAL COMPOSITION AND EVOLUTION

As we have seen, samples of particulate material in rivers can be used to obtain estimates of upper crustal composition. However, because the Sm/Nd ratio changes little during production of sediment, these sediment samples also contain information on the age of the rocks they are derived from through Nd model ages (or crustal residence time). Sm/Nd and \(^{147}\text{Sm}/^{144}\text{Nd}\) ratios in major rivers draining about 25% of the exposed continental crust (excluding Antarctica and Australia) as well as samples of loess and aeolian dusts were analyzed by Goldstein and O’Nions (a different Steve Goldstein than the Steve Goldstein of the Goldstein and Jacobsen papers) are shown in Figure 16.11. The Nd isotope ratios are fairly homogeneous. Sm/Nd ratios are quite uniform, illustrating a point that was already well known, namely that rare earth patterns of continental crustal material show relatively little variation. A further illustration of this point is shown in Figure 16.12. Virtually all crustal rocks have \(^{147}\text{Sm}/^{144}\text{Nd}\) ratios at the extreme end of the range observed in mantle-derived rocks, and the range of \(^{147}\text{Sm}/^{144}\text{Nd}\) ratios in crustal material is small compared to the range observed in mantle-derived rocks. Figure 16.12 suggests there is a major fractionation of the Sm/Nd when crust is formed from the mantle, but thereafter processes within the crust tend to have only second-order effects on the Sm/Nd ratio. This is the main reason why crustal residence time calculated from \(^{147}\text{Sm}/^{144}\text{Nd}\) and \(^{143}\text{Nd}/^{144}\text{Nd}\) is such a robust parameter.

By studying sediments of various ages, we should be able to draw some inferences about the rates of continental growth. Goldstein and O’Nions (1984) found that the mean crustal residence time (\( \tau_{\text{DM}} \) calculated from \(^{147}\text{Sm}/^{144}\text{Nd}\) and \(^{143}\text{Nd}/^{144}\text{Nd}\)) of the river particulates they studied was 1.7 Ga, which they interpreted as the mean age of the crust being eroded. However, they estimated the mean crustal residence time of the entire sedimentary mass to be about 1.9 Ga. Figure 16.13 compares the stratigraphic...
age* of sediments with their crustal residence ages. Note that in general we expect the crustal residence age will be somewhat older than the stratigraphic age. Only when a rock is eroded into the sedimentary mass immediately after its derivation from the mantle will its stratigraphic ($\tau_{ST}$) and crustal residence age ($\tau_{CR}$) be equal.

The top diagram illustrates the relationships between $\tau_{ST}$ and $\tau_{CR}$ that we would expect to see for various crustal growth scenarios, assuming there is a relationship between the amount of new material added to the continents and the amount of new material added to the sedimentary mass. If the continents had been created 4.0 Ga ago and if there had been no new additions to continental crust since that time, then the crustal residence time of all sediments should be 4.0 Ga regardless of stratigraphic age, which is illustrated by the line labeled 'No New Input'. If, on the other hand, the rate of continent growth through time has been uniform since 4.0 Ga, then $\tau_{ST}$ and $\tau_{CR}$ of the sedimentary mass should lie along a line with slope of 1/2, which is the line labeled 'Uniform Rate'. The reason for this is as follows. If the sedimentary mass at any given time samples the crust in a representative fashion, then $\tau_{CR}$ of the sedimentary mass at the time of its deposition (at $\tau_{ST}$) should be $(4.0 - \tau_{ST})/2$, i.e., the mean time between the start of crustal growth (which we arbitrarily assume to be 4.0 Ga) and $\tau_{ST}$. A scenario where the rate of crustal growth decreases with time is essentially intermediate between the one-time crust creation at 4.0 and the uniform growth rate case. Therefore, we would expect the decreasing rate scenario to follow a trend intermediate between these two, for example, the line labeled 'Decreasing Rate'. On the other hand, if the rate has increased with time, the $\tau_{CR}$ of the sedimentary mass would be younger than in the case of uniform growth rate, but still must be older than $\tau_{ST}$, so this scenario should follow a path between the uniform growth rate case and the line $\tau_{ST} = \tau_{CR}$, for example, the line labeled 'Increasing Rate'.

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* The stratigraphic age is the age of deposition of the sediment determined by conventional geochronological or geological means.

† One way to rationalize this equation is to think of newly deposited sediment at $\tau_{ST}$ as a 50-50 mixture of material derived from the mantle at 4.0 Ga and $\tau_{ST}$. The equation for the $\tau_{CR}$ of this mixture would be:

$$\tau_{CR} = \frac{4.0 + \tau_{ST}}{2}.$$ 

At time of deposition, its crustal residence age would have been:

$$\tau_{CR} = \frac{4.0 + \tau_{ST}}{2} - \tau_{ST} = \frac{4.0\tau_{ST}}{2}.$$ 

You could satisfy yourself that a mixture of material having $\tau_{CR}$ of all ages between 4.0 Ga and $\tau_{ST}$ would have the same $\tau_{CR}$ as given by this equation.
Geol. 655 Isotope Geochemistry

Lecture 16

Spring 2009

Line A in Figure 16.13b is the uniform growth rate line with a slope of 1/2. Thus the data seem to be compatible with a uniform rate of growth of the continental crust. However, the situation is complicated by various forms of recycling, including sediment-to-sediment and sediment-to-crystalline rock, and crust-to-mantle. Goldstein and O’Nions noted sedimentary mass is cannibalistic: sediments are eroded and redeposited. In general, the sedimentary mass follows an exponential decay function with a half-mass stratigraphic age of about 500 Ma. This means, for example, that half the sedimentary mass was deposited within the last 500 Ma, the other half of the sedimentary mass has a depositional age of over 500 Ma. Only 25% of sediments would have a depositional (‘stratigraphic’) age older than 1000 Ma, and only 12.5% would have a stratigraphic age older than 1500 Ma, etc. Line B represents the evolution of the source of sediments for the conditions that the half-mass stratigraphic age is always 500 Ma and this age distribution is the result of erosion and re-deposition of old sediments. The line curves upward because in younger sediments consist partly of redeposited older sediments. In this process, \( \tau_{ST} \) of this cannibalized sediment changes, but \( \tau_{CR} \) does not. Goldstein and O’Nions noted their data could also be compatible with models, such as that of Armstrong, which have a near constancy of continental mass since the Archean if there was a fast but constantly decreasing rate of continent-to-mantle recycling.

We should emphasize that the \( \tau_{CR} \) of sediments is likely to be younger than the mean age of the crust. This is so because sediments preferentially sample material from topographically high areas and topographically high areas tend to be younger than older areas of the crust (e.g. the shields or cratons) because young areas tend to be still relatively hot and therefore high (due to thermal expansion of the lithosphere).

REFERENCES AND SUGGESTIONS FOR FURTHER READING


Figure 16.13. Relationship between stratigraphic age of sediments and the crustal residence age of material in sediments. See text for discussion (from Goldstein and O’Nions, 1984).