Phil. Trans. R. Soc. Lond. A **301**, 381–399 (1981)
Printed in Great Britain

The composition and evolution of the continental crust: rare earth element evidence from sedimentary rocks

By S. R. Taylor and S. M. McLennan
Research School of Earth Sciences, Australian National University, Canberra, 2600 *Australia*

The composition of the present-day upper crust, inferred from the uniformity of sedimentary rock r.e.e. (rare earth element) patterns, is close to that of granodiorite. A revised 'andesite' model is used to obtain total crustal composition. The lower crust is the composition remaining, assuming that the upper crust, one-third of the total, is derived from intracrustal partial melting. The upper-crustal r.e.e. pattern has pronounced Eu depletion (Eu/Eu* = 0.64), the lower-crustal pattern has Eu enrichment (Eu/Eu* = 1.17) and the total crust has no Eu anomaly relative to chondritic abundances. The Eu depletion in the upper crust is attributed to retention of Eu in plagioclase in the lower crust. Because plagioclase is not stable below 40 km (> 10 kbar), the anomaly is intracrustal in origin.

The Archaean upper crust has a different r.e.e. pattern to that of the present-day upper crust, being lower in total r.e.e., and La/Yb ratios, and lacking an Eu anomaly. These data are used to infer the Archaean upper-crustal composition, which resembles that of the present-day total crust, except that Ni and Cr contents are higher. The Archaean crustal composition can be modelled by a mixture of tholeiites and tonalite trondhjemites. The latter have steep light r.e.e.-enriched – heavy r.e.e.-depleted patterns, consistent with equilibration with garnet and hence probable mantle derivation. There is little reason to suppose that the Archaean lower crust was different in composition from the upper crust, except locally where partial melting episodes occurred. The r.e.e. evidence is consistent with isotopic and geological evidence for a low continental growth rate in the early Archaean, a massive increase (to about 70% of the total crust) between about 3000 and 2500 Ma B.P. and a slow increase until the present day. The change from Archaean to post-Archaean r.e.e. patterns in the upper crust is not isochronous, but is reflected in the sedimentary rock r.e.e. patterns at differing times in different continents. On the basis of a model composition for the mantle, 36% of the potassium, 30% of uranium, 15% of lanthanum and 3% of ytterbium are concentrated in the present continental crust. This enrichment is related to ionic size and valency differences from common mantle cations (e.g. Mg, Fe). Pre-3.9 Ga B.P. crusts were obliterated by meteorite bombardment. No geochemical evidence exists for primordial anorthositic, sialic or mafic crusts.

1. RARE EARTH ELEMENTS AND CRUSTAL AVERAGES

The formation of sedimentary rocks, involving weathering, erosion, transportation, deposition and diagenesis, provides a sampling of the exposed continental crust. How representative this sampling is depends both on the type of sediment and on the behaviour of the chemical elements during its formation. Thus readily soluble elements, being concentrated in the oceans and in evaporite deposits, will not record their crustal abundances in any simple fashion. Elements with low solubility and mobility, and with short residence times in sea water, will be transferred quantitatively into detrital sediments and preserve a record of the source rock compositions. The rare earth elements (r.e.e.) (La–Lu) appear to meet those latter conditions. The uniformity of their abundance patterns in clastic sedimentary rocks argues both for the widespread sampling being carried out by sedimentary processes, and for the inherent stability of the r.e.e. during those events.

The r.e.e. have negligible residence times in sea water (50–600 a (Goldberg *et al.* 1973; Piper 1974)). Haskin & Paster (1979) note that the r.e.e. concentration in the oceans 'is less than the amount present in the upper 0.2 mm of ocean floor sediment'. Accordingly, the distribution of the r.e.e. in sedimentary rocks is essentially unaffected by the marine environment. Other studies demonstrate the immobility of the r.e.e. during diagenesis and metamorphism (see, for example, Muecke *et al.* 1979), except in some examples occurring during low-grade metamorphism of basalts (see, for example, Floyd 1977).

The uniformity of r.e.e. patterns in sedimentary rocks was first used by Taylor (1964) to derive an element abundance table for the continental crust. Since that time, the data have improved both in quality and quantity. Thus the r.e.e. abundance patterns in post-Archaean shales have $\mathrm{La_N/Yb_N}$ ratios of 9.2 ± 1.5 (where N stands for chondrite normalized) and $\mathrm{Eu/Eu^*}$ values of 0.64 ± 0.05 (Nance & Taylor 1976; Taylor & McLennan 1981*a*). ($\mathrm{Eu^*}$ is the theoretical Eu concentration for no Eu enrichment or depletion relative to neighbouring r.e.e.). This uniformity, which occurs in all clastic sedimentary rocks except for some first-cycle sediments (Nance & Taylor 1977), is interpreted to indicate that the processes of formation of clastic sediments are carrying out a widespread and average sampling of the upper crust, which preserve and record its average rare earth abundance pattern.

2. Composition of the present upper crust

In discussing crustal abundances, it is necessary to distinguish the upper crust (10–15 km thick), exposed to weathering and erosion, from both the generally inaccessible lower crust, and the total crust (typically 40 km thick) above Mohorovičić discontinuity. Heat-flow data and element balance calculations combine to indicate that the observed upper crustal abundances, for example of K, U and Th, cannot extend to the base of the crust.

The sedimentary rocks provide a measure of the r.e.e. abundance patterns, but the wide dispersal of many other elements (e.g. Rb) during their formation does not make them a suitable base from which to calculate crustal compositions. Even the use, by Goldschmidt (1954), of glacial clays derived from mechanically abraded rock flour produces compositions depleted in Na and Ca among the major elements. In the models used here, the assumption is made that the crust is ultimately derived from the mantle, and that the upper crust is derived from the lower crust by igneous processes. Accordingly, the upper-crustal r.e.e. *patterns*, deduced from sedimentary rocks, represent those of the igneous source material. The r.e.e. upper-crustal patterns are equivalent to those observed in granodiorites. This assumption is confirmed from direct sampling of upper-crustal rocks (Eade & Fahrig 1971, 1973; Shaw *et al.* 1967, 1976), which produce 'granodioritic' average compositions. The upper-crustal composition derived in this manner is given in table 1. It is distinguished by high concentrations of the large-ion lithophile (l.i.l.) elements and is characterized by a marked Eu depletion.

Three observations combine to suggest that the absolute upper-crustal abundances of the r.e.e. are about 20% lower than in the post-Archaean average Australian sedimentary rock (p.A.a.s.): (*a*) upper crustal estimates based on sampling of igneous rocks give average La abundances of about 30 μg/g (Shaw *et al.* 1976); (*b*) fine-grained sediments such as shales contain about 20–30% more r.e.e. than do coarser clastic sediments of the same major-element chemistry; and (*c*) clastics constitute about 70–80% of sedimentary rocks at the surface of the crust, but contain the bulk of the r.e.e., carbonates and evaporites having very low abundances.

Accordingly, the post-Archaean sedimentary r.e.e. abundances have been reduced by 20% (e.g. La from 38 to 30 μg/g) to give the upper-crustal values. The relative patterns remain parallel. The total r.e.e. in the upper crust becomes 148 μg/g. With the exception of this change in the r.e.e. abundances, the other values are unchanged from those of Taylor (1977, 1979*a*).

3. MODELS FOR TOTAL-CRUSTAL COMPOSITION

The upper-crustal composition has a reasonably established observational base and provides one boundary for total-crustal compositions. The derivation of the continental crust from various sources was discussed by Taylor (1967) who concluded that island-arc volcanism provided the only presently observable viable source, both from volume and compositional considerations. The average composition of present-day island-arc rocks was accordingly used, on uniformitarian grounds, to establish the bulk composition of the continental crust and has become labelled as the 'andesite' model. Although considerable progress has been made since that time, only slight modifications to the crustal averages appear to be needed at present (tables 1, 2).

The average composition of erupted material at island arcs has been much debated, but consensus appears to be emerging, based on statistical treatment, that the average silica content, for example, is about 57–58% SiO_2 (Ewart 1976). The Th/U ratio used in previous estimates (Taylor 1977, 1979*a*) was based heavily on that of intraoceanic arcs. These modern arcs may be sampling somewhat depleted mantle, since their Th/U ratios are similar to those of mid-ocean ridge basalt (m.o.r.b.). Many calc-alkaline volcanic rocks in continental or continental margin arcs have Th/U close to 3.8 (e.g. in northeast Japan (Masuda & Aoki 1979)). Accordingly the Th/U ratio has been revised to 3.8. The U abundance has been raised from 1.0 to 1.25 μg/g to be consistent with an overall K/U ratio of 10^4. These new values provide a total crustal heat flow of 0.032 W/m^2 (0.76 heat flow units (h.f.u.)) for a crust 40 km thick. For a total continental heat flow of 0.053 W/m^2 (1.2 h.f.u.) (Lee 1970) the average mantle-derived heatflux accordingly is 0.0183 W/m^2 (0.44 h.f.u.) for a crust 40 km thick, or 0.0263 W/m^2 (0.63 h.f.u.) for a crust 30 km thick (M. Harrison, personal communication). Rb has been lowered to 42 μg/g to give K/Rb = 300. The model composition is otherwise unchanged from that of Taylor (1977, 1979*a*).

4. PRESENT LOWER-CRUSTAL COMPOSITION

A model for overall lower-crustal composition may be calculated from the total-crustal composition by subtracting the observed upper-crustal composition. The results of this calculation are given in table 1, based on the assumption that the upper crust constitutes one-third of the total. This calculation is based also on the model that the upper crust is derived by intracrustal partial melting (see § 9). Among the characteristics of the lower-crustal composition modelled here, is the presence of enrichment in Eu compared to the neighbouring r.e.e. The light r.e.e. still display some enrichment relative to chondrites (figure 1).

Table 1. Elemental abundances for C-1 carbonaceous chondrites (volatile-free), the primitive Earth mantle, present Earth continental crust, present upper and lower continental crusts and the Archaean upper crust

element	C-1 volatile-free µg/g	source†	primitive Earth mantle µg/g	source†	present Earth crust (25) µg/g	percentage in crust	present upper crust (25) µg/g	present lower crust (25) µg/g	Archaean upper crust µg/g	source†
Li	2.4	2	0.78	24	10	5.5	—	—	—	—
Be	0.072	4	0.10	15	1.5	6.6	—	—	—	—
B	2.4	3	3.3	15	—	—	—	—	—	—
Na	7900	1	2500	21	26000	5.0	28200	25200	23000	26
Mg	141000	1	240000	14	21100	0.038	13900	24700	31400	26
Al	12900	1	17500	14	95000	2.3	84700	100000	82500	26
Si	156000	1	210000	14	271000	0.56	308000	252000	268000	26
K	890	2	180	19	12500	35.7	27400	5000	7500	26
Ca	13900	1	18900	14	53600	1.22	25000	67900	52200	26
Sc	7.8	2	10.6	15	30	1.22	10	40	25	29
Ti	660	5	900	14	4800	2.3	3600	5400	5400	26
V	62	2	84	17	175	0.90	60	230	150	29
Cr	3500	1	3000	14	55	0.007	35	65	140	26
Mn	2700	1	1000	18	1100	0.47	600	1350	1300	26
Fe	272000	1	62000	14	58300	0.41	35000	70000	73800	26
Co	765	6	100	16	25	0.12	10	33	30	26
Ni	15100	1	2000	14	30	0.007	20	35	90	29
Cu	160	2	28	30	60	0.92	25	78	80	26
Zn	455	2	50	30	—	—	52	—	100	26
Ga	14	6	3	30	18	2.6	—	—	—	—
Rb	3.45	2	0.48	20	42	38.0	110	8	25	28
Sr	11.4	2	15.5	15	400	11.0	350	425	300	29
Y	2.1	2	2.9	15	22	3.4	22	22	15	27
Zr	5.7	5	7.8	15	100	5.5	240	30	100	29
Nb	0.45	7	0.60	15	11	7.9	25	4	5	29
Cs	0.29	8	<0.016	22	1.7	>45.0	3.7	0.7	—	—
Ba	3.6	2	4.9	15	350	30	700	175	240	26
La	0.367	9	0.50	15	19	15.4	30	14	12.6	27
Ce	0.957	9	1.30	15	38	12.6	64	25	26.8	27
Pr	0.137	10	0.19	15	4.3	9.8	7.1	2.9	3.1	27
Nd	0.711	9	0.967	15	16	7.1	26	11	13.0	27
Sm	0.231	9	0.314	15	3.7	5.1	4.5	3.3	2.78	27
Eu	0.087	9	0.12	15	1.1	4.0	0.88	1.2	0.90	27
Gd	0.306	9	0.42	15	3.6	3.7	3.8	3.5	2.85	27
Tb	0.058	10	0.079	15	0.64	3.5	0.64	0.64	0.48	27
Dy	0.381	9	0.52	15	3.7	3.0	3.5	3.8	2.93	27
Ho	0.085	9	0.12	15	0.82	2.9	0.80	0.83	0.63	27
Er	0.249	9	0.34	15	2.3	2.9	2.3	2.3	1.81	27
Tm	0.036	10	0.048	15	0.32	2.9	0.33	0.32	0.26	27
Yb	0.248	9	0.34	15	2.2	2.8	2.2	2.2	1.79	27
Lu	0.038	9	0.052	15	0.30	2.4	0.32	0.29	0.28	27
Hf	0.17	11	0.23	15	3.0	5.6	5.8	1.6	3	28
Tl	0.22	8	0.005	—	—	—	0.5	—	—	—
Pb	3.6	2	—	—	10	—	15	7.5	10	29
Bi	0.17	8	—	—	—	—	—	—	—	—
Th	0.051	12	0.070	15	4.8	30	10.5	1.95	2.9	28
U	0.014	8	0.018	15	1.25	30	2.5	0.63	0.75	28

TABLE 1—*cont.*

† *Notes on table 1*

source

1 Wiik (1971). Data multiplied by 1.479 to allow for loss of H_2O, C, S.

2 Mason (1979). Date for Orgueil from table 84, pp. B111–B112. Values multiplied by 1.5 to allow for volatile loss.

3 Weller *et al.* (1978). Data multiplied by 1.5.

4 From Li/Be = 33 (Dreibus *et al.* 1976).

5 Shima (1979). Data multiplied by 1.5.

6 Mason (1979). Average C-1 value (× 1.5).

7 Mason (1979). Ivuna data (× 1.5).

8 Krahenbuhl *et al.* (1973). Data multiplied by 1.5.

9 Evensen *et al.* (1978). Mean C-1 values multiplied by 1.5.

10 Extrapolated value from neighbouring r.e.e.

11 From Zr/Hf = 33.

12 From Th/U = 3.8.

13 J. H. Crocket, personal communication (1977).

14 Taylor (1979*c*). Values based mainly on nodule data. Ratio of Al, Ca and Ti to volatile-free chondritic values (column 1) is 1.36. This ratio is used to obtained mantle abundances for refractory elements from the chondritic data, with the assumption of no relative fractionation.

15 Refractory trace element value from col. 1 of Taylor (1979*c*), multiplied by 1.36 (see note 14).

16 From Ni/Co = 20 (C-1 value).

17 C-1 value multiplied by 1.36.

18 From FeO/MnO = 60 (Dreibus 1977).

19 K from K–Ar (Taylor 1979*c*) and K/U = 10^4.

20 Rb from Rb/Sr = 0.031 (O'Nions *et al.* 1979). This gives K/Rb = 375. This implies loss of Rb relative to K during accretion of Earth since K/Rb for C-1 is 260 (see note 25).

source

21 From Na/K = 14 (Taylor 1980).

22 Rb/Cs in crust is *ca.* 30 compared to 12 in chondrites. Cs will be enriched in crust relative to Rb. Thus 0.016 µg/g (from Rb/Cs = 30) is an upper limit.

23 From Tl/U = 0.022 (Krahenbuhl *et al.* 1973).

24 From Li/Zr = 0.10 (Dreibus *et al.* 1977).

25 Total crustal data based on 'andesite model' from Taylor (1979*a*). The uranium abundance has been revised from 1.0 to 1.25 µg/g by means of K/U = 10^4. Th has been raised to 4.8 µg/g to conform to Th/U = 3.8. Hf has been revised to 3.0 µg/g (from 2.2 µg/g) to be consistent with chondritic Zr/Hf = 33. Tl from Shaw *et al.* (1976). R.e.e. data have been lowered by 20% from those quoted by Taylor (1977, 1979*a*). See text for discussion. Rb has been lowered from 50 (Taylor 1977) to 42 µg/g to be consistent with K/Rb = 300. This ratio implies that Rb is partitioned into the crust preferentially to K (cf. Cs, source note 22).

26 Major element data, Ba, Co, Cu and Zn from Goodwin (1977).

27 R.e.e. from average Archaean sedimentary rock (a.A.s.) (Taylor & McLennan 1981*a*).

28 Rb from K/Rb = 300, U from K/U = 10^4, Th from Th/U = 3.8, Hf from Zr/Hf = 33.

29 Sr, Zr, V, Ni from tholeiite/tonalite–trondhjemite mix (50:50). Tholeiite data from Sun & Nesbitt (1977, 1978). Tonalite and trondhjemite data from Glikson (1978).

30 H. Palme (personal communication).

TABLE 2. MAJOR-ELEMENT COMPOSITIONS FOR C-1 CARBONACEOUS CHONDRITES (VOLATILE-FREE), PRIMITIVE EARTH MANTLE, PRESENT EARTH CONTINENTAL CRUST, PRESENT UPPER AND LOWER CONTINENTAL CRUSTS AND THE ARCHAEAN UPPER CONTINENTAL CRUST

(Data are percentages (by mass). Sources given in notes following table 1.)

	C-1 volatile-free	primitive Earth mantle	present Earth crust	present upper crust	present lower crust	Archaean upper crust
SiO_2	33.3	45.0	58.0	66.0	54.0	57.4
TiO_2	0.11	0.15	0.8	0.6	0.9	0.9
Al_2O_3	2.43	3.3	18.0	16.0	19.0	15.6
FeO	35.0	8.0	7.5	4.5	9.0	9.5
MgO	23.4	40.0	3.5	2.3	4.1	5.2
CaO	1.94	2.65	7.5	3.5	9.5	7.3
Na_2O	1.05	0.34	3.5	3.8	3.43	3.1
K_2O	0.11	0.02	1.5	3.3	0.6	0.9
NiO	1.92	0.25	—	—	—	—
Cr_2O_3	0.52	0.44	—	—	—	—
MnO	0.34	0.13	—	—	—	—
Σ	100.12	100.1	100.3	100.0	100.4	99.9

The general problem of arriving at estimates of lower-crustal compositions is severe. No average sample analogous to that produced for the r.e.e. by sedimentary processes appears to be available. The available material from high-grade granulite terrains (either exposed, or from xenoliths) might be atypical, similar to random sampling of upper-crustal rocks, which would provide a wide diversity of compositions.

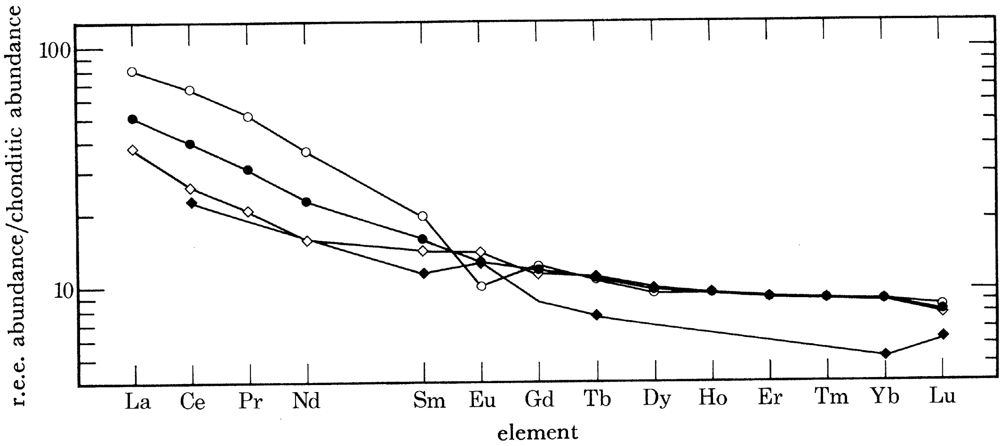

FIGURE 1. R.e.e. patterns, normalized to chondritic abundances (C-1, volatile-free) for the upper, lower and total continental crust. Data from table 1; ○, upper crust; ●, total crust; ◇, lower crust; ◆, Scourian terrain and Lesotho xenoliths. The upper-crustal patterns are parallel to those of p.A.a.s. (post-Archaean sedimentary rock), but the abolute abundances are 20 % lower (see text). The lower crust is assumed to be two-thirds of the total crust. Note relative Eu enrichment in the lower-crustal model composition. The average Scourian terrain r.e.e. patterns (Muecke *et al.* 1979) and those from Lesotho xenoliths (Rogers 1977) closely conform to the predicted lower-crustal patterns. Note that Sm/Nd ratios in the lower crust are close to chondritic values. (Reproduced from Taylor & McLennan (1981*a*).)

The present model provides some testable predictions. *Average* lower-crustal samples should have positive Eu anomalies. The calculated Sm/Nd average ratio is 0.30, similar to chondritic values of 0.325, much higher than the upper-crustal average of 0.18. The estimated lower-crustal value is thus not readily distinguishable from chondritic or undepleted mantle values, if we allow for the uncertainties in the calculations. The normative composition for the lower crust has about 60 % plagioclase (Taylor 1977), consistent with the high Ca, Al, Sr and Eu contents.

Samples that display some characteristics of the model proposed here may be found both in xenoliths from Lesotho kimberlites (Rogers 1977) and in the exposed Lewisian gneisses (Muecke *et al.* 1979). The well studied Scourian terrain of the Archaean Lewisian gneiss complex (Holland & Lambert 1975) is enriched in Eu relative to the other r.e.e. (Muecke *et al.* 1979). The r.e.e. patterns of the constituent minerals of the granulites indicate equilibration with a granodioritic liquid, consistent with the model adopted here. The age relationships of these events (Hamilton *et al.* 1979) will be discussed in a later section.

5. UNIFORMITY OF CRUSTAL COMPOSITION WITH TIME

The r.e.e. patterns in sedimentary rocks are uniform back to the early Proterozoic. The Eu depletion, La/Yb and Σl.r.e.e./Σh.r.e.e. ratios are uniform within analytical error over this immense span of time (Taylor 1979*a*). When rocks of similar major-element composition are considered, there is some suggestion of a decrease in total r.e.e. with age (Nance & Taylor 1976). Since the r.e.e. patterns maintain a uniform Σl.r.e.e./Σh.r.e.e. ratio, this effect is probably mainly due to recycling of sediments (Veizer 1973; Veizer & Jansen 1979). Igneous processes would alter the slopes of the patterns. Accordingly, one could apply Lyellian principles of uniformitarianism over this period, and extrapolate present-day models for continental growth back to the base of the Proterozoic.

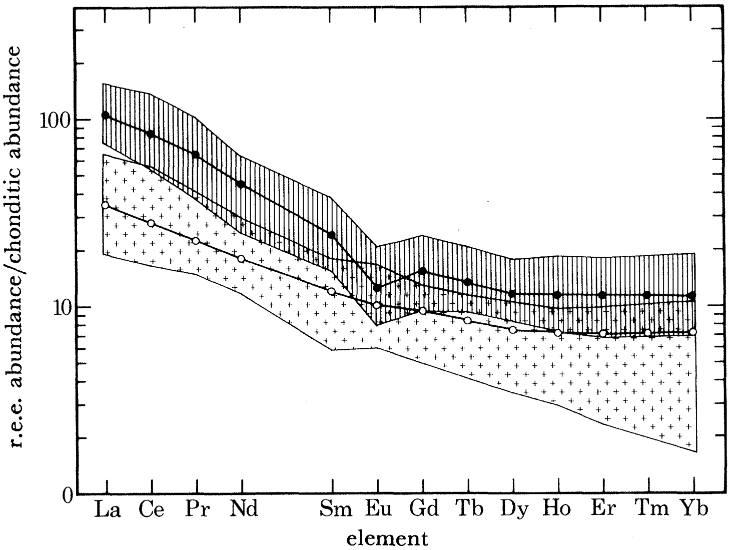

FIGURE 2. R.e.e. patterns, normalized to chondritic abundances (C-1, volatile-free), for the post-Archaean average Australian sedimentary rocks (p.A.a.s., ●), and for the average Archaean sedimentary rocks a.A.s., ○). The fields for individual samples of typical Archaean (a.A.s., +) and post-Archaean (p.A.a.s.||||) sedimentary rocks are also shown. Data from Nance & Taylor (1976, 1977), Bavington & Taylor (1980), from this laboratory. (Reproduced from Taylor & McLennan (1980*a*).)

6. R.E.E. PATTERNS IN ARCHAEAN SEDIMENTARY ROCKS

Typical r.e.e. patterns in Archaean sedimentary rocks differ in three significant respects from those in post-Archaean sedimentary rocks. They have lower La/Yb ratios (average $\mathrm{La_N/Yb_N} = 4.6$), no detectable Eu anomaly ($\mathrm{Eu/Eu^*} \approx 1$) and lower total r.e.e. abundances (Σr.e.e. = 70, compared to 185 for p.A.a.s.). They have somewhat more variable abundance patterns (e.g. La/Yb) than do post-Archaean sediments. Rarely, Eu enrichments are observed, due either to local plagioclase accumulation (Nance & Taylor 1977) or to local hydrothermal effects (Kerrich & Fryer 1979; Cullers *et al.* 1973; Bavinton & Taylor 1980). Some Archaean sedimentary rocks have steep l.r.e.e.-enriched patterns indicating local derivation from acidic parent rocks (e.g. at Kalgoorlie (Nance & Taylor 1977)). Examples of flat r.e.e. patterns, indicating derivation from basaltic parents, occur in the Alkilia and Malene supracrustal rocks (unpublished data from the authors' laboratory). Most Archaean sedimentary rocks, however, exhibit patterns with the characteristics noted earlier. The average pattern so calculated is given in figure 2. Some questions must be addressed before this average Archaean sedimentary pattern can be used, in the same fashion as that for post-Archaean sedimentary rocks, to deduce the average composition of the exposed Archaean crust.

(*a*) Do the Archaean sedimentary rocks represent a widespread sampling of the exposed crust? Although localized exceptions occur, as noted above, the uniformity of patterns encountered in Archaean sediments is impressive. Similar r.e.e. patterns occur at Kalgoorlie (Nance & Taylor 1977) and Kambalda (Bavington & Taylor 1980) in Australia, Wind River in Wyoming, U.S.A., Barberton Mountain Land in South Africa (Wildeman & Condie 1973; Wildeman & Haskin 1973), Isua, Alkilia and Malene supracrustals in Greenland (Mason 1975); unpublished data from the authors' laboratory), Yellowknife (Jenner *et al.* 1981) and Knife Lake (Arth & Hanson 1975) and the Abitibi greenstone belt (Kerrich & Fryer 1979) in Canada. The similar characteristics of all these patterns indicate that sedimentary processes in the Archaean are homogenizing the r.e.e. abundances. This uniformity of patterns extends to very small occurrences of sedimentary rocks, where local derivation might have been expected. Thus the sedimentary lenses at Kambalda (Bavinton & Taylor 1980), enclosed in basalt, and rarely more than a few hundred metres in lateral extent, display patterns indistinguishable from those of the other Archaean localities.

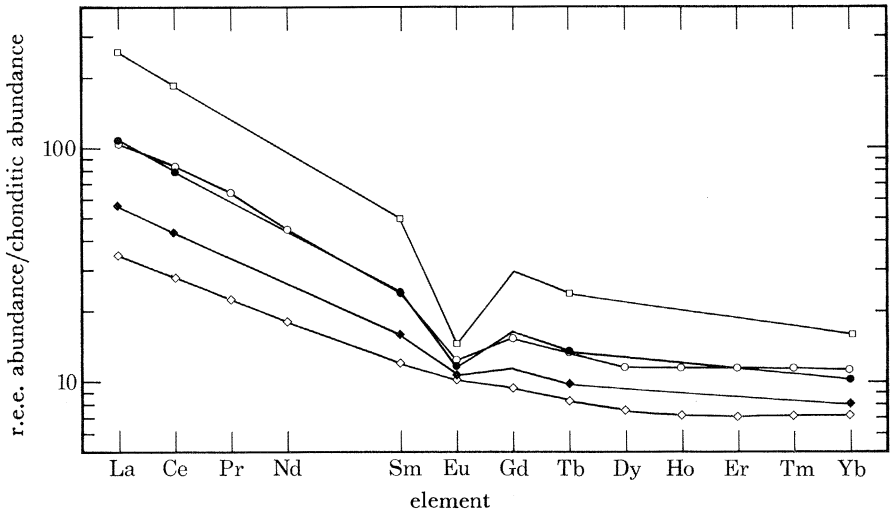

FIGURE 3. R.e.e. patterns, normalized to chondritic abundances (C-1, volatile-free) for p.A.a.s., a.A.s. and 36 late Archaean K-rich granites (Taylor & McLennan 1981*a*); ○, p.A.a.s.; ◇, a.A.s.; □, K granite; ●, 2 a.A.s.: 1 K granite mixture; ◆, 9 a.A.s.:1 K granite mixture. Note that even 10% of K-rich granite patterns mixed with a.A.s. produces a discernible Eu depletion. This indicates that K-rich granites are not a significant source for average Archaean sedimentary rocks. A mix of two-thirds a.A.s. and one-third granite produces r.e.e. patterns similar to p.A.a.s. (Reproduced from Taylor & McLennan (1981*a*).)

(*b*) Is the sampling biased toward sedimentary rocks in greenstone belts? Could the uniformity claimed in the preceding section be due to biased sampling? Many Archaean sedimentary rocks are located in greenstone belts, but several observations indicate that an overall sampling of the Archaean terrain is being accomplished. These include the following.

(i) The volcanic rocks in greenstone belts have flat r.e.e. patterns (Sun & Nesbitt 1977). They have only about a 10% felsic component, insufficient to account for the observed sedimentary r.e.e. patterns (Goodwin 1977, fig. 2).

(ii) Petrographic evidence in the Kambalda sediments indicates the presence of 'granitic' components (Bavinton & Taylor 1980), exotic to the immediate environment.

(iii) Steep r.e.e. patterns derived wholly from tonalitic–trondhjemitic debris are rare. Most of this material is mixed with basaltic debris to provide the observed patterns.

(iv) The lowermost Proterozoic Huronian sediments, derived from a complex provenance, have patterns similar to those of typical Archean sediments (McLennan *et al.* 1979).

Accordingly, it appears that the sedimentary sequences are sampling both the greenstone belt and the 'granitic' or felsic Archaean terrains. The r.e.e. patterns are consistent with about equal contributions from both sources (Nance & Taylor 1977; Taylor 1977). There is no discernible input into the Archaean sedimentary record from typical K-rich granites or granodiorites with negative Eu anomalies. Figure 3 shows the effects of such additions to the Archaean pattern. If more than 10% granite is added, a detectable Eu depletion will appear in the sedimentary record. Accordingly we judge that such contributions were less than that value, although occasional examples of granodiorites with Eu depletion occur (e.g. at Amîtsoq (Mason 1975)).

7. ARCHAEAN UPPER CRUSTAL COMPOSITIONS

The philosophy of deriving crustal estimates from r.e.e. patterns in sedimentary rocks was noted earlier. The sedimentary r.e.e. patterns provide the crustal r.e.e. pattern. This is in fact the average pattern of the igneous rocks constituting the crust. These can be identified from the r.e.e. pattern, enabling average major- and trace-element abundances to be obtained. The Archaean average sedimentary rock pattern derived in the previous section is close both to average island-arc compositions and to mixture of tholeiite and tonalite–trondhjemite suites. An appropriate major-element composition is that derived by Goodwin (1977) for the average volcanic rock of the Superior Province.

The Archaean sedimentary rock record is probably dominated by clastics, consistent with the rapid erosion apparent in such terrains (Walker 1978). Chemical sediments such as iron formations and carbonates make up an unknown but probably small amount of the total sedimentary record. Accordingly, the r.e.e. abundances in the Archaean sediments may be taken as indicative of the *abundances* as well as the relative r.e.e. patterns of the Archaean crust. The suggested composition is given in table 1 from Taylor & McLennan (1981*b*).

8. COMPARISON OF ARCHAEAN AND POST-ARCHAEAN CRUSTAL COMPOSITIONS (figure 4)

The composition of the exposed Archaean upper crust derived in the previous section and listed in table 1 resembles that of the total post-Archaean crust, except that it contains more Ni and Cr. It is thus much less evolved geochemically than that of the post-Archaean *upper* crust. No overall Eu depletion is present, so that the evidence that is widespread for intracrustal melting in the post-Archaean crust is absent in the Archaean. The occasional granodiorites and felsic volcanics with negative Eu anomalies, as noted earlier, do not contribute significantly to the sedimentary record of upper crustal composition. The lack of evidence of much intracrustal melting indicates that the composition of the unexposed portions of the Archaean crust were probably similar to that of the eroding crust. Accordingly there is no reason to postulate a separate lower crustal composition in the Archaean. The existence of the sedimentary rocks themselves and the presence of distantly derived 'granitic' detritus at Kambalda, for example, indicate that substantial areas of crust were above sea level (cf. Hargraves 1976).

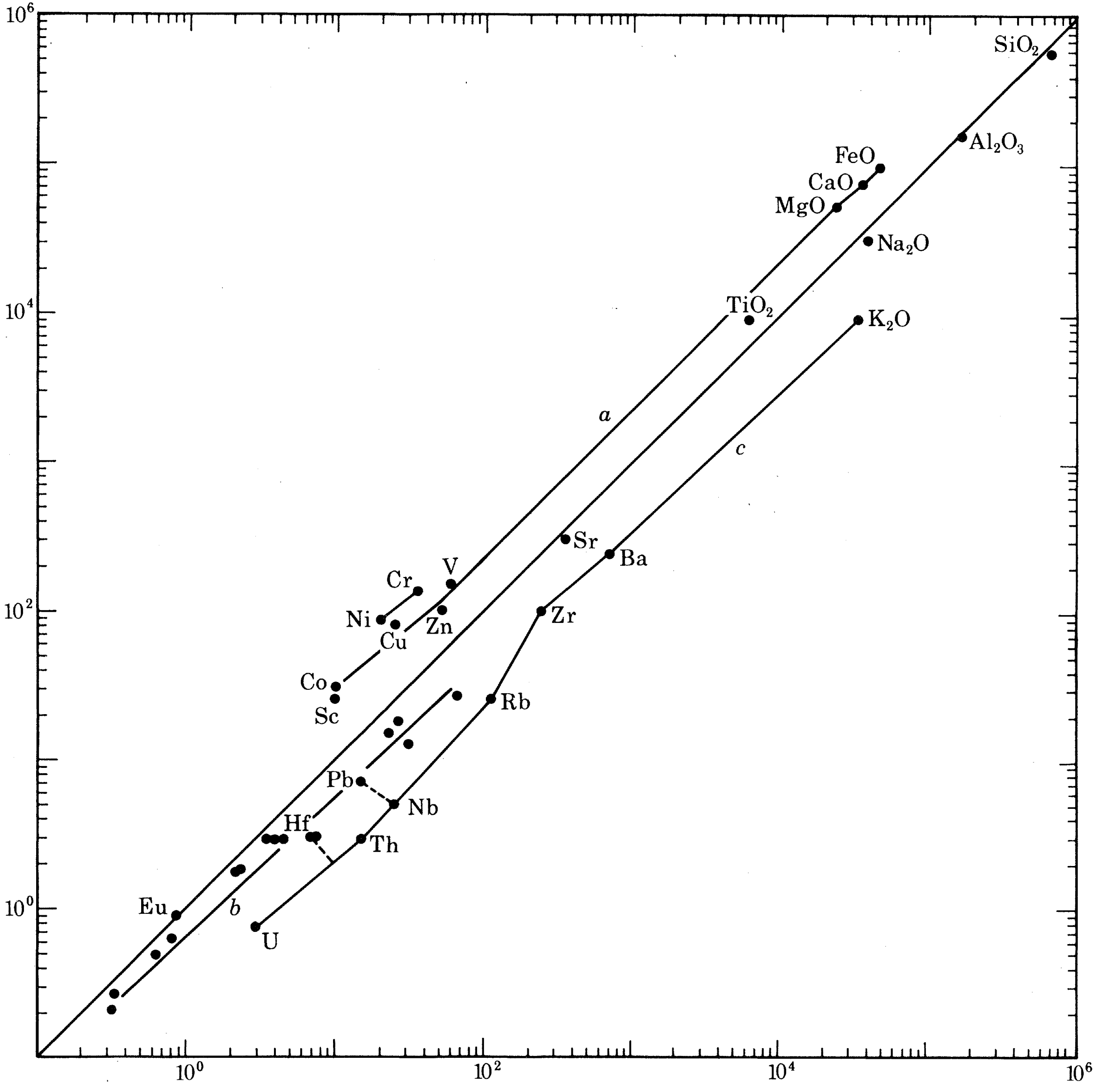

FIGURE 4. The relative compositions of the Archaean and post-Archaean upper continental crusts. Note the enrichment of ferromagnesian elements (line *a*) and the depletion of l.i.l. elements (line *c*) in the Archaean upper crust. Rare earth elements, line *b*.

9. CAUSES OF THE COMPOSITIONAL CHANGE IN THE UPPER CRUST

The change in the composition of the post-Archaean upper crust, as documented by the r.e.e. patterns, is inferred to be a consequence of massive intracrustal partial melting, which forms magmas predominantly of granodioritic composition. These rise into and intrude the upper crust, principally as batholiths (Fyfe 1973*a*, *b*). During this intracrustal melting event the melt becomes enriched in the light r.e.e. and depleted in Eu. The depletion in Eu is attributed to the differing crystal–liquid partition behaviour of divalent compared to the other trivalent r.e.e. Eu^{2+}, of nearly the same radius as Sr^{2+}, is preferentially incorporated in feldspar, particularly plagioclase. Charge-balance difficulties and smaller radii exclude the trivalent r.e.e. and concentrate them in the melt. Thus granodioritic melts forming from material of island-arc composition within the stability field of plagioclase (< 10 kbar; < 40 km) will develop a negative Eu anomaly.

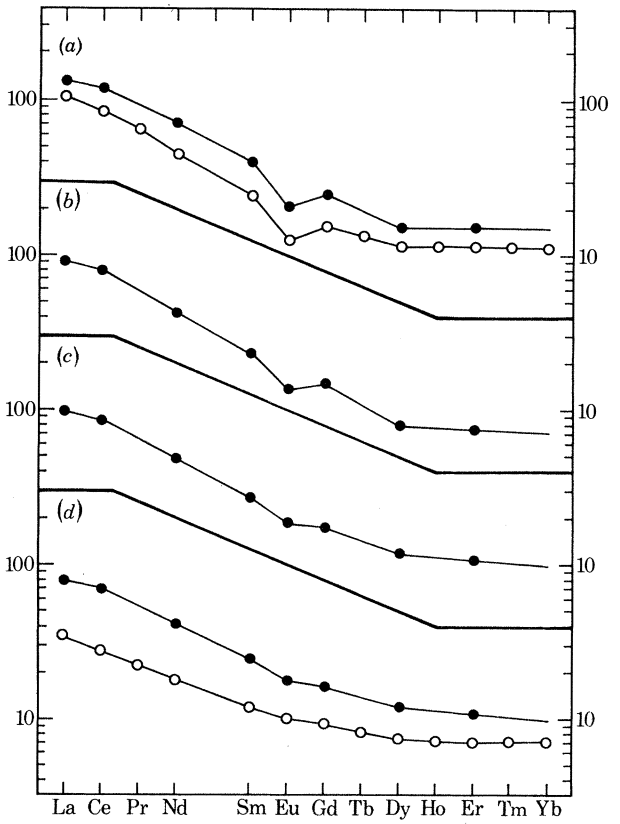

FIGURE 5. The change in r.e.e. patterns (●) in the Huronian sequence. The McKim Formation (*d*) has patterns close to those of a.A.s. (○, Eu/Eu* = 1.0; shown for comparison) but with slight Eu depletion (Eu/Eu* = 0.83) higher La_N/Yb_N and Σr.e.e. The Eu depletion gradually increases through the Pecors ((*c*), Eu/Eu* = 0.81) and the Serpent ((*b*), Eu/Eu* = 0.72 until in the Gordon Lake Formation ((*a*)Eu,/Eu* = 0.64), near the top of the sequence, the Eu depletion equals that of p.A.a.s. (○, shown in (*a*) for comparison) (adapted from McLennan *et al.* (1979)).

The sedimentary sampling of the upper granodioritic crust inherits this Eu depletion, even though the europium may be oxidized to Eu^{3+} during weathering. An important point is that the development of the upper-crustal Eu depletion is *intracrustal* in origin. A principal observation in support of this is that no systematic enrichment or depletion in Eu is observed in igneous rocks of undoubted mantle origin.

The r.e.e. evidence for this major episodic change is supported by the Sm–Nd isotopic systematics of crustal rocks, which indicate a massive increase in crustal growth at that time (McCulloch & Wasserburg 1978). Veizer & Jansen (1979) note the $^{87}Sr/^{86}Sr$ ratio in carbonates (reflecting seawater composition) is rather low in the Archaean but shows a major increase ($\Delta ^{87}Sr/^{86}Sr) = 0.0025$) over the time interval from about 2.5 to 2.0 Ga, with a more gradual increase up to the present. These trends are consistent with a large addition of Rb to the upper-crustal weathering régime between 3 and 2 Ga. The change in upper-crustal composition is not expected to be isochronous on a world-wide scale, but to occur at somewhat different times (within 0.5 Ga) on different continents. The change in the r.e.e. patterns in sedimentary rocks also has to await the unroofing of the granodioritic batholiths.

The predictable change from Archaean to post-Archaean r.e.e. patterns has been demonstrated by McLennan *et al.* (1979) in the lowermost Proterozoic Huronian succession in Canada. R.e.e. patterns at the base (McKim Formation) have Archaean-type patterns.

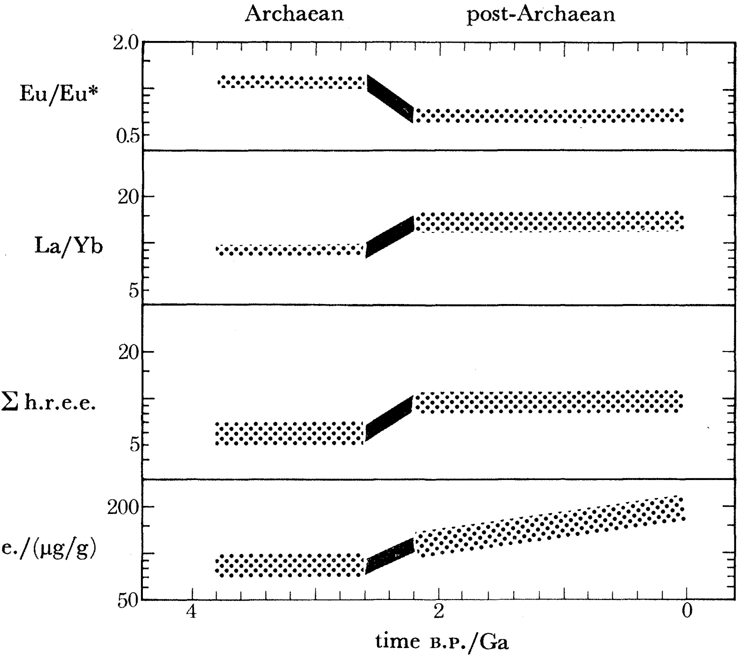

FIGURE 6. The evolution of the rare earth elements in *sedimentary* rocks with time. The change from Archaean to post-Archaean patterns in sedimentary rocks is here represented as taking place during the age limits of Huronian sedimentation. (Adapted from Taylor (1979).)

A gradual change occurs upward in the 12000 m thick succession until the highest unit (Gordon Lake Formation) contains typical post-Archaean r.e.e. patterns, complete with negative Eu anomalies (figure 5). Studies in this laboratory indicate that the change from Archaean to post-Archaean patterns in southern Africa occurs before the deposition of the Pongola system, dated at about 3 Ga B.P. In Australia, the oldest post-Archaean sedimentary patterns recognized occur in the Hamersley Group dated at about 2.5 Ga B.P. (W. Compston, personal communication). Thus the major crustal evolution event that is reflected by the change from Archaean to post-Archaean r.e.e. patterns is not isochronous. This change would form a better marker for the Archaean–Proterozoic boundary than by fixing it at an arbitrary age of 2.5 Ga (James 1978). The overall variation in the r.e.e. with time is given in figure 6.

10. MODELS FOR THE DERIVATION OF THE ARCHAEAN CRUST

The most reasonable model for the formation of present-day continental crust is that it grows by accretion, dominated by island-arc-type volcanism (§3). Does this model hold for the Archaean crust? The r.e.e. evidence is enigmatic (Taylor 1977). The abundance patterns could be produced either by island-arc volcanics, or from a mixture of the tholeiitic basalt and tonalite–trondhjemite suites so common in Archaean terrains (Barker & Peterman 1974). Although island-arc volcanics resembling closely those of present-day arcs do occur in the Archaean (e.g. Marda Complex (Taylor & Hallberg 1977)), the weight of geological evidence favours the bimodal model. This indicates a differing tectonic style in the Archaean. The model developed by Tarney *et al.* (1979) appears reasonable. In this model, partial melting in the mantle produces basalts, which are recycled back into the mantle and are transformed to eclogite. Partial melting at depths where garnet is stable, produce tonalites and trondhjemites (but not K-rich granites) with steep r.e.e. patterns. This model accounts well for the observed r.e.e. patterns, and can provide for those observed in the sedimentary rocks, by an appropriate mixture.

11. The growth rate of the crust

The r.e.e. evidence from the sedimentary rocks indicates a massive change in upper-crustal composition at about the Archaean–Proterozoic boundary. The Archaean upper crust contains less than 10 % granodiorites or K-rich granites, but the post-Archaean upper crust is dominated by granodiorite. Thus the characteristics of the present-day crust were established by about 2.5 Ga ago. The sedimentary rock r.e.e. evidence is thus consistent with a major episodic growth of the continental crust in the period 3.0–2.5 Ga ago, as is strongly indicated by the Nd isotopic evidence (McCulloch & Wasserburg 1978). A massive increase in the areal extent of the upper crust, as inferred from the isotopic data (McCulloch & Wasserburg 1978; Moorbath 1975, 1977), implies a two-stage growth model: (*a*) igneous rocks, derived from the mantle are added to the crust; (*b*) subsequent intrastructural melting produces the granodioritic upper crust.

Growth rates for the continental crust, which are low in the Archaean, increase rapidly between 3 and 2.5 Ga B.P. and then decrease in the Proterozoic and Phanerozoic, are consistent with the r.e.e. data.

12. The evolution of the continental crust

A summary of the preceding discussion is presented in figure 7. This model can be used to address several questions connected with continental growth.

(*a*) The rather slow growth rate for the past 2.5 Ga assists in explaining the 'freeboard' question raised by Armstrong (1969). If 70 % of the volume of continents was accreted before 2.5 Ga B.P. the Proterozoic and later sedimentary sequences will remain close to present-day sea level.

(*b*) The parallel r.e.e. patterns in sedimentary rocks since 2.5 Ga B.P. indicate no fresh input from other than 'granodioritic' material over this period. The slow increase in total r.e.e. is best explained by recycling of sedimentary material in a rather stable cratonic environment (Veizer & Jansen 1979).

(*c*) During post-Archaean time, island-arc volcanic rocks may provide some first-cycle volcanogenic sedimentary rocks (e.g. Baldwin Formation, New South Wales (Nance & Taylor 1977)) with r.e.e. patterns that 'mimic' Archaean-type patterns.

(*d*) The massive increase in continental volume at 3–2.5 Ga B.P. will induce a change in tectonic régime and in the style of igneous activity. The presence of large continental blocks, incapable of being subducted, will produce the change to linear-style belts characteristic of present-day tectonic régimes.

(*e*) What type of mantle igneous activity produced the massive increase in crustal volume between 3.0 and 2.5 Ga B.P.? Was it a continuation of the early Archaean bimodal tholeiitic–felsic régime, or did calc-alkaline island-arc volcanism begin to provide the major contribution to continental accretion, as it has done in later ages? The answer to these problems may well lie in a close examination of the Scourian Terrain (Holland & Lambert 1975; Muecke *et al.* 1979; Hamilton *et al.* 1979).

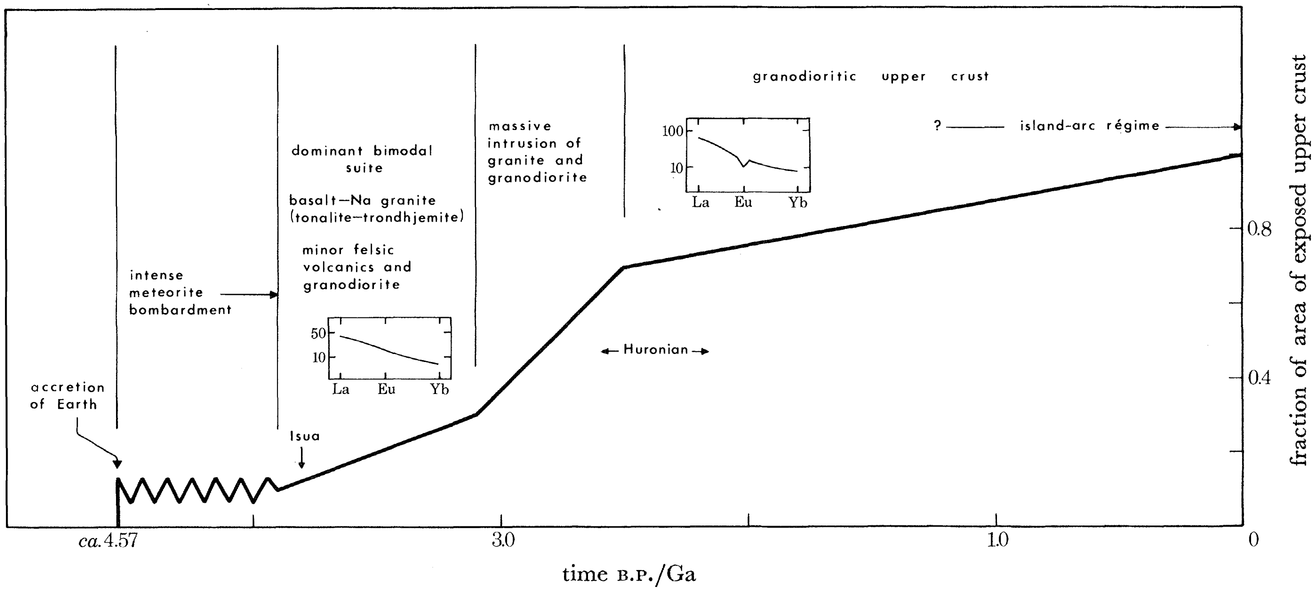

FIGURE 7. Model for the evolution of the continental crust throughout geological time. Average sedimentary r.e.e. patterns are indicated, as are some of the principal events of crustal evolution. The changes in upper-crustal composition, and a massive increase in continental growth, takes place between 3.0 and 2.5 Ga B.P. This change will be reflected a little later in the sedimentary record (e.g. Huronian, cf. figure 6).

(*f*) The isotopic evidence (see, for example: Moorbath 1975, 1977; Hamilton *et al.* 1979) is consistent with a short time interval (*ca.* 0.2 Ga) between derivation of material from the mantle, and final differentiation to produce an evolved continental crust. This production of l.i.l. element enriched upper, and depleted lower crust, with retention of Eu and Sr in the lower crust is consistent with studies of the origin of granites (see, for example: Tuttle & Bowen 1958; Fyfe 1973*a*, *b*).

13. RECYCLING OF CONTINENTAL CRUST THROUGH THE THE MANTLE

Although this debate turns principally on the Sr, Nd and Pb isotopic systematics (see, for example: Moorbath 1975, 1977), the trace-element abundances provide some additional constraints. Subduction of material derived from the present upper continental crust will introduce some characteristic trace-element signatures into the mantle. These will include an Eu depletion of the order of 30% (Eu/Eu* = 0.64). Igneous rocks derived from the mantle contain no memory of such a depletion of Eu in their source regions. Ratios of Th/U are typically greater than four and average five in post-Archaean sedimentary rocks (McLennan

& Taylor 1980*b*). Present-day igneous rocks derived from the mantle frequently have ratios of less than three. Boron abundances in sedimentary rocks are typically about 120 μg/g, much higher than the abundances of about 20 μg/g observed in igneous rocks (Christ & Harder 1969). It is possible that such subducted material is so heavily diluted by rehomogenization in the mantle that these characteristic signatures are swamped. If such is the case, then, the sedimentary trace-element evidence cannot be used to test the hypothesis of recycling, except to place some stringent upper limit (10% ?) on the reappearance of such subducted crustal material.

14. RELATIONSHIP OF THE CRUST TO THE MANTLE AND THE AMOUNT OF TERRESTRIAL DIFFERENTIATION

It remains to consider the relationship of the present-day continental crust to that of the primitive mantle composition. The overall composition of the mantle may be established in the following manner. Core compositions are excluded. It is assumed that most of the core elements Fe, Ni and Co were accreted as separate metallic phases. During segregation of the core material only partial equilibration with the mantle silicates was achieved. This is shown, *inter alia*, by the large abundances of Ni and by trace siderophile elements in the present upper mantle. These elements may have been added from meteorite bombardment (Morgan *et al.* 1980) occurring after core formation. Together with those of the highly volatile trace elements (e.g. Bi), their abundances, except those for Ni and Co, are too uncertain both in the mantle and the crust to warrant further treatment at this stage

The major-element composition of at least the upper mantle, derived, by Taylor (1979*b*), from nodule and pyrolite data, is given in table 2. The refractory elements Ca, Al and Ti are not expected to be fractionated from one another during the accretion of the Earth. Accordingly the ratio of these elements to those in the primitive solar nebula can be used to arrive at values for the refractory trace elements.

Values for the 'solar nebula' abundances are listed in table 1. They are derived by assuming that the abundances in type 1 carbonaceous chondrites give the best estimate of the primitive abundances. This assumption is strengthened by the similarity between solar photosphere and coronal abundances and those in type 1 carbonaceous chondrites (C-1). Full references are given following table 1. The values listed in table 1 have been multiplied by 1.5 to allow for loss of H_2O, C etc. The average of Ca, Al and Ti in the Earth's mantle to the volatile-free C-1 abundances is 1.36. The mantle for the refractory trace elements so obtained are given in table 1. The values for the other elements in the mantle are obtained as listed below table 1. In general, an interlocking set of ratios are used to provide consistent data, and to meet the various elemental and isotopic constraints. It is assumed here that these abundances are typical of the whole mantle and that the lower mantle is not chemically different from the upper mantle. No K, U or Th is assumed to enter the core.

The percentage of each element now residing in the continental crust, relative to that present in the primitive mantle, is given in table 1, the values being based on the assumption that the crust is 0.43% of the mass of the mantle. Figure 8 illustrates that the degree of enrichment of an element in the crust depends on the difference between its ionic radius and valency and those of the elements forming the major mantle minerals (Mg, Fe). This indicates that crystal–liquid fractionation is the principal factor responsible for the derivation of the crust from the mantle (Taylor 1964, 1967). The very large percentage of elements such as Cs, Rb, K, Ba, U, Th, La and Ce in the continental crust indicates the processing of very large fractions of the mantle to produce the crust. The question of the percentage of mantle from which this element extraction has occurred cannot be answered satisfactorily until the question of upper, lower, or total mantle involvement is settled. This basically geophysical question is related as well to possible initial differences in major element composition between the upper and lower mantle, another unresolved question.

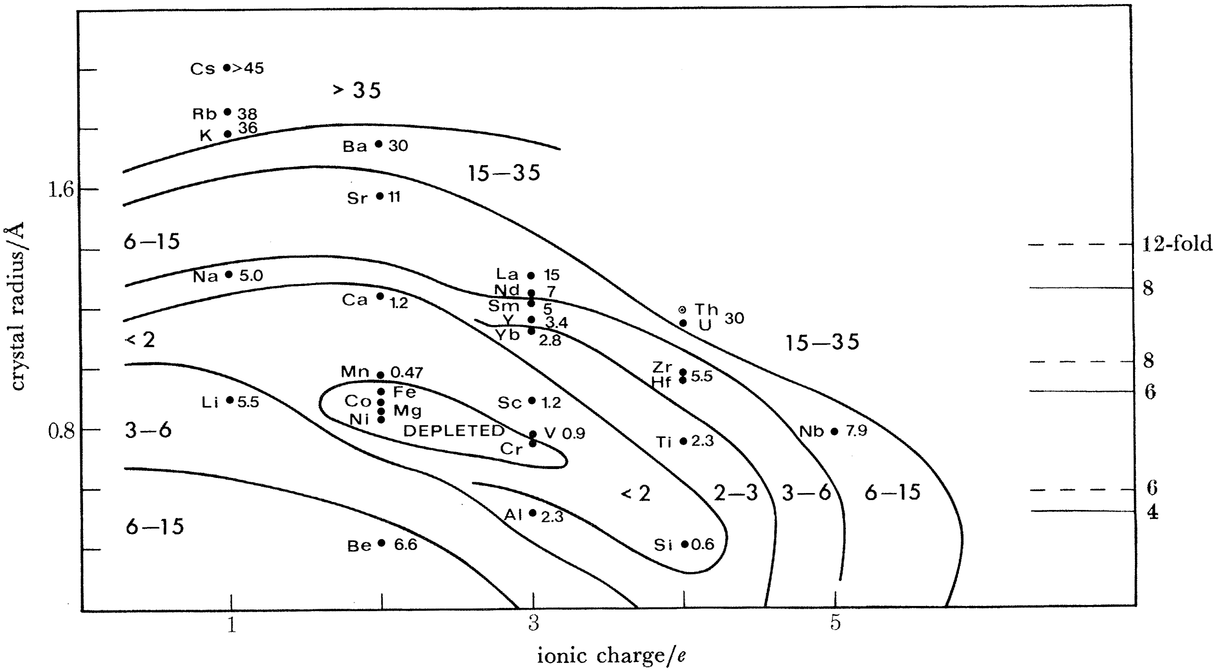

FIGURE 8. The enrichment of elements in the continental crust, relative to the abundances in the primitive mantle (data from table 1), plotted against crystal radius and valency. Crystal radii are used in preference to ionic radii as being probably more representative of physical ion size in crystals (Shannon 1976). On the right side coordination numbers for both 0^{2-}, radii 1.40 (– – – –) and 1.26 Å (——), are shown. Note that the relative enrichment in the crust depends on the difference from the radii and valency of the principal mantle cations (e.g. Mg, Fe).

15. PRE-ARCHAEAN CRUST?

Any crust existing before 3.9 Ga B.P. would be massively altered by large-scale meteoritic bombardment. This has occurred throughout the inner Solar System, from Mercury to Mars, while the surfaces of Ganymede and Callisto bear witness to a probable extension to the neighbourhood of Jupiter. The termination of the bombardment is dated only from the moon, but the widespread morphological evidence for similar cratering encourages the view that such events were common in the period up to 3.9 Ga B.P. There is no reason to suppose that the Earth escaped, a view consistent with the absence of identifiable crust older than 3.8 Ga.

Is there any geochemical evidence available to cast light on these early events? A primitive anorthositic crust, from analogy with the lunar highlands, appears improbable. There is no evidence for a large Eu or Sr spike, nor for the reservoir of very primitive $^{87}Sr/^{86}Sr$ ratios, which would result from the Rb–Sr fractionation inherent in such an event. Walker & Hays (1978), in some definitive experiments, noted that, although a plagioclase-rich crust will float over an anhydrous lunar magma ocean, it will sink in the wet terrestrial environment. Isotopic constraints provide the major argument against primeval sialic crusts, although they might also generate negative Eu spikes (as yet unobserved). Early basaltic crusts are less easy to disprove, but the Sm–Nd evidence appears to rule out any major early mantle differentiation (McCulloch & Wasserburg 1978).

REFERENCES (Taylor & McLennan)

Armstrong, R. L. 1968 *Rev. Geophys.* **6**, 175–199.
Arth, J. G. & Hanson, G. N. 1975 *Geochim. cosmochim. Acta* **39**, 325–362.
Barker F. & Peterman, Z. E. 1974 *Precambr. Res.* **1**, 1–12.
Bavinton, O. A. & Taylor, S. R. 1980 *Geochim. cosmochim. Acta* **44**, 639–648.
Chou, C.-L. 1978 *Lunar Sci.* **9**, 163–165.
Christ, C. L. & Harder, H. 1969 Boron. In *Handbook of Geochemistry* (ed. K. H. Wedepohl), pp. K.1–K.13. Berlin Heidelberg and New York: Springer-Verlag.
Cullers, R. L., Medaris, L. G. & Haskin, L. A. 1973 *Geochim. cosmochim. Acta* **37**, 1499–1512.
Dreibus, G. *et al.* 1976 *Proc. 7th Lunar Sci. Conf.*, p. 3383. New York: Pergamon Press.
Dreibus, G. *et al.* 1977 *Proc. 8th Lunar Sci. Conf.*, p. 211. New York: Pergamon Press.
Drury, S. A. 1978 *Precambr. Res.* **7**, 237–257.
Eade, K. E. & Fahrig, W. F. 1971 *Bull. geol. Surv. Can.* **179**.
Eade, K. E. & Fahrig, W. F. 1973 *Geol. Surv. Can. Pap.* no. 72–46.
Evensen, N. M. *et al.* 1978 *Geochim. cosmochim. Acta* **42**, 1203.
Ewart, A. 1976 *Earth planet. Sci. Lett.* **31**, 417–432.
Fyfe, W. S. 1973*a* *Phil. Trans. R. Soc. Lond.* A273, 457–462.
Fyfe, W. S. 1973*b* *Tectonophysics* **17**, 273–283.
Floyd, P. A. 1977 *Nature, Lond.* **269**, 134–137.
Glickson, A. 1978 *Earth Sci. Rev.* **15**, 1.
Goldberg, E. D., Koide, M., Schmitt, R. A. & Smith, R. H. 1963 *J. geophys. Res.* **68**, 4209–4217.
Goldschmidt, V. M. 1954 *Geochemistry.* London: Oxford University Press.
Goodwin, A. M. 1977 *Spec. Pap. geol. Ass. Can.* **16**, 205–241.
Hamilton, P. K., Evenson, N. M., O'Nions, R. K. & Tarney, J. 1979 *Nature, Lond.* **277**, 25–28.
→ Hargraves, R. B. 1976 *Science, N.Y.* **193**, 363–371.
Haskin, L. A. & Paster, T. P. 1979 In *Handbook of physics chemistry of the rare earths*, vol. 3, ch. 21, p. 18. Amsterdam: North Holland.
Holland, J. G. & Lambert, R. St. J. 1975 *Precambr. Res.* **2**, 161–188.
James, H. L. 1978 *Precambr. Res.* **7**, 193–204.
Jenner, G. A., Fryer, B. J. & McLennan, S. M. 1981 *Geochim. cosmochim. Acta.* (In the press.)
Krahenbuhl, U. *et al.* 1973*a* *Geochim. cosmochim. Acta* **37**, 1353.
Krahenbuhl, U. *et al.* 1973*b* *Proc. 4th Lunar Sci. Conf.*, p. 1325.
Kerrich, R. & Fryer, B. J. 1979 *Can. J. Earth Sci.* **16**, 440–458.
Lee, W. H. K. 1970 *Phys. Earth planet. Int.* **2**, 332–341.
Mason, B. H. 1975 *Grön. Geol. Under. Rep.* **71**, 1–11.
Mason, B. H. 1979 *Prof. Pap. U.S. geol. Surv.* no. 440–B–1.
Masuda, Y. & Aoki, K. 1979 *Earth planet. Sci. Lett.* **44**, 139–149.
→ McCulloch, M. T. & Wasserburg, G. J. 1978 *Science, N.Y.* **200**, 1003–1011.
McLennan, S. M., Fryer, B. J. & Young, G. M. 1979 *Geochim. cosmochim. Acta* **43**, 375–388.
McLennan, S. M. & Taylor, S. R. 1980 *Nature, Lond.* **285**, 621–624.
Moorbath, S. 1975 *Proc. Geol. Ass.* **86**, 259–279.
Moorbath, S. 1977 *Chem. Geol.* **20**, 151–187.
Morgan, J. W., Wandless, G. A., Petrie, R. K. & Irving, A. J. 1981 *Tectonophysics.* (In the press.)
Muecke, G. K., Pride, C. & Sarkar, P. 1979 *Phys. Chem. Earth*, **11**, 449–464.
Nance, W. B. & Taylor, S. R. 1976 *Geochim. cosmochim. Acta* **40**, 1539–1551.
Nance, W. B. & Taylor, S. R. 1977 *Geochim. cosmochim. Acta* **41**, 225–231.
O'Nions, R. K. *et al.* 1979 *A. Rev. Earth planet. Sci.* **7**, 11.
Piper, D. Z. 1974 *Chem. Geol.* **14**, 285–304.
Rogers, N. W. 1977 *Nature, Lond.* **270**, 681–684.
Shannon, R. D. 1976 *Acta crystallogr.* A **32**, 751.
Shaw, D. M., Dostal, J. & Keays, R. R. 1976 *Geochim. cosmochim. Acta* **40**, 73–83.
Shaw, D. M., Reilly, G. A., Muysson, J. R., Pattenden, G. E. & Campbell, F. E. 1967 *Can. J. Earth Sci.* **4**, 829–854.

Shima, M. 1979 *Geochim. cosmochim. Acta* **43**, 353.
Sun, S.-S. & Nesbitt, R. W. 1977 *Earth planet. Sci. Lett.* **35**, 429–448.
Sun, S.-S. & Nesbitt, R. W. 1978 *Contr. Miner. Petr.* **65**, 301.
Taylor, S. R. 1964 *Geochim. cosmochim. Acta* **28**, 1989–1998.
Taylor, S. R. 1967 *Tectonophysics* **4**, 17–34.
Taylor, S. R. 1977 *Am. geophys. Un. Ewing Ser.* **1**, 325–335.
Taylor, S. R. 1979*a* *The Earth: its origin, structure and evolution* (ed. M. W. McElhinny), ch. 11, pp. 353–376. London: Academic Press.
Taylor, S. R. 1979*b* *Proc. lunar planet. Sci.* **10**, 2017–2030.
Taylor, S. R. 1979*c* *Proc. 10th Lunar Sci. Conf.*, table 1. New York: Pergamon Press.
Taylor, S. R. 1980 *Proc. 11th Lunar Planet. Sci. Conf.* New York: Pergamon Press. (In the press.)
Taylor, S. R. & Hallberg, J. A. 1977 *Geochim. cosmochim. Acta* **41**, 1125–1129.
Taylor, S. R. & McLennan, S. M. 1981*a* In *Precambrian plate tectonics* (ed. A. Kröner). Amsterdam: Elsevier. (In the press.)
Taylor, S. R. & McLennan, S. M. 1981*b* *J. geol. Soc. Aust.* (In the press.)
Tarney, J., Weaver, B. & Drury, S. A. 1979 In *Trondhjemites, dactites and related rocks* (ed. F. Barker), ch. 8, pp. 275–299. New York and Amsterdam: Elsevier.
Tuttle, O. F. & Bowen, N. L. 1958 *Mem. geol. Soc. Am.* no. 74.
Veizer, J. 1973 *Contr. Miner. Petr.* **38**, 261–278.
Veizer, J. & Jansen, S. L. 1979 *J. Geol.* **87**, 341–370.
Walker, D. & Hays, J. F. 1977 *Geology* **5**, 425–428.
Walker, R. G. 1978 *Can. J. Earth Sci.* **15**, 1213–1218.
Weller, M. R. *et al.* 1978 *Geochim. cosmochim. Acta* **43**, 999.
Wiik, H. B. 1971 In *Handbook of elemental abundances in meteorites* (ed. B. H. Mason), p. 8, tbl. 3, analysis 1. London: Gordon & Breach.
Wildeman, T. R. & Condie, K. R. 1973 *Geochim. cosmochim. Acta* **37**, 439–453.
Wildeman, T. R. & Haskin, L. A. 1973 *Geochim. cosmochim. Acta* **27**, 419–438.

Discussion

R. L. Armstrong (*Department of Geological Sciences, University of British Columbia, 2075 Wesbrook Place, Vancouver, Canada V6T 1W5*). I do not understand how the rare-earth patterns in sediments provide a measure of crustal volume. Why can't the change 2500 Ma ago be due to a change in processes or rates of processes?

S. R. Taylor. The r.e.e. patterns of post-Archaean sedimentary rocks display a uniform pattern, thus providing a widespread sampling and average for the upper crust. Archaean sedimentary rocks likewise display a uniform pattern, but one that is distinct from the post-Archaean in having a lower La/Yb ratio, and no relative Eu depletion. Archaean-type sedimentary patterns do not persist in rocks younger than about 2500 Ma. The upper crustal r.e.e. patterns after that time are dominated by 'granodioritic' patterns, which have overwhelmed the contributions from the Archaean terrains. Accordingly they must be derived from a much wider area than have the Archaean patterns. Accordingly the sedimentary r.e.e. patterns observed in post-Archaean time are consistent with a massive increase in crustal area (and hence volume) between about 3000 and 2500 Ma B.P.

J. Tarney (*Department of Geology, University of Leicester, Leicester LE1 7RH, U.K.*). Average r.e.e. patterns of deep crustal granulite terrains such as the Lewisian do have a net positive Eu anomaly that might appear to be complementary to the negative Eu anomaly in upper crustal rocks. However, equating the two in this way requires some caution. Our r.e.e. data for the Lewisian rocks (Weaver & Tarney 1981), like those cited by Muecke *et al.* (1979), show that it is the more silicic trondhjemitic gneisses that have large positive Eu anomalies whereas the more basic granulites have either no Eu anomaly or slight negative anomalies. This is the opposite of the relationship expected from intracrustal melting models that propose that the lower crust is more mafic and more refractory through having lost a silicic mobilizate.

S. R. Taylor. Muecke *et al.* (1979) calculated an *average* r.e.e. pattern for the Scourian terrain, which has a pronounced positive Eu anomaly, complementary to the negative Eu anomaly observed in upper crustal rocks. Although they include a majority of 'felsic' rocks in this average, their use of 'felsic' is apparently equivalent to tonalitic (p. 457) and they note that the Scourian terrain has an 'overall intermediate composition and mineralogy' (p. 456).

References

Muecke, G. K., Pride, C. & Sarkar, P. 1979 *Phys. Chem. Earth* **11**, 449–464.
Weaver, B. & Tarney, J. 1981 *Earth planet. Sci. Lett.* (In the press.)