Chapter 2
Chemical Composition of Continental Crust and the Primitive Mantle

2.1 Continental Crust

To study the origin of continental crust, we need to first know its structure and chemical composition, as well as the chemical composition of the primitive mantle that is the source of continental crust.

The Earth's continental crust comprises large continents, such as Eurasia and Africa, and islands with continental crust such as Greenland and New Guinea. This definition also includes mature island arcs, such as the Japanese islands and the Sunda arc, as well as submarine plateaus, such as the Seychelles plateau. The total area of continental crust is 2.1×10^8 km^2, covering 41.2% of the solid Earth's surface (Cogley 1984). Its mass is 2.09×10^{22} kg, corresponding to 0.52% of the mass of the mantle (Taylor and McLennan 1985). Its thickness ranges from 14 to 80 km, averaging 41.1 km (Christensen and Mooney 1995).

Continental crust differs from oceanic crust in origin, chemical composition and structure. Oceanic crust comprises three layers (Raitt 1963). Layer 1 represents a sedimentary cover of the crust. Layer 2 has P-wave velocities from 4.5 to 5.6 km/s, and comprises pillow basalts and underlying sheeted dikes. This layer ranges in thickness from 1.5 to 2 km. Layer 3 represents a gabbroic layer with P-wave velocities from 6.5 to 7.0 km/s. It ranges in thickness from 4.5 to 5.0 km. The boundary between this third layer and the underlying hartzburgite is the Mohorovičić discontinuity, or Moho, which separates the crust from the mantle.

However, based on a characteristic P-wave velocity profile, two to four layers are recognized within the continental crust. The four-layer model of Rudnick and Fountain (1995) is presented here to explain its structure. The first is a surface layer with P-wave velocities of less than 5.7 km/s. It consists mostly of sedimentary and volcanic rocks. The second layer ranges in P-wave velocities from 5.7 to 6.4 km/s, and consists of granitic plutons and low-grade metamorphic rocks. The third layer has P-wave velocities ranging from 6.4 to 7.1 km/s, and is made of gabbroic cumulate with small amounts of intercalated pelitic metamorphic rocks of the granulite facies. P-wave velocities of the fourth layer vary from 7.1 to 7.6 km/s.

T. Yanagi, *Arc Volcano of Japan*, Lecture Notes in Earth Sciences 136,
DOI 10.1007/978-4-431-53996-4_2, © Springer-Verlag Berlin Heidelberg 2011

In most cases, this layer is either very thin or missing. Accordingly, in the latter case, the third layer directly lies on the mantle. However, the second and third layers always are present in continental crust. The boundary between these two layers commonly lies at a depth of 20–28 km beneath the surface of shields, platforms, Mesozoic to Cenozoic orogenic belts and mature island arcs. Since clastic sedimentary and metamorphic rocks are, on average, both granitic in composition, the first and second layers together represent the granitic upper crust. The third layer represents the gabbroic lower crust. It ranges in thickness from 9 to 21 km. The granitic layer is seismologically transparent, while the third layer has many parallel reflectors, suggesting stratification of materials of different physical properties.

2.2 Chemical Composition of Continental Crust

Oceanic crust is chemically homogeneous and basaltic (Table 2.1). Accordingly, its bulk chemical composition can be determined rather easily. On the other hand, it is not easy to estimate the bulk chemical composition of continental crust, because it consists of various types of rocks with different compositions. Estimation of the bulk chemical composition of continental crust of today's definition was made first by Poldervaart (1955). Based on both average chemical compositions and seismic velocities of individual different rock types, he determined chemical composition from rock configuration estimated from the seismic velocity structure model of continental crust. His estimation has since been improved with the same approach by Ronov and Yaroshevsky (1969), and recently by Wedepohl (1995) (Tables 2.2–2.4).

Unlike these estimates, average chemical compositions at the surface of the continental crust (Tables 2.2 and 2.3) were determined by analyzing composite samples produced by mixing of rocks collected from intersections of a lattice network laid over different regions in Canada (Fahrig and Eade 1968; Shaw et al. 1986). In China, similar work was conducted by Gao et al. (1998). They collected

Table 2.1 Chemical composition of the oceanic crust

	Taylor and McLennan (1985) (wt.%)	Condie (1997) (wt.%)
SiO_2	49.5	50.5
TiO_2	1.5	1.6
Al_2O_3	16.0	15.3
T.FeO	10.5	10.4
MgO	7.7	7.6
CaO	11.3	11.3
Na_2O	2.8	2.7
K_2O	0.15	0.2
Total	99.5	99.6

T.FeO Total iron as FeO

Table 2.2 Chemical composition of the upper continue crust

	Shields	Ukrainian-Baltic shield	Upper continental crust	Canadian shield surface	Canadian shield surface	Upper continental crust	Upper contineneta crust [a] before restoration	Upper contineneta rust[a] after restoration	Surface of continental crust of Central East China[b]
	Poldervaart (1955)	Ronov and Yaroshevky (1969)	Ronov and Yaroshevky (1969)	Fahrig and Eade (1968)	Shaw et al. (1986)	Taylor and McLennan (1985)	Condie (1993)	Condie (1993)	Gao et al.(1998)
SiO_2	66.4	66.0	65.2	66.1	66.71	66.0	66.86	67.80	67.81
TiO_2	0.6	0.6	0.6	0.5	0.53	0.5	0.64	0.59	0.67
Al_2O_3	15.5	15.3	15.6	16.1	15.03	15.2	15.26	15.11	14.14
Fe_2O_3	1.8	1.9	2.1	1.4	1.40				2.43
FeO	2.8	3.1	2.8	3.1	2.83	4.5	4.90	4.52	3.13
MnO	0.1	0.1	0.1	0.1	0.07	0.1			0.10
MgO	2.0	2.4	2.3	2.2	2.30	2.2	2.26	1.98	2.61
CaO	3.8	3.7	4.7	3.4	4.23	4.2	3.57	3.25	3.43
Na_2O	3.5	3.2	3.1	3.9	3.55	3.9	3.34	3.36	2.85
K_2O	3.3	3.5	3.3	2.9	3.18	3.4	3.01	3.26	2.67
P_2O_5	0.2	0.2	0.2	0.2	0.15		0.14	0.13	0.16
Total	100.0	100.0	100.0	99.9	99.98	100.0	99.98	100.0	100.0

[a] 2.5 ~ 1.8 Ga.(wt. %)

[b] H_2O and CO_2 were eliminated

Table 2.3 Incompatible element composition of the upper continental crust

	Upper continental crust	Surface of Canadian shield	Upper continental crust[a] before restoration	Upper continental crust[a] after restoration	Upper continental crust	Surface of continental crust Central East China[b]
	Taylor and McLennan (1985)	Shaw et al. (1986)	Condie (1993)	Condie (1993)	Wedepohl (1995)	Gao et al. (1998)
Rb (ppm)	112	110	92	99	110	85
Sr (ppm)	350	316	287	280	316	276
Y (ppm)	22	21	30	32	20.7	18
Zr (ppm)	190	237	174	180	237	195
Nb (ppm)	25	26	11.2	12.1	26	12
Cs (ppm)	3.70				5.80	3.67
Ba (ppm)	550	1070	684	700	668	702
La (ppm)	30	32.3	28.8	30.9	32.3	36.0
Pb (ppm)	20	17	17	18	17	19
Th (ppm)	10.7	10.3	9.2	10.4	10.3	9.27
U (ppm)	2.8	2.45	2.40	2.6	2.5	1.61

[a] 2.5 ~ 1.8 Ga

[b] H_2O and CO_2 were eliminated

Table 2.4 Compositions of the lower continental crust and bulk composition of the continental crust

	Lower continental crust	Lower continental crust	Continental crust	Continental crust	Continental crust
	Taylor and McLennan (1985) (wt.%)	Rudnick and Fountain (1995) (wt.%)	Taylor and McLennan (1985) (wt.%)	Rudnick and Fountain (1995) (wt.%)	Wedepohl (1995) (wt.%)
SiO_2	54.4	52.3	57.3	59.1	61.5
TiO_2	1.0	0.8	0.9	0.7	0.68
Al_2O_3	16.1	16.6	15.9	15.8	15.1
T.FeO	10.6	8.4	9.1	6.6	6.28
MgO	6.3	7.1	5.3	4.4	3.7
CaO	8.5	9.4	7.4	6.4	5.5
Na_2O	2.8	2.6	3.1	3.2	3.2
K_2O	0.34	0.6	1.1	1.88	2.4
Total	100.04	97.80	100.1	98.08	98.36
Rb (ppm)	5.3	11	32	58	78
Sr (ppm)	230	348	260	325	333
Y (ppm)	19	16	20	20	24
Zr (ppm)	70	68	100	123	203
Nb (ppm)	6	5	11	12	19
Cs (ppm)	0.1	0.3	1.0	2.6	3
Ba (ppm)	150	259	250	390	584
La (ppm)	11	8	16	18	30
Pb (ppm)	4.0	4.2	8.0	12.6	14.8
Th (ppm)	1.06	1.2	3.5	5.6	8.5
U (ppm)	0.28	0.2	0.91	1.42	1.7

Total iron as FeO

rock samples from different sites of individual tectonic regions in China, and then mixed them to form composite samples, which were analyzed to obtain average compositions for the individual different tectonic regions.

Instead of analyzing composite samples, sedimentary rocks can also be measured to estimate average composition of surface layers of continental crust. Elements insoluble to water are generally included in clastic particles, while they are carried by river water from continental source areas to submarine depositional sites. Since the smallest particles are least fractionated during transport, fine sediments such as shale maintain the insoluble-element ratios of the source areas. If La content of a shale sample is 30 ppm, then Th content of the source area is estimated at 10.7 ppm, since La/Th ratio averages 2.8 ± 0.2 at the surface of continental crust. The average value of 3.8 for Th/U ratios at the surface of continental crust yields 2.8 ppm as the U content of the source area. Furthermore, the average value of 1.0×10^4 for K/U ratios gives 2.8% by weight as the K content of the source. Based on this method, Taylor and McLennan (1985) determined average abundances of 62 metal elements in the upper continental crust (Tables 2.2 and 2.3).

In addition to these average abundances of elements, we need to identify the effect of erosion on composition at the surface of the upper continental crust. Condie (1993) noticed that the geological configuration at the surface of the continental crust changes systematically in such a way that the surface area of plutonic rocks increases and exposure of sedimentary rocks decreases with advancing erosion. Thus, it is reported that 5–10 km of erosion at the surface does not significantly affect composition at the surface of the upper continental crust (Tables 2.2 and 2.3).

Because surface geology of the analyzed regions consists mostly of metamorphic rocks and granitic plutons, and the second layer of the seismological model of continental crust is thought to be made of low grade metamorphic rocks and granitic plutons, these reported compositions probably represent bulk composition of the second layer and/or a combination of the first and second layers of the seismological model of continental crust.

Estimation of lower crustal composition is much more difficult. Poldervaart (1955) provided the first estimations, and assigned an average composition of basalts that had seismic velocities corresponding to those of the lower crust. His estimation was improved by Ronov and Yaroshevsky (1969). Unlike these estimates, Taylor and McLennan (1985) used a combination of the average mass ratio between upper and lower crust, and the composition of upper continental crust. Their work was based on the assumption that the bulk composition of continental crust is identical to the average composition of arc andesites (andesite model of Taylor and McLennan 1981). The compositions estimated by Taylor and McLennan (1985) are listed in Table 2.4.

In addition, Rudnick and Fountain (1995) determined the lower crust composition by selecting rocks with seismic properties corresponding to those of the lower crust from rocks produced under physical conditions equivalent to those of the lower crust. There are two types of these corresponding rocks. One consists of metamorphic rocks of orogenic belts, and the other comprises xenoliths carried by basaltic lavas from depth. Rudnick and Fountain (1995) assumed that granulite-facies rocks in cratons that have isobaric cooling histories might represent the lower crust. They are mostly basaltic in composition, although they have a few pelitic intercalations. In the estimation of lower crustal composition, they combined xenoliths and granulite-facies metamorphic rocks to explain P- and S-wave velocities of the lower crust. Then, they added some pelitic rocks sufficient to explain terrestrial heat flow rates in continental crust. Their estimated composition is listed in Table 2.4.

2.3 Chemical Composition of the Primitive Mantle

The primitive mantle is defined as the mantle before segregation of the crust. Its chemical composition has been estimated from chemical compositions of chondrites, primitive basalts, peridotites from orogenic belts, and peridotite inclusions in basalts. These peridotites are materials that represent the upper mantle. Except for gaseous elements, the average elemental abundances in CI chondrites are identical to those in the chromosphere of the sun. Therefore, the average composition of the CI chondrites has been considered to represent the chemical composition of solar nebulae.

There are two estimation methods, one mainly from peridotites (e.g., Ringwood, 1991; McDonough and Sun, 1995) and the other from chondrites (e.g., Anderson 1983; Taylor and McLennan, 1985) Ringwood assumed that the combination of primitive basalts and peridotites might represent the primitive mantle (Ringwood, 1991), and he called it pyrolite. The composition as reported in 1991 is presented in Table 2.5. McDonough and Sun (1995) found that peridotite samples were arranged in a line on rectangular coordinates having axes of two refractory element ratios. Based on these findings, they assumed that the chemical composition of primitive mantle might be estimated by reconstituting the peridotite compositions so as to have the same refractory element ratios as those of the chondrites. The titanium abundance in the primitive mantle, for instance, was estimated at 1,200 ppm by extrapolating the linear array of peridotite samples on the coordinates, until the ratio of the linear array became the average value of chondrites. The coordinates had axes of titanium abundance and a refractory element ratio, such as Sc/Yb, Ca/Yb and Sm/Yb. As estimated by the same method, the abundance of any refractory element in the primitive mantle was found to be 2.75 times the abundance of the element in the chondrites. This fact made it possible to estimate abundances of volatile elements in the primitive mantle using various element ratios of the mantle. For instance, since uranium abundance in CI chondrites was 7.4 ppb, the uranium content of the primitive mantle was estimated at 20.3 ppb by multiplying its abundance in the chondrites by 2.75. The U/K ratio of 1.2×10^4 of the mantle gave 240 ppm of potassium content, and the K/Rb ratio of 400 of the mantle yielded 0.6 ppm of the rubidium content of the primitive mantle. Since strontium is a

Table 2.5 Primitive and depleted mantle compositions

	Primitive mantle	Primitive mantle	Primitive mantle	Primitive mantle	Depleted mantle
	Anderson (1983) (wt.%)	Taylor and McLennan (1985) (wt.%)	Ringwood (1991) (wt.%)	McDonough and Sun (1995) (wt.%)	Condie (1997) (wt.%)
SiO_2	49.3	49.90	44.78	45.0	43.6
TiO_2	0.21	0.16	0.21	0.201	0.134
Al_2O_3	3.93	3.64	4.46	4.45	1.18
FeO	7.86	8.0	8.40	8.05	8.22
MgO	34.97	35.1	37.22	37.8	45.2
CaO	3.17	2.89	3.60	3.55	1.13
Na_2O	0.27	0.34	0.34	0.36	0.02
K_2O	0.018	0.02	0.029	0.029	0.008
Total	99.73	100.05	99.04	99.44	99.49
Rb (ppm)	0.39	0.55	0.635	0.60	0.12
Sr (ppm)	16.2	17.8	21.05	19.9	13.8
Y (ppm)	3.26	3.4	4.55	4.30	2.7
Zr (ppm)	13	8.3	11.22	10.5	9.4
Nb (ppm)	0.97	0.56	0.713	0.658	0.33
Cs (ppb)	20	18	33	21	
Ba (ppb)	5220	5100	6989	6600	
La (ppb)	570	551	708	648	330
Pb (ppb)	120	120	185	150	
Th (ppb)	76.5	64	84.1	79.5	18
U (ppb)	19.6	18	21	20.3	3

refractory element, the combination of 7.25 ppm of strontium content of CI chondrites and of 0.03 of the Rb/Sr ratio of the mantle again gave 0.6 ppm for rubidium content of the primitive mantle. Furthermore, since barium also is a refractory element, the combination of 2,410 ppb of barium content of CI chondrites and of 11 for the Ba/Rb ratio of the mantle yielded 0.6 ppm for rubidium content of the primitive mantle, indicating that all these estimates were internally consistent. The primitive mantle composition estimated in this manner by McDonough and Sun (1995) is listed with other estimates in Table 2.5.

2.4 Mass of Primitive Mantle Necessary for Formation of Continental Crust

As already stated, the Earth's continental crust has a two-layer structure and is chemically characterized by the composition of the upper crust. Herein, the amount of primitive mantle necessary to form the present total mass of the upper crust is estimated based on its chemical composition and the composition of the primitive mantle. For this purpose, we will adopt the composition estimated by Taylor and McLennan (1985) for the upper continental crust, and the composition estimated by McDonough and Sun (1995) for the primitive mantle. Potassium, rubidium, cesium, lead, thorium, and uranium are concentrated in the upper continental crust by 118, 187, 176, 133, 135, and 138 times those contents of the primitive mantle, respectively. We may adopt 150 times as a rough enrichment ratio by taking the median. Since all these elements now in the upper continental crust were originally contained in the primitive mantle, this enrichment ratio gives a minimum amount of primitive mantle necessary to produce the present mass of the upper continental crust. If the upper 20 km of the continental crust of an area of 2.1×10^8 km^2 is assumed to approximately represent a total amount of the present upper continental crust, then the minimum amount necessary to produce it may amount to a volume corresponding to a ~1,600 km-thick primitive mantle. The mass of mantle from which continental crust is segregated is called the depleted mantle (see Condie (1997) in Table 2.5 for chemistry). Thus, this estimation indicates that the present amount of this depleted mantle is enormous.

Next, let us estimate the amount of primitive mantle necessary for formation of upper continental crust for the Japanese Islands. For this purpose, we assume that the upper continental crust is 15 km thick. It is a little less than half of the 32–35 km-thick crust of the Japanese Islands. Then, an amount of the primitive mantle reaching to a depth of nearly 2,900 km beneath the Japan arcs is necessary to make this much upper crust. In general, a subducting oceanic plate separates the mantle prism from the lower thick mantle beneath an island arc. In the case of the Northeast Japan arc, for instance, the volume of such a prism beneath the arc is about 1/24 times the necessary amount, and thus it is too small. Even if some portion of the crust is assumed to have formed from sediments carried from the nearby Eurasian continent, an enormous amount of the primitive mantle must have been utilized to form the arc crust. To achieve this amount in the prism beneath the arc, convection could not be avoided in the prism. The incompatible-element enrichment ratios indicate that a large material exchange has taken place between the prism beneath the arc and the adjacent exterior large mantle via this convection. In other words, a great amount of the primitive mantle material must have flowed into the prism, and almost the same amount of depleted mantle material has flowed out from the prism.

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http://www.springer.com/978-4-431-53995-7

Arc Volcano of Japan
Generation of Continental Crust from the Mantle
Yanagi, T.
2011, XI, 123 p. 42 illus., 41 illus. in color., Hardcover
ISBN: 978-4-431-53995-7