UNESCO Encyclopedia of Life Support Systems
Oxford, U.K., 2002
www.eolss.net

Section 16.6.1.3

CONTINENTAL CRUST

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KEYWORDS:

anisotropy, brittle crust, composition, continental arc, continental margin, cratons, crustal thickness, crustal types, ductile crust, extended crust, juvenile crust, lower crust, Moho discontinuity, orogens, plate tectonics, platforms, Poisson's ratio, rift, reflectivity, rheology, seismic velocities, shear-wave splitting, shields, subcrustal velocity, transitional crust, upper crust, upper mantle

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GLOSSARY

Amphibolite facies: The set of metamorphic mineral assemblages in which mafic rocks are composed of amphibole and plagioclase.

Conrad discontinuity: The seismic boundary distinguished in some blocks of continental crust and defined as the transition to the compressional wave velocities VP>6.5 km s-1.

Convection: transfer of heat by the physical movement of the material.

Craton: Large stable block of the continent which was not subject to deformation since the Precambrian.

Crustal types: Segments of the crust with similar geophysical and geologic characteristics.

Curie temperature: Temperature above which the mineral cannot be permanently magnetized.

Ductility: The capacity of a material to sustain substantial deformation without gross faulting.

Eclogite: A high-pressure mafic rock composed of garnet and Na-rich clinopyroxene.

Facies: Appearance, composition or physical environment of a rock.

Felsic rocks: Quartz-rich rocks composed mostly of light-colored minerals.

Heat flow provinces: Regions with similar values of the near-surface heat flow.

Heat producing elements: The elements that generate heat as a result of their radioactive decay (e.g., U, K, and Th).

Igneous rocks: Once molten rocks that have cooled and solidified.

Kimberlites: An explosively emplaced ultramafic volcanic rock, which typically contains many xenoliths.

Komatiites: Mg-rich ultramafic rock, chiefly Archean in age, with unusually high eruption temperatures.

Lithosphere: Crust and uppermost mantle behaving in a quasi-rigid way over geologic time-scales; its thickness varies between tectonic provinces and typically is in the range 80 to 200-250 km.

Mafic rocks: Quartz-poor rocks composed mainly of magnesium- and iron-bearing minerals.

Mantle: Portion of the Earth below the crust and above the core; is divided into the upper mantle (from the Moho down to 670 km depth) and the lower mantle (from 670 km to 2891 km).

Mohorovičić (or Moho) discontinuity: The seismic boundary between the crust and the mantle marked by a change in P-wave seismic velocity to above 7.8 km s-1.

Ophiolite: A piece of oceanic crust and the uppermost mantle that has been uplifted and exposed at the surface.

Orogen: Region which was deformed together to form a mountain belt.

Petrologic Moho: The base of the lowermost crust, which marks the transition from mafic to ultramafic composition. In some regions (chiefly in orogens), it is the base of the eclogitic layer below the seismic Moho which does not show a seismic discontinuity.

Plate tectonics: The process of movement of large lithospheric plates at the Earth's surface; most of tectonic activity occurs at the plate boundaries.

Pn velocity: The compressional wave velocity at the top of the upper mantle.

Poisson's ratio: The ratio of elastic contraction to elastic expansion of a material under uniaxial compression. In seismology related to P- and S-wave velocities as $\sigma = 1 - 1/\left(V_P^2 / V_S^2 - 1\right)$ ($0<\sigma<0.5$).

Quartz: Crystals of silicon dioxide, SiO2.

Rayleigh number: Dimensionless number characterizing the efficiency of convection: the ratio of heat transferred by convection to heat transferred by conduction.

Seismic anisotropy: Variation of velocity as a function of direction, usually reported as the percentage of faster to slower velocity.

Shear-wave splitting: Seismic analogue to optical birefringence, when the incident seismic shear wave is polarized into two orthogonal directions travelling with different velocities.

Tectono-thermal age: The age of the last great metamorphic-tectonic event.

Trench: Long, narrow depression resulting from bending of the lithospheric plate as it subducts into the mantle.

Xenolith: A rock that occurs as a fragment in another igneous rock. Such rocks are often brought to the surface during magmatic or volcanic activity.

SUMMARY

This article summarizes the structure, composition and evolution of continental crust. The major characteristic of continental crust is its thickness, which can vary from less than 20 km to more than 70 km. Seismic velocities in the crust increase with depth and at the base of the crust (the Moho) there is usually a pronounced jump to the upper mantle velocities. In some regions, high-velocity rocks underneath the seismic Moho may have originally been part of the crust; thus the idea of the petrologic Moho is introduced.

Continental crust is highly heterogeneous in three dimensions. However, it is useful to divide it into several layers, differing by seismic velocities and composition. The three-layer model, distinguishing the upper, middle and the lower crust, is the most common. The P-wave velocities in the layers are in the ranges 5.7-6.4 km s^{-1}, 6.4-6.8 km s^{-1} and 6.8-7.6 km s^{-1}, correspondingly. The upper crust has felsic composition (granite/granodiorite), the middle crust - intermediate-to-felsic, while the lower crust of stable continents is mafic with the composition close to basalts. Bimodal distribution of seismic velocities and strong seismic reflectivity, observed in the lower crust in many regions, suggests that it can be formed by a layered sequence of felsic and mafic rocks.

Continental crust may be subdivided into crustal types, i.e. segments of the crust with similar geophysical and geologic characteristics. Such subdivision provides a useful tool for generalized models of the velocity structure and composition of the highly heterogeneous crust of the continents. The primary types of continental crust include shields, platforms, orogens, extended crust, and continental margins.

Continental crust is formed primarily at the continental magmatic arcs and oceanic island arcs, both of which are associated with subduction zones. The uneven age distribution of the juvenile continental crust is related to the secular changes in the mantle convection.

1. INTRODUCTION

The Earth's crust is the outermost part of the lithosphere with thickness ranging from <10 km in the oceans to more than 70 km in continental regions. Three crustal types are recognized – continental, oceanic (*see Oceanic Crust*) and transitional (the latter includes primarily continental margins). Continental crust includes the major continents, their margins, and several submerged microcontinents. It constitutes only 0.4% of the Earth's mass, but covers about 41% of the Earth's surface and comprises 79% of the total crustal volume.

The crust differs from the underlying mantle in seismic velocity and density, reflecting their different composition. The base of continental crust is defined as the *Mohorovičić* (or *Moho* for short) seismic discontinuity, named after the Croatian seismologist who discovered it in 1909. At the Moho seismic velocities abruptly increase from 6-7 km s^{-1} in the crust to about 8 km s^{-1} in the upper mantle. Density and seismic velocities are closely related and the base of the crust is also associated with a density increase. Thus, the existence of this boundary provides striking evidence for the differentiation of the Earth.

Continental crust provides the most complete record of the Earth's geologic history. Its mean age is about 2.5 Ga, while its oldest fragments found in the central parts of the continents are >4.0 Ga old. In contrast, the oldest oceanic crust is only about 160 Ma old because of the rapid recycling of oceanic lithosphere at subduction zones. Thus, studies of continental crust provide a unique opportunity to understand the geologic and geodynamic evolution of the Earth.

Although continental crust is accessible for geologic, geophysical and geochemical studies, its structure is still much less known than the structure of the oceanic crust due to its greater degree of heterogeneity. Most of the knowledge on the nature and composition of continental crust is based on seismic and heat flow studies, complemented by gravity and electromagnetic studies, geologic mapping, stress measurements, geochemical studies, continental drilling, and age determinations.

2. METHODS OF CONTINENTAL CRUST STUDIES

Seismic studies. The most detailed information regarding the structure and composition of the crust is based on seismic refraction and reflection methods. Indeed, the discovery of the base of the crust came from seismological studies which identified the Moho as a sharp seismic boundary, which was defined as the base of the crust. Globally, seismic methods provide high-resolution images of continental crust (Figure 1). In crustal studies two seismic methods play the leading role: reflection and refraction surveys which are mainly based on the use of artificial energy sources (e.g., explosions or air guns) supplemented by natural events (earthquakes).

The first *refraction seismic experiment* was done in 1860s. Since 1920s the refraction seismic method has been routinely used in oil exploration and since early 1940s it provides the basis for determination of the velocity structure of the entire crust. The accuracy of the interpreted seismic velocities is 3% or better. Most modern analyses of refraction data include calculations of both seismic travel-times and amplitudes supplemented by calculations of synthetic (theoretical) seismograms. Wide-angle refraction surveys are broadly used nowadays for studies of the continental (especially lower) crust and provide depth estimates of crustal layers and crustal thickness better than 10%.

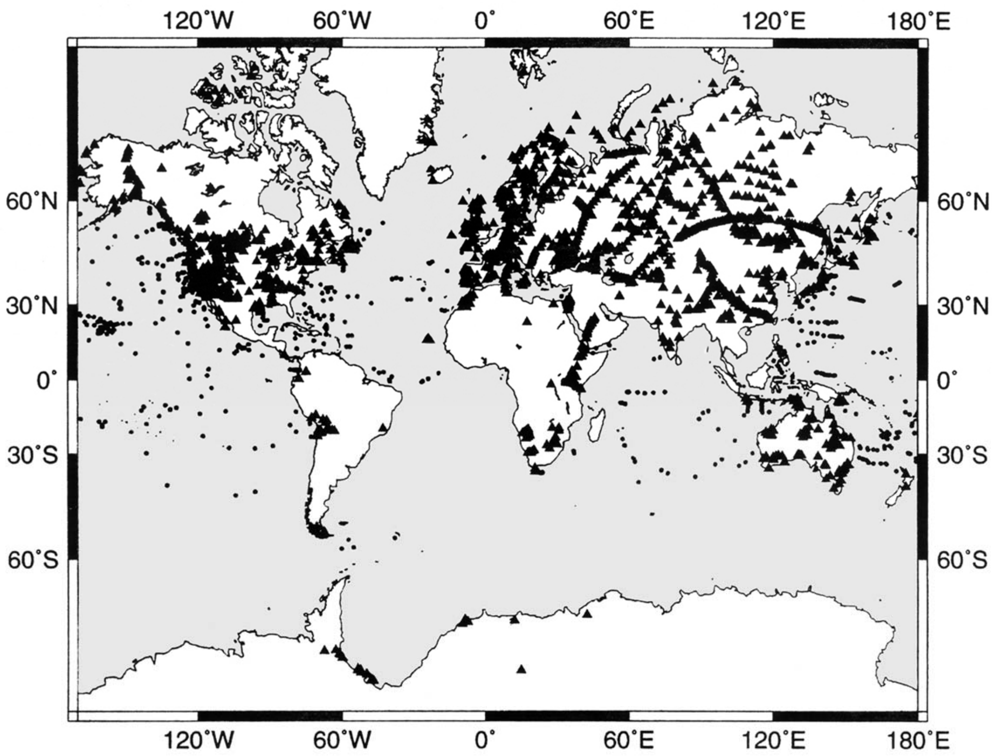

Figure 1: Location of seismic refraction profiles within continents (triangles) and oceans (circles). From Mooney et al. (1998).

Seismic reflection methods provide the most detailed, high-resolution information on the structure of the continental crust. The vertical resolution of this method is some tens of meters for the typical crustal velocities and frequencies used in normal-incidence reflection seismology. However, reflection methods generally do not resolve seismic velocities within the deep crust. Thus, the reflection profiles can be interpreted in terms of the crustal composition only if additional information on seismic velocities is available.

Geologic mapping. Geologic studies of basement outcrops provide a firm basis for models of the composition of continental crust. In some continental regions, deep crustal rocks, that were originally at a depth of 20 to 30 km or more are exposed at the surface as a result of tectonic processes. Such locations permit direct studies of the deep parts of continental crust, their properties and composition.

Petrologic studies. Xenolith studies also provide important information on the composition of continental crust, especially its deep parts, which rarely are available for study at the Earth's surface.

Heat flow studies (*see Terrestrial Heat Flow*). Geothermal modeling permits discrimination between models of the depth distribution of heat producing (radioactive) elements in continental crust and, when combined with laboratory measurements of heat production in different rock types, verification of the composition of the crust derived from seismic experiments. Regional heat flow provinces are typically well correlated with tectonic provinces, as based on distinct crustal structure.

Electromagnetic studies (*see Electric Field of the Earth*). The conductivity structure of the crust is related to its composition and to the presence of pore fluids. The depth to the Curie isotherm, based on magnetic investigations, provides additional control for geothermal constrains. The conductivity contributes auxiliary information on the crustal composition and, in particular, on the depth distribution of crustal fluids.

Gravity studies (*see Applications of Gravimetry and Methods of Survey*). Rock density and seismic velocity are closely related, and thus combined gravity and seismic data can provide the basis for assessing density distribution in continental crust. Gravity studies can help in discriminating between competing seismic models and distinguish density inhomogeneities in the deep crust that may not be evident in seismic data.

Laboratory ultrasonic measurements. The laboratory measurements of V_P and V_S velocities in different rock types provide the basis for models of the crustal composition derived from seismic velocities. Usually these experiments are made at high pressures and temperatures to simulate the *in situ* conditions in continental crust. Additional parameters derived from laboratory measurements, such as Poisson's ratio and seismic anisotropy, are important for discrimination between competing models of the crustal composition.

Continental drilling. Drilling of continental crust is an extremely technically complicated and expensive enterprise. However, it provides a unique opportunity for the direct study of the structure and properties of continental crust. The first deep drilling project was started in the early 1970s in northwestern Russia and still continues. At present the Kola Superdeep Borehole is the deepest drilled borehole with a depth of about 13 km deep. Other deep drilling projects on the continents include the KTB (Continental Deep Borehole) in Germany and the deep borehole in the Urals (Russia).

Geochronology. Age dating of the crustal rocks is important for understanding of the timing and the thermal (e.g., magmatism and metamorphism) and tectonic (e.g. extension and thrusting) processes by which continental crust is formed and modified (*see Tectonic Processes*). In most cases these processes are coupled and are usually referred to as tectono-thermal events.

3. AVERAGE SEISMIC STRUCTURE OF CONTINENTAL CRUST

3.1. Crustal thickness and seismic velocities

Seismically, the crust is defined as the outer layer with compressional (or primary, *P*-) wave velocities (V_P) less than 7.6-7.8 km s^{-1} (with an average velocity of 6.45 km s^{-1}) and shear (or secondary, *S*-) wave velocities (V_S) less than 4.3 km s^{-1} (with an average velocity of 3.65 km s^{-1}). The average V_P crustal velocity range from 5.6 to 7.4 km s^{-1}; however in 85% of continental crust $6.1 \leq V_P \leq 6.7$ km s^{-1}. Typically, seismic velocities in continental crust increase with depth (Table 1).

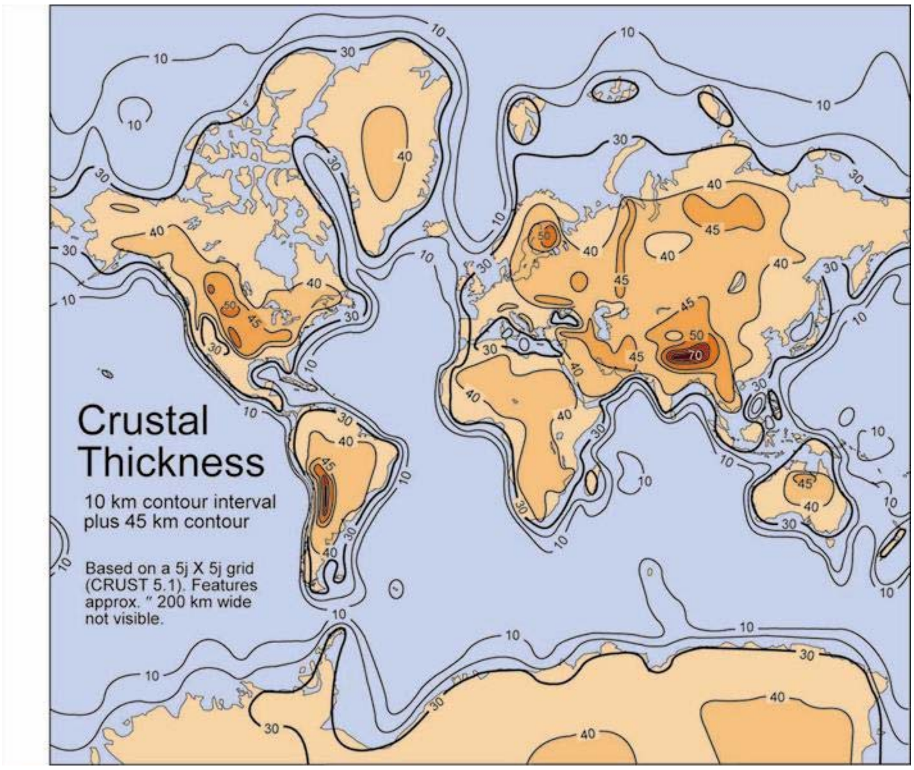

Figure 2: Mercator projection of crustal thickness based on seismic refraction profiles shown in Figure 1. From Mooney et al. (1998).

Table 1: Average crustal velocities weighted at 5-km interval. After Christensen and Mooney (1995).

Depth, km	5	10	15	20	25	30	35	40	45	50
V_P, km s^{-1}	5.95± 0.32	6.10± 0.25	6.30± 0.30	6.45± 0.30	6.65± 0.30	6.78± 0.35	6.92± 0.30	7.02± 0.32	7.10± 0.38	7.15± 0.40

Total thickness is a basic parameter characterizing continental crust (Figure 2). On the continents it varies from 16 km in the Afar Triangle (Ethiopia) to 72 km in Tibet (China) (with an accuracy of about 10%). About 95% of all seismic measurements indicate continental crust 22 km to 57 km thick with the most typical values of about 35-45 km. The mean thickness of continental crust is 39.2 ± 8.5 km. The average (weighted by area) crustal thickness on the continents was recently estimated to be 41 km.

Continental crust is very variable and does not have a standard structure. Two- or three-layer models of the crystalline continental crust based on seismic data are most common, although one-layer models or models with more than three layers were proposed for some regions. For example, some shield areas may be characterized by a one-layer crust formed by high-grade metamorphic rocks with V_P =6.5-6.6 km s^{-1}.

The sedimentary cover forms an additional uppermost layer of the crust. Its thickness can vary from zero on the shields to more than 20 km in deep sedimentary basins (e.g., the Caspian Basin). On continents, compressional wave velocity is 1.5-3.5 km s^{-1} in *unconsolidated (soft) sediments* and 3.5-5.8 km s^{-1} in the *consolidated (hard) sediments*. As the rocks of the sedimentary cover are often metamorphosed and become seismically indistinguishable from the crystalline rocks of the basement, the upper part of the crystalline crust is not always well defined, at least in the regions with thick sediments. Usually its top is assumed to have compressional wave velocities V_P > 5.3-5.8 km s^{-1}. In many regions, metamorphosed Paleozoic sediments are included as part of the upper crystalline crust.

The upper continental crust, constituted chiefly by gneisses, granites and granodiorites, has 5.6-5.8 < V_P < ~6.4 km s^{-1} and a typical thickness of 10-25 km. In the deeper crust an increasing mafic content and metamorphic grade both raise seismic velocities. In the *middle crust* (usually 5-15 km thick), which is usually composed of rocks in amphibolite facies, velocities are ~6.4 < V_P < ~6.8 km s^{-1}. *P*-wave seismic velocities in *the lower crust*, which is formed by metamorphic rocks in granulite facies (chiefly diorites, gabbros, amphibolites and granulites), range from ~6.8 to ~7.2 km s^{-1}. In Precambrian shield and platform areas *the lowermost crust* may have very high *P*-wave velocities (~7.2 < V_P< ~7.6 km s^{-1}).

In some continental regions a small boundary at mid-crustal levels (usually between 15 and 25 km depth), referred to as the *Conrad discontinuity*, is recognized. This is a gradational boundary, separating the upper "granitic" and the middle (or in some cases lower "mafic") continental crust and was first identified by Conrad in 1925 as the discontinuity where *P*-wave velocities become higher than 6.5 km s^{-1}. In some regions a low velocity zone is found at the base of the upper crust, enhancing the sharpness of the Conrad discontinuity. The Conrad discontinuity varies in depth and character from region to region, suggesting it is not a global feature as it was previously believed and its origin can be diverse. However, in some regions this boundary is significant and most likely reflects the result of the differentiation of crustal material into light sialic rocks (granites) in the upper crust and heavier mafic rocks of the deeper crust.

The compressional (*P*-) wave velocity at the top of the ("normal") peridotitic upper mantle is often referred to as the *Pn* velocity. *Pn* velocities usually are in the range from 7.6 to 8.8 km s^{-1}, however values less than 7.8 km s^{-1} and exceeding 8.4 km s^{-1} are not common. The global average for *Pn* velocity in the continents is 8.07 ± 0.21 km s^{-1}.

3.2. Crustal reflectivity

High-resolution seismic reflection studies provide detailed information on the structure of continental crust. Crustal reflectivity typically appears within continental crust as reflecting fault zones in the commonly seismically transparent upper crust and as subhorizontal laminated strong reflections in the “mafic” lower continental crust.

The seismic reflection programs in North America (COCORP) and Europe (BIRPS, ECORS, DEKORP) show that the reflections from the upper continental crust are chiefly produced by single faults or fault zones. Examples can be found in many tectonically young regions, as for example at the North Variscan Deformation Front. In several cases the origin of the upper crustal reflectivity was determined by drilling or by field-mapping studies of outcrops (an example is the Sijian meteorite impact structure in Sweden, where strong reflections are from horizontal dolerite intrusions within the granitic host rock).

The origin of the lower crustal reflectivity is still a subject for speculation. Usually four origins of the layered reflectivity of continental crust are considered: (1) Igneous (compositional) layering caused by mafic intrusions into the crust (e.g. magma chambers) or by lenses of partial melt of the lower crust; the Basin and Range province in western USA is an example. (2) Metamorphic layering caused by regional metamorphism. Ductile flow in the warm crust during a thermal event can produce subhorizontal layering of melting products. The examples of such lamellae reflectivity are known in the southern Appalachians (USA) and in the Archean granulite terrains. (3) Dynamic layering in the crust caused by partial melting or mylonitization within shear zones and fault zones. (4) Pore pressure layering associated with suture zones or fluid-filled cracks.

The large diversity in the seismic velocities observed in the lower crust suggests that the composition of the lower crust can vary within a wide range and therefore the lower crustal reflectivity may be due to different mechanisms in various geologic environments. The reflectivity pattern is very consistent within similar tectonic provinces, which implies that they have undergone common processes of tectonic evolution.

The general character of the crustal reflectivity correlates with the thermo-tectonic age. Usually the reflectivity is high in tectonically young and warm Phanerozoic areas, such as continental rift zones and regions with extended crust (e.g. in the western part of USA). In these regions the zone of high reflectivity usually extends from near the Conrad discontinuity down to the Moho discontinuity, where it abruptly disappears. Typically the crustal reflectivity in ancient tectonic provinces (the Precambrian continental shields and platforms) is very weak, especially in their lower crust. In these regions the reflection Moho cannot be reliably determined.

3.3. The Moho discontinuity

The Moho is defined as the transition boundary where *P*-wave seismic velocities, as measured by seismic *refraction data*, increase from crustal to mantle values. This change in seismic velocities is typically large 0.5 to 1.5 km s^{-1}. In some regions seismic velocities may change gradually from crustal to mantle values and the Moho does not appear as a sharp seismic discontinuity. Detailed seismic reflection and refraction studies indicate that in many regions the Moho is not a simple boundary, but may occasionally be offset by faults, either strike-slip (e.g., the Pyrenees) or thrust faults in orogen (e.g., the Alps). In such regions double seismic Mohos may be observed.

The reflection Moho is often not as strong as expected from the velocity contrast. In many regions the reflections from the Moho are much weaker than from mid-crustal depths. Based on the most commonly observed pattern, the reflection Moho is defined as the lowest strong (sub-horizontal) normal-incidence reflection in the deep crust, which often coincides with the transition to the seismically transparent upper mantle. The refraction Moho is a global discontinuity, whereas the reflection Moho usually has a piecewise character typically does not appear as a sharp boundary, but rather as a zone, usually nearly horizontal at large distances. Regions with a flat reflection Moho are often interpreted as modified by igneous intrusions and ductile deformation at the base of the crust, such that the Moho, which now is a transitional (3-5 km thick) zone formed by a lamination of thin layers with different seismic properties. In most regions where modern seismic data are available, the depth to both the reflection and the refraction Moho is the same within the resolution of the techniques.

In some collisional orogens the seismic Moho may not always coincide with the base of the petrologic crust. Under high pressures, a phase transition of granulites in the lower crust into high-density eclogite facie rocks may occur. Seismic velocities in eclogites are similar to those in the peridotitic uppermost mantle and thus the petrologic base of the crust (i.e., the bottom of the eclogitic layer) may not show a seismic discontinuity. However, the eclogitic roots have a chemical composition typical of the crust and, therefore, in a petrologic sense are still part of the crust. Thus, three types of Moho are distinguished: The *seismic refraction Moho* is defined by an increase in seismic velocities. The *seismic reflection Moho* corresponds to the depth where the deepest set of normal incidence reflections vanishes. The *petrologic Moho* refers to the transition from felsic-mafic to ultramafic composition typical of the upper mantle. However, usually the term Moho refers to the seismic refraction Moho.

4. Types of Continental Crust

Continental crust may be subdivided into crustal types, i.e. segments of the crust with similar geophysical and geologic characteristics. Such subdivision provides a useful tool for generalized models of the structure and composition of the highly heterogeneous crust of the continents. The subdivision of the crust into the types is primarily based on the geologic ages of the near-surface rocks, surface topography, tectonic history and geologic setting. Geophysical studies have shown that crustal types distinguished on the basis of the near-surface, chiefly geologic, observations exhibit many common features in their deep structure.

The primary types of continental crust include (Figure 3): Precambrian (older than 570 Ma, subdivided into Archean and Proterozoic) shields, Precambrian platforms, Paleozoic (250-570 Ma) orogens, Meso-Cenozoic (younger than

250 Ma) orogens, continental rifts, regions with extended crust, and (in some studies) large volcanic (igneous) provinces. Active and passive continental margins comprise transitional crust.

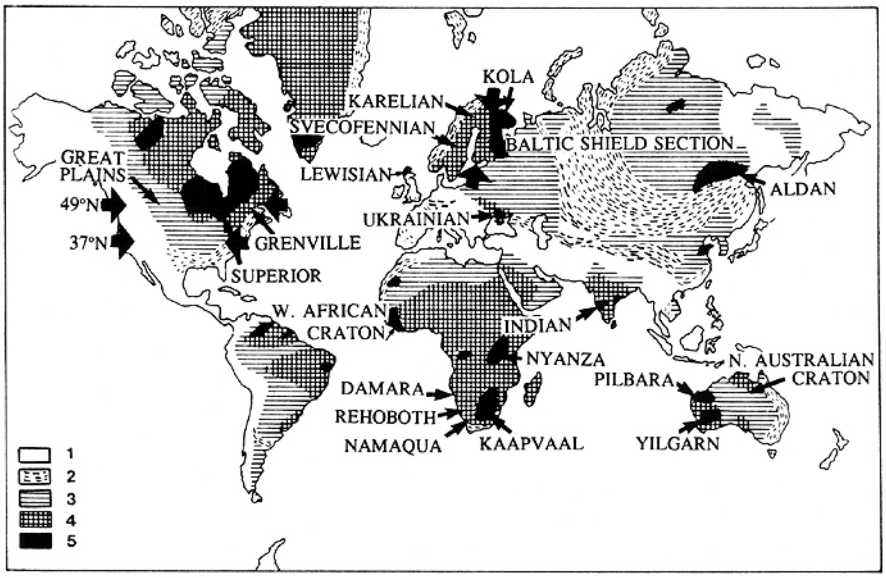

Figure 3: World map of geologic age provinces. (1) Mesozoic and Cenozoic (< 250 Ma) orogenic belts, (2) Paleozoic (250-570 Ma) orogenic belts, (3) Proterozoic (570-2500 Ma) platforms, (4) Proterozoic (570-2500 Ma) shields, (5) Archean (2.5-4.0 Ga) shields. From Durrheim and Mooney (1994).

Table 2: Geophysical characteristics of continental crust (volume-weighted). After Christensen and Mooney (1995).

Crustal type	Fraction of total crustal volume (of continental crust), %	Average crustal thickness, km	Thickness of:				Average crustal velocity, km s^{-1}	*Pn* velocity at the top of the mantle, km s^{-1}
			sedimentary cover, km (V_P<5.7 km s^{-1})	upper crust, km (5.7<V_P< <6.4 km s^{-1})	middle crust, km (6.4<V_P< <6.8 km s^{-1})	lower crust, km (6.8<V_P< <7.8 km s^{-1})		
Precambrian shields and platforms	46 (58)	41.5±5.8	2	15	13	11.5	6.42±0.20	8.13±0.19
Pz and Mz-Cz orogens	27 (34)	46.3±9.5	2	19.5	13.5	11	6.39±0.25	8.01±0.22
Continental rifts	1 (1)	36.2±7.9	2	15.5	9	10	6.36±0.23	7.93±0.15
Extended crust	1 (1)	30.5±5.3	3	17	7	3.5	6.21±0.22	8.02±0.19
Continental arcs	4 (6)	38.7±9.6	0	17.5	11	10	6.44±0.25	7.95±0.23

Though a subdivision of the highly heterogeneous continental crust into a limited number of types is a strong oversimplification, it permits to distinguish systematic differences in the deep structure of different tectonic provinces that may be related to the processes of their formation and tectonic reworking (Table 2). The thickest continental crust is formed beneath young collisional orogens and Precambrian shields. Values close to the continental average are observed in the platforms (40 km), which form about 40% by volume of continental crust and about 32% of the total crustal volume. Relatively thin crust (30 km or even less) is characteristics of Phanerozoic extensional areas of the continents. The thinnest continental crust (<30 km) is typical for the continental margins (shelves, arcs and forearcs) and some highly extended rift zones.

4.1. Shields and platforms

The Precambrian (i.e. Archean and Proterozoic) shields and platforms (often referred to as *cratons)* are the oldest provinces on the Earth (older than 570 Ma) and together they comprise more than 50% of the continental crustal volume (Table 2). Cratons are the most tectonically stable provinces of the Earth, which have not undergone major deformation since the Precambrian. Igneous activity within the cratons is usually very limited in space and volume and is presented chiefly by kimberlites and lamproites. Average surface heat flow is 41 ± 11 mW m^{-2} in the Archean terrains and increases to 55 ± 14 mW m^{-2} in late Proterozoic structures. Thus, a further subdivision of the Precambrian cratons into the age groups: Archean (>2.5 Ga), early (2.5 to 1.6 Ga), middle (1.6 to 0.9 Ga) and late Proterozoic (900-570 Ma) is sometimes useful.

Shields are distinguished from platforms by the absence of sedimentary cover. They have a highly variable crustal thickness, which ranges from 26 km to 65 km (as in the north-eastern part of the Baltic Shield) and can vary by about 20 km at distances of several hundred kilometers (e.g., in the Abitibi province of the Canadian Shield). The average crustal thickness in shields is about 45 km, with about equal thickness of the upper, middle and lower crusts; a large part of the Archean shields is covered by ice (as in Greenland and Antarctica).

Precambrian platforms (e.g., the East-European Platform) and Paleozoic platforms (e.g., West Siberia) have a highly variable thickness of sedimentary cover (typically ranging from 1 to 3 km, but exceeding 20 km in some structures, e.g. in the Caspian Basin) and a relatively thin lower crust (about 10 km compared to 16 km beneath shields). Thickness of the upper and middle crustal layers and total crustal thickness (40-45 km) is about the same as in shields.

In many Proterozoic cratons the lowermost crustal layer has a high *P*-wave velocity (>7.2 km s^{-1}) and results in a very high average crustal velocity (6.4 to 6.7 km s^{-1}). *Pn* velocities beneath the shields are very high, typically in the range 8.1-8.4 km s^{-1}, but in some regions (e.g. East Siberia and Central Canada) they can be as high as 8.6 km s^{-1}, probably partly due to the anisotropy of the olivine-rich upper mantle rocks. The typical *Pn* velocities of Precambrian platforms are slightly higher than the continental average (8.2 km s^{-1}).

A number of *large volcanic (igneous) provinces* (LIPs) are known chiefly at the cratonic margins. The largest include the Siberian trapps and the Deccan trapps in India. The models of their formation include a plume origin or small-scale convection at the cratonic edges where the lithospheric thickness changes. The volcanic provinces along the Atlantic coasts are associated with the opening of the Atlantic (the Parana Basin in South America is the largest). Crustal thickness in LIPs is about 40 km, with 10 km thick upper and middle crust above a very thick lower crust (20 km), which may be ascribed to thickening by intrusions of basaltic magmas. This results in relatively high average crustal velocities in LIPs, similar to those observed in shields; but *Pn* velocities are much lower than the continental average because of higher subcrustal temperatures.

4.2. Collisional orogens

The collisional orogens are mountain belts, ranging from hundreds to tens of thousands of kilometers in length, which were formed as the result of plate collisions. The pre-Paleozoic orogens are parts of cratons (e.g., the Trans-Hudson Orogen in the Canadian Shield). The Uralian, Variscan (250-320 Ma) and Caledonian (400-440 Ma) orogenic belts are examples of Paleozoic orogens. The Meso-Cenozoic orogens (younger than 250 Ma) include the high mountain belts on all of the continents.

The crustal structure of young (tectonically active) and old (stable at present) collisional orogens is substantially different. Young collisional orogens have the thickest crust, regionally >70 km with the average value of about 54 km and with upper, middle and lower crustal thickness of about 22, 14 and 18 km, respectively. The thickness of the upper crust is increased by the presence of granites and granodiorites which are igneous rocks formed as the result of extensive crustal melting during the collisional orogeny. In Tibet the crustal thickness is doubled due to the overthrusting of the Eurasian plate over the Indian plate. In the intracontinental orogens the crustal thickness usually decreases with age. In some Paleozoic orogens (e.g. the Caledonides and Variscides of Europe), the crust is thin (about 33 km) due to the relatively thin upper crust (about 10 km) that was eroded since the end of the tectonic activity and an apparent loss of the lower crustal root, probably during the post-tectonic extensional collapse or eclogite formation.

A characteristic feature of all intra-continental orogens is well-developed, subhorizontal seismic reflectors in the lower crust. In young orogens the reflectors in the middle and lower crust may have a very complex pattern, dipping to the center of the crustal root (as for example in the Alps) and suggesting crustal wedging. The velocity contrast at the Moho beneath the intra-continental collisional orogens is one of the largest, 0.5-1.5 km s^{-1}.

4.3. Continental rifts and extended crust

Continental rifts are long narrow depressions bounded by normal faults. Usually continental rifts are associated with volcanism, extension and show an anomalous seismic structure of the crust and the subcrustal lithosphere. The continental rifts are closely related to the plate tectonic processes (*see Tectonic Processes*) and in some cases (e.g. the

Afar Triangle in Ethiopia) provide a unique opportunity to observe the initial stage of the continental break-up and the evolution into mid-oceanic ridges.

The major active continental rifts include: (a) the European Cenozoic rift system, which extends from the Mediterranean to the North Sea over a distance of 1000 km and is a continuous system of rift structures, the largest of which is the Rhine Graben; (b) the East African Rift system, which includes the Kenya Rift, the Western Rift and the Ethiopian Rift; (c) the Rio Grande Rift in USA; (d) the Baikal Rift in Siberia. The tectonically active Basin and Range province in western USA is an example of extended continental crust, which may be considered as a multiple-rift system.

Paleo-rifts (i.e. rift structures which are not tectonically active at present) include the Oslo Rift and Central Graben in Europe and the Midcontinent Rift system in the central USA. Paleo-rifts are also distinguished on all other continents. The oldest known continental rifts are Precambrian in age (e.g. the Keweenawan rift in north-central USA).

Geodynamic models of the formation of the continental rifts and large extensional structures can be divided into the "active" and "passive" models. The models of an *active rift* suggest that crustal extension and associated rifting is driven by a thermal pulse in the mantle. This results in an ascent of hot mantle material to shallow depths (in some cases to the base of the crust) and subsequent lithospheric extension and thinning.

The *passive models* imply that lithospheric extension (caused by plate boundary forces or by basal drag of the lithosphere with respect to the underlying mantle) is the primary mechanism of rifting, while ascent of hot mantle material into the lithosphere is a secondary process. The Baikal Rift in Siberia, the Basin and Range province in western USA and the Rhine Graben in Europe are possible examples of passive rifting.

Extended continental lithosphere has a typical crustal thickness of 30 km. The average crustal velocities in regions with a highly extended crust and the rift zones are low (about 6.2 km s^{-1}) due to stretching of the lower crust to 4-14 km (Table 2) and high crustal temperatures. The lower crust is often highly reflective (e.g., the Basin and Range Province in western USA). The reflection Moho in extensional provinces is nearly flat, suggesting that the Moho is a young tectonic boundary. Low *Pn* velocities (7.6-8.1 km s^{-1}) are characteristic of the extensional provinces and is generally considered to be the result of high mantle temperatures.

4.4. Continental margins

Active continental margins are associated with subduction zones and include the continental arcs and forearcs. These are widely developed along the margins of the Pacific Ocean. Continental arcs are formed as the result of subduction of an oceanic plate underneath a continent (e.g., the Cascades in North America). They are highly volcanically active, with a gradual lateral change of the composition of magmas across them, associated with varying pressure-temperature conditions at the wedge of the subducting slab as it moves downwards.

In the continental arcs, where the resolution of seismic data is low, the crustal structure appears to be highly heterogeneous. Crustal thickness can vary from about 5 km to more than 40 km., with an average of about 30 km. The upper crust of the arcs is typically thickened to about 10-20 km by the products of magmatic activity. Average velocities in the crust of arcs and forearcs are very low, 6.09-6.14 km s^{-1}. However, two- or three-layer crustal models cannot be applied to many of the arcs. Forearcs (e.g. the Pacific coasts of North America) are marine sedimentary basins on the trench (oceanic) side of arcs with almost no volcanic activity and a thin transitional crust (about 26 km) formed by several kilometers of sediments (chiefly volcanogenic turbidites) on the top of the metamorphic basement.

Passive continental margins occur at the transition from the continents to the oceans (*see Continents on the Move*) and include the continental shelves. Crustal thickness is typically less than 25 km. The upper part of the crust is formed mostly by sediments, while the lower crust can be the oceanic (as for example the continental shelf off eastern North America). Subsidence at passive margins may be a result of a crustal thinning due to flow of the ductile lower crust towards the ocean and the subsequent deposition of sediments in a newly formed basin. Another, thermal, mechanism (e.g., a mantle plume) for the formation of passive continental margins implies lithospheric thinning and associated crustal erosion, followed by cooling, subsidence and basin formation. Some passive margins were initiated as continental rifts as indicated by their rift-type crustal structure.

5. PHYSICAL PROPERTIES OF CONTINENTAL CRUST

5.1. Seismic velocities in typical crustal rocks

Data on seismic velocities in continental crust provides a basis for petrologic models of crustal composition. Laboratory velocity measurements (the accuracy of each measurement is better than 1%) for the common crustal rocks are presented in Figure 4 (average velocities and standard deviations). The wide range of seismic velocities measured in crustal rocks is related to their variable mineralogy, porosity and, sometimes, anisotropy. In ultramafic rocks the variations in seismic velocity chiefly reflect serpentinization and anisotropy.

The number of crystalline rocks with velocities less than 6.0 km s^{-1} is very limited. Velocities in the range 6.0 to 6.4 km s^{-1} (upper crust) are typical for a large number of rock types, including granites, tonalites, felsic gneisses, phyllite, mica quartz schist, granite and felsic granulite facies rocks. Most rocks with middle and lower crustal velocities (6.4 to 7.2 km s^{-1}) are mafic in composition (the exceptions are marble, anorthosite and hornblendite) and include diorite, diabase, greenschist facies basalt, mafic granulite, amphibolite and gabbro. No rocks appear to have seismic velocities in the range 7.2 to 7.6 km s^{-1}. High seismic velocities (>7.6 km s^{-1}) are typical for pyroxenite, eclogite and dunite.

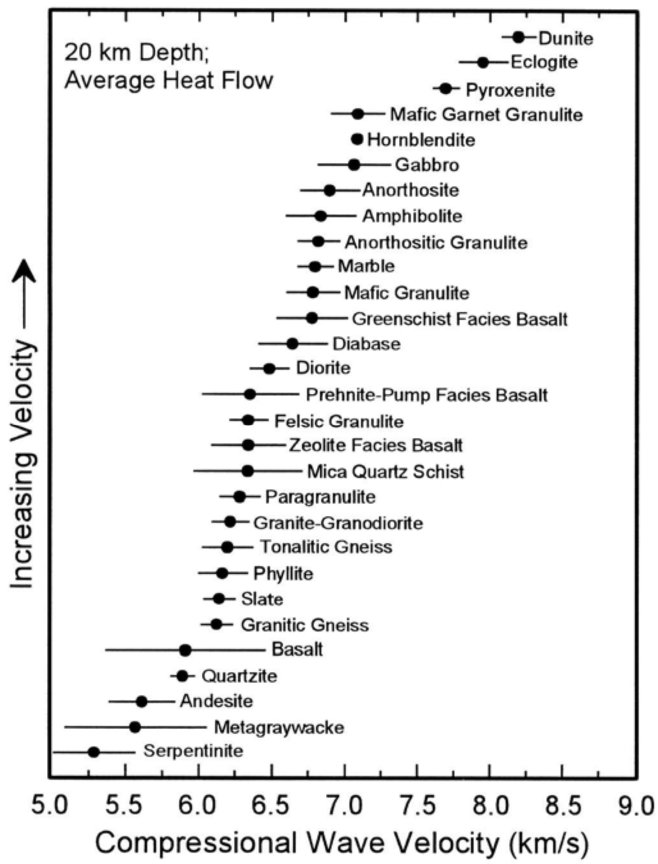

Figure 4. Compressional wave velocities and standard deviations at pressures corresponding to 20 km depth and 309°C (which corresponds to the average surface heat flow) for major rock types. From Christensen and Mooney (1995).

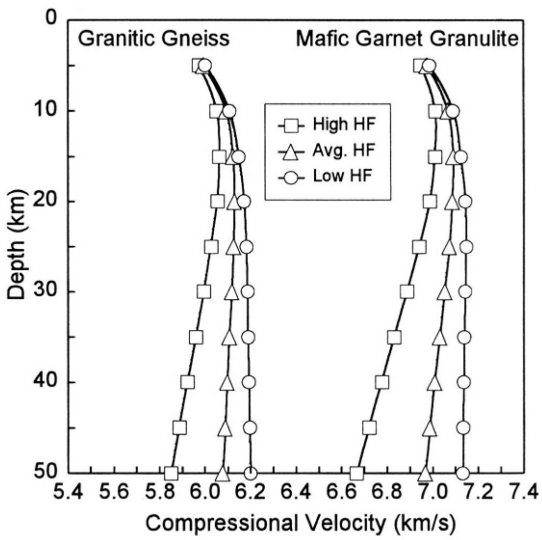

Figure 5. Compressional wave velocity versus depth and heat provinces for average granite/granodiorite and mafic garnet granulite. Heat provinces approximately correspond to the surface heat flow of 35 mW m^{-2} (low), 60 mW m^{-2} (average) and 90 mW m^{-2} (high). From Christensen and Mooney (1995).

Both temperature and pressure affect seismic velocities. The total pressure-temperature effect is non-linear and is shown in Figure 5 for the upper and lower crustal rocks. The increase in the upper crust is caused by the closure of grain boundary cracks. At deeper levels the temperature effect dominates and seismic velocities may decrease with depth in the lower crust.

5.2. Seismic anisotropy in continental crust

Seismic anisotropy exists in all layers of continental crust. Most observations distinguish between azimuthal and transverse anisotropy. Azimuthal anisotropy produces shear-wave splitting and is detected on seismic refraction profiles recorded at different azimuths. Transverse anisotropy is produced by a hexagonal type of symmetry and is observed as dispersion of surface waves and travel time differences between shear waves vibrating in a vertical and a horizontal planes (*SV*- and *SH*-waves, correspondingly). The depth resolution in anisotropy studies is usually very poor.

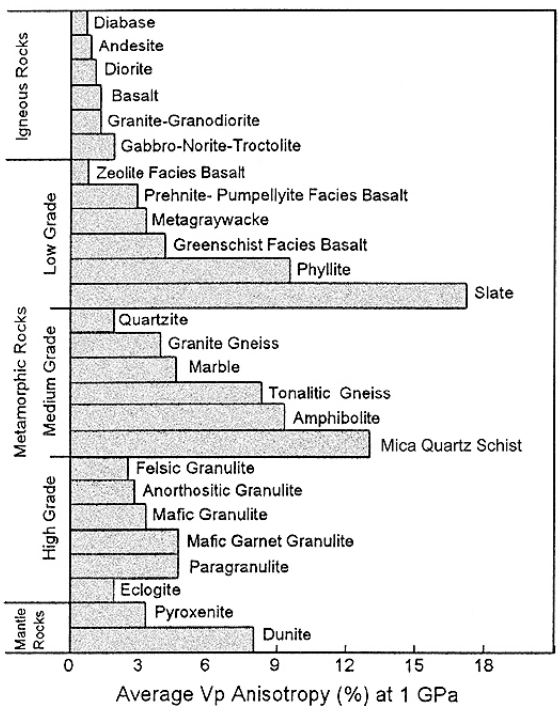

Figure 6. Average anisotropies $(V_{max}-V_{min})/V_{ave}100\%$ for major crustal rocks at 35 km depth. From Christensen and Mooney (1995).

The causes of seismic anisotropy of the continental crust are different at various depths. The principle mechanisms include: (1) preferred orientations of anisotropic rocks, (2) aligned cracks and micro-cracks, (3) periodic layering of thin isotropic layers of contrasting velocity. The latter causes only transverse anisotropy, while the first two mechanisms can produce both types of seismic anisotropy.

Igneous rocks are essentially isotropic (their anisotropy is less than 3%), while most of metamorphic rocks are highly anisotropic (the highest values are observed in slates and schists) (Figure 6). It implies that preferred orientation of anisotropic metamorphic rocks can be important in producing chiefly upper and middle crustal anisotropy.

Layering of thin isotropic layers plays an important role in producing seismic anisotropy in sediments. The major mechanism of seismic anisotropy in the upper 10-15 km of the crust is the presence of stress-aligned cracks and micro-cracks, which can produce as much as 4% difference in *S*-wave velocities. At greater depths, most of the cracks are closed and crustal anisotropy results from the presence of metamorphic rocks and/or the preferred orientation of anisotropic minerals aligned during ductile flow of the lower crust. In case highly anisotropic rocks (like schists) are present, the lower crustal anisotropy can be as large as 15% (e.g., below the Urals mountains). Lower crustal rocks which were transported to upper crustal depths by a tectonic process, may exhibit a "frozen-in" anisotropy caused by foliation and preferred orientation of anisotropic minerals (the Ivrea zone in Italy is an example). A part of lower crustal anisotropy can be attributed to a near-horizontal layering of felsic and mafic rocks, which produces lamellae-type reflectivity.

5.3. Poisson's ratio

Poisson's ratio is proportional to the ratio of *P*- to *S*-wave velocity $\sigma = 1 - 1/\left(V_P^2 / V_S^2 - 1\right)$ ($0<\sigma<0.5$) and is more sensitive to the composition of the crustal rocks than either *P*- or *S*-wave velocities. For common rock types σ varies from 0.20 to 0.35, increasing with a higher mafic content or with a lower quartz content (for quartz Poisson's ratio is only about 0.08). Because the *silica* (*SiO_2*) *content* in continental crust decreases with depth, Poisson's ratio typically increases with depth.

The average values of Poisson's ratio in continental crust depend on the crustal type. The highest values ($\sigma=0.29\pm0.02$) are characteristic of the Precambrian Shields and reflect a more mafic composition of their lower crust, compared to the platforms and orogens of the Paleozoic age (for both of which $\sigma=0.27\pm0.03$). Poisson's ratio clearly indicates the existence of two compositionally different crustal layers in the Precambrian shields and platforms with $\sigma\leq0.26$ in the upper crust and around 0.30 in the lower crust. Low ($\sigma<0.26$), intermediate ($0.26<\sigma<0.28$) and high values ($\sigma>0.28$) are characteristic, respectively, of felsic (quartz-rich), intermediate and mafic (quartz-poor) deep crustal rocks, which are indistinguishable from *P*-wave velocity alone.

The Mesozoic-Cenozoic orogens have much lower average values of Poisson's ratio than old tectonic provinces of the continents, $\sigma=0.25\pm0.04$, which is indicative of a predominantly felsic composition of the crust in young collisional orogens. The crust of the continental arcs has $\sigma=0.25$. Thus, the average values of Poisson's ratio in continental crust typically increase with the crustal age. The average Poisson's ratio for the bulk continental crust is 0.27, however some orogens have an unusually low $\sigma<0.20$, while some platforms and Paleozoic orogens have a very high ratio, $\sigma>0.36$.

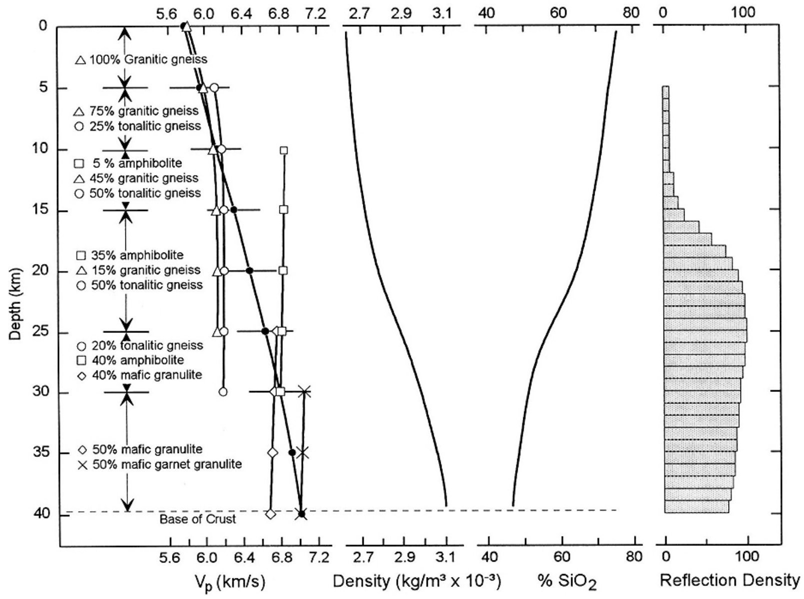

Figure 7. A model for average crustal petrology consistent with depth profiles for average *P*-wave velocity, density, silica content and reflection density. From Christensen and Mooney (1995).

5.4. Crustal density

The density profile of continental crust can be estimated from seismic data by an empirical relation between density and seismic velocity derived from laboratory measurements. The best simple correlation between density and *P*-wave velocity has the form of: $\rho = a + b/V_P$, where the numerical coefficients *a* and *b* vary slightly with depth. Typical values are, respectively, 4.93 g cm^{-3} and –13.29 g cm^{-3}/km s^{-1} for 10 km depth and 5.28 g cm^{-3} and –15.17 g cm^{-3}/km s^{-1} for 50 km depth.

The density of continental crust typically increases from about 2600-2800 kg m^{-3} at the surface to about 3100 kg m^{-3} at the base of the crust (Figure 7). The steepest growth of density occurs at the transition from the upper to the lower crust. The average density value assumed for continental crust is about 2830 kg m^{-3}.

5.5. Crustal rheology, brittle-ductile crust

The continental crust is usually considered to have two domains: brittle and ductile. Deformations caused by external stress can cause fracturing of the upper, brittle, layer of the crust and result in earthquakes. The temperature regime, crustal composition and pore fluids control the thickness of the brittle crust. Theoretical studies show that on geologic times the continental lithosphere cannot support elastic stresses at temperatures above approximately 350^{o} C. Thus, to a first approximation the transition from the brittle to the ductile deformation coincides with this isotherm, the depth to which can vary from about 10 km in young tectonically active regions to 50 km and more in cold cratons. The lower, warm, part of the crust can sustain essential deformations and exhibit a ductile flow under tectonic stresses. The majority of the crustal earthquakes occur in the upper, brittle, crust.

Crustal composition and the presence of pore fluids can shift the depth of the brittle-ductile transition."Soft" rheology of the crust is controlled by the presence of wet quartz, while "hard" rheology is controlled by dry Ca-rich plagioclase. In the case of "soft" rheology, the thickness of the brittle layer varies from <10 km in "hot" tectonic regions (with surface heat flow of 90 m W m^{-2}) to 15-20 km in "cold" regions (with heat flow of 50 m W m^{-2}), but not 50 km as expected from the thermal regime of the crust alone). In case of "hard" rheology, the brittle-ductile transition shifts to greater depths. *See Terrestrial Heat Flow, Geomagnetism and Geoelectricity* for the discussion of thermal, electrical and magnetic properties of continental crust.

6. COMPOSITION OF CONTINENTAL CRUST

6.1. Methods of estimating crustal composition

Only the uppermost part of the continental crust is accessible for direct studies and, thus, a series of indirect methods for estimating crustal composition were developed.

Seismic models are based on seismic refraction data that provide seismic velocities and Poisson's ratio along seismic profiles, laboratory measurements of ultrasonic velocities in rocks and minerals, and velocity-density relationships. However, the correlation between seismic velocity and composition is essentially non-unique, which implies a non-uniqueness of compositional interpretations.

Seismic velocities in the crust can be matched by an assemblage of igneous rocks and the velocity increase with depth can be explained by a gradual change from granite/granodiorite to gabbro. Alternatively, seismic velocities in the upper crust can be fit by metamorphic rock compositions and an increase in seismic velocities at mid-crustal depth can be explained by a progressive metamorphism. High velocities in the deepest crustal layers can originate from a transition from a predominantly mafic lower crust to mafic garnet granulite.

Petrologic models are based primarily on detailed studies of petrology, geochemistry and isotopic ages of sediments, basements outcrops, crustal xenoliths and exposed cross-sections of the deep continental crust. The best-known petrologic model is the so-called andesite model of continental crust, which implies a more sialic bulk crustal composition than suggested by recent seismic models (56-57% of SiO_2 versus 59-62%).

Geologic models, based chiefly on studies of exposed basement rocks, which can be non-representative of the middle and especially lower crust, seem to overestimate the sialic content in continental crust (63-64% SiO_2) and suggest the most felsic compositions of the bulk crust.

The most reliable and accurate recent models of crustal composition are based on seismic interpretations, petrologic studies and heat flow models, which permit to narrow the range of possible compositions of the deep continental crust. These models incorporate as well data on the plate tectonic processes responsible for the formation of the continents. For the reasons discussed below, Archean and post-Archean crustal compositions are determined independently and considered separately. The silica content in the bulk continental crust is estimated to be about 62%.

6.2. The upper and middle continental crust

The composition of the upper continental crust is relatively well known from sampling basement outcrops, especially in the shield areas, geochemical studies, and studies of exposed crustal sections. The upper crust is felsic in composition (the silica content is about 70% in the upper 5 km of the continental crust and decreases with depth).

Seismic velocities in the upper crust (<6.4 km s^{-1}) are characteristic primarily of gneisses, granites and granodiorites; however medium-grade metamorphic rocks (e.g. quartzite, slates and phyllites) have similar seismic velocities and are often present in the exposed rocks in shields. Usually it is accepted that the average composition of the upper continental crust is similar to a granodiorite.

Seismic velocities in the middle continental crust between a depth of 15 and 30 km are in the range 6.4 to 6.8 km s^{-1}. The increase in seismic velocity from the upper to the middle crust is usually associated with a larger volume of mafic mineral assemblages in the middle crust. Many cross-sections show that at these depths a transition in metamorphic grade from amphibolite facies to granulite facies occurs. Usually the middle crust is modeled as a mixture of mafic, intermediate and felsic amphibolite facies gneisses.

6.3. The lower continental crust

The direct study of the deep continental crust is possible only in few locations, where the crust originally at a depth of 20-30 km is now exposed (e.g., the Vredefort structure in South America, the Pikwitonei and Kapuskasing provinces in North America, an exposed crustal cross-section in Southern Norway, the Ivrea zone in northern Italy and an almost complete cross-section through a Variscan lower crust in Calabria, southern Italy), and by studies of lower crustal xenoliths brought to the surface by magmatic processes. However, such sampling may be biased because of the special tectonic conditions required to bring either lower crustal sections or xenoliths and high-grade amphibolite and granulite complexes to the surface.

At present it is generally accepted that the lower crust of stable continental regions has a mafic composition (the SiO_2 content is only about 47% near the Moho), though some studies suggest that in some non-cratonic regions it may be comprised mostly of rocks of intermediate silica composition. The bimodal distribution of seismic velocities in the lower crust (with peaks at about 6.8-6.9 and 7.3-7.4 km s^{-1}) suggests that the deep crust can be sometimes constituted by layered sequences of felsic and mafic rocks. This conclusion is supported by seismic reflection studies, which show that the lower continental crust is typically highly reflective, and by reconstruction of the lower crust and upper mantle based on xenolith data, which indicate the presence of multiple mafic layers at these depths.

The average composition of the bulk lower crust is close to basalt; however its composition is highly heterogeneous and varies between different tectonic provinces. In cratons, the mafic lower crust may contain high-grade metamorphic rocks in the granulite and/or eclogite facies; high seismic velocities in the eclogite facies rocks make them seismically indistinguishable from the upper mantle.

7. CRUSTAL EVOLUTION

7.1. Hypotheses for the continental crust origin

The growth of the continental crust began at around 3.8-4.3 Ga and though its volume increased in time, the rate of the crustal production is gradually decreasing: it is about $6.1x10^{15}$ g/year over the past 4.0 Ga and only $2x10^{15}$ g/year at present. The crust has reached 50% of its present volume at about 2.5 Ga, 70% at about 1.7 Ga and 85% at about 500 Ma. However, the rate and the volume of crustal recycling into the mantle at different geologic times are essentially unknown.

Several mechanisms are considered as playing the major role in the formation of continental crust.

1. New crust is extracted from the mantle primarily at the continental and island arcs, where mantle melting above the wedge of the downgoing subducting slab is initiated by fluids from the slab.
2. About 5 to 10% of continental crust growth is produced by intraplate volcanism within the continents (in orogenic belts and continental rifts) due to intrusion and underplating of mantle-derived magmas at the base of continental crust.
3. A small fraction of continental crust is produced by accretion of oceanic terrain to continental margins.
4. Minor volumes of continental crust are formed by ophiolites (obducted oceanic crust).

The high rate of crustal accretion in the Archean suggests that different mechanisms of the crustal growth could have operated throughout the history of the Earth. Hotter temperatures of the Archean mantle imply that magmas derived from the mantle were formed at different pressure-temperature conditions than ever since and thus had a different composition. This is supported (a) by the worldwide difference in the composition of Archean greenstone and post-Archean cratonic sediments, which are thought to be representative of the bulk crustal composition, (b) by a voluminous production of komatiites and TTG (tonalite-trondhjemite-granodiorites) series, most of which have the Archean ages. Besides, melting in subduction zones in the Archean may not have occurred within the mantle wedge above the downgoing slabs, but within the slabs, producing magmas (and hence continental crust) of a different composition.

Studies of rare earth element distributions in the Archean and post-Archean crust suggest that the Archean crust was not strongly differentiated into crustal layers. Since the Archean, no systematic changes in the composition of continental crust have occurred. Thus, the end of the Archean (ca. 2.5 Ga) is an important boundary in the chemical evolution of continental crust.

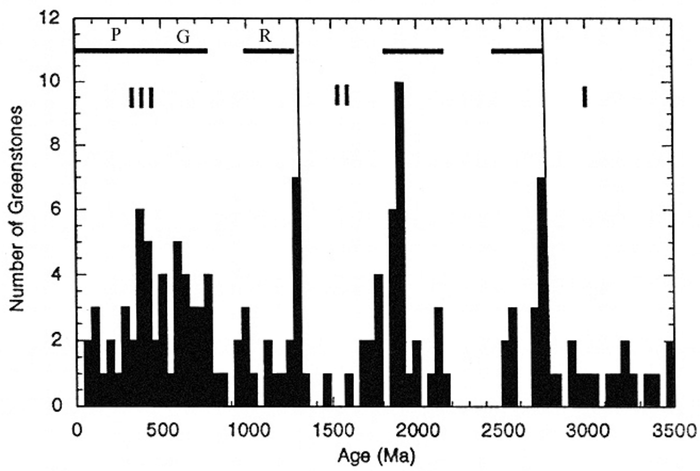

Figure 8. Histogram of greenstone eruption ages and proposed three stages in Earth history. Known or possible supercontinents are shown along the top (R=Rodinia, G=Gondwana, P=Pangea). From Condie (1995).

7.2. Age distribution of continental crust

The age distribution of juvenile continental crust appears to be very uneven with major peaks at about 0.3-0.5 Ga, 1.0-1.3 Ga, 1.7-1.9 Ga, 2.6-2.7 Ga and 3.2-3.5 Ga (Figure 8). The 2.7-2.6 and 1.9-1.7 Ga events are the times of major juvenile crust production. This result is rather surprising and unexpected from the models of a monotonous secular cooling of the core and the mantle, which would suggest a smooth distribution of the ages of the juvenile crust formed as the result of partial melting of the mantle. Three major explanations for the episodic age distribution of the continental crust can be considered: (a) incomplete sampling of the continental crust; (b) its uneven preservation; (c) non-uniform crustal generation throughout the geologic history.

New data on the distribution of crustal ages, accumulated during the past decades has not changed the main peaks at about 1.0, 1.9 and 2.7 Ga. Recent progress in the resolution of isotopic dating methods also has not smoothed the age distribution pattern of the crust. The gaps in the ages of greenstone belts at 1.35-1.65 Ga and at 2.2-2.5 Ga seem to indicate either that the plate tectonics did not operate prior to 1.35 Ga or that greenstones with these ages were entirely recycled into the mantle. However, the models of the monotonously cooling Earth cannot provide reasonable mechanisms for episodic turns on and off of the plate tectonics. A selective recycling of greenstones also seems unlikely, as other supracrustal rocks with the ages 1.35-1.65 Ga and at 2.2-2.5 Ga are widely preserved.

The existence of supercontinents appear to correlate with the peaks in the crustal ages (Figure 8), thus supporting thus the idea of the episodic growth of the continental crust. This seems to be the only possible explanation of the worldwide-observed age distribution of granitoids, greenstone belts, and orogenic events on different continents, which very well correlate with peaks in the production of the juvenile crust.

7.3. The formation of continental crust and mantle dynamics

The episodic formation of the juvenile continental crust may be closely related to the secular cooling of the Earth and the thermal processes in the mantle (primarily, mantle convection) and mantle dynamics. Studies of komatiites indicate that since the end of the Archean the average temperature of the Earth's mantle has decreased by approximately 200^o K. The model of the gradual transition from a two-layered (above and below 670 km) to the whole mantle convection associated with the secular cooling of the Earth may provide an explanation for the periodicity of the processes of the crustal formation and orogeny.

Higher temperatures and thus higher Rayleigh numbers of the Archean mantle would favor layered mantle convection. The phase boundary at the depth of 670 km (separating the upper and the lower mantle) is often assumed to play the key role in the evolution of the convecting system. At this transition zone the Clapeyron curve has a negative slope and will tend to resist the penetration of cold mantle material from above and hot material from below. Mantle convection may have been two-layered till ca. 2.5-2.8 Ga. An insufficient heat transfer across the phase boundary, where the subducted slabs were accumulated, may result in periodic massive overturns of the mantle, bringing very high-temperature lower mantle material to the shallow depths and thus producing voluminous melting at the upper mantle depths.

The overturns became more frequent as the Earth cooled, although the maximum temperature of the upper mantle decreased with each cycle. This means that melting of the mantle occurred at shallower depths than in Archean. The early-middle Proterozoic time marks the transition from layered to whole mantle convection. Some large overturns, which were likely to happen at this time, were probably associated with sinking of the lithospheric plates into the lower mantle. However, no catastrophic mantle overturns took place since Archean. Gradual evolution of the previous regime, associated with the Earth's cooling, resulted at about 1.3 Ga in the final transformation of the mantle convection from a layered to the modern whole-mantle convection regime. Seismic data does not support slab accumulation at the 670-km discontinuity at present and thus does not predict any catastrophic overturns of the mantle in the near future.

This model, relating the processes in continental crust with mantle dynamics, provides an explanation of the secular decrease in the volume of produced continental crust and episodisity of its formation.

Acknowledgements. Special thanks are due to W.D. Mooney for providing the information on seismic properties of continental crust. Valuable comments of W.D. Mooney and H. Thybo on the early versions of the manuscript helped to improve the quality and clearness of the presented material. The comments of O. Fitzpatrick and anonymous reviewers are greatly appreciated.

BIBLIOGRAPHY

Christensen N.I. and Mooney W.D. (1995). Seismic velocity structure and composition of the continental crust: A global view. *J. Geophys. Res.,* v. 100, B7, 9761-9788. [The paper discusses the composition of the continental crust based on the largest set of seismic velocity data and laboratory measurements of rock velocities and densities]

Condie K.C. (1997). *Plate tectonics and crustal evolution.* Oxford: Butterworth-Heinemann Publishers. [The monograph addresses most of the questions related to plate tectonics, the inner structure of the Earth and its evolution through the entire geologic history]

Crampin S. (1994). The fracture criticality of crustal rocks. *Geophys. J. Int.,* 118, 428-438. [The review of the worldwide evidence for crustal anisotropy]

Fountain D.M., Arculus R. and Kay R.W. (Editors) (1992). *Continental lower crust.* Elsevier. [The book contains papers which address seismic, electrical, thermal, magnetic properties of the lower crust, its rheology and composition]

Fowler C.M.R. (1990). *The solid Earth.* Cambridge University Press. [The advanced textbook on the global geophysics]

McLennan S.M. and Taylor S.R. (1999). Earth's continental crust. In: C.P. Marshall and R.W. Fairbridge, *The Encyclopedia of Geochemistry*. Kluwer Academic Press. [The article summarizes geochemical data on the composition and evolution of the continental crust]

Meissner R. and Wever Th. (1989). Continental crustal structure. In: D.James (Editor), *The Encyclopedia of Solid Earth Geophysics*. [The encyclopedia article which discusses seismic properties of continental crust]

Mooney W.D., Laske G. and Masters T.G. (1998). CRUST 5.1: A global crustal model at 5^ox5^o. *J. Geophys. Res*., v. 103, 727-747. [The global averaged model of both continental and oceanic crust based on seismic refraction data compilation]

Ranalli G. (1995). *Rheology of the Earth.* 2nd ed. Chapman & Hall, London. [The monograph which addresses all problems related to the Earth's rheology]

Rudnick R. and Fountain D.M. (1995). Nature and composition of the continental crust: A lower crustal perspective. *Rev. Geophys.,* 33, 267-309. [This review article reports recent geochemical results for the composition of the middle and lower crust]

Savage M.K. (1999). Seismic anisotropy and mantle deformation: what have we learned from shear wave splitting? Rev. Geophys., 37, 65-106. [The review article on seismic anisotropy of the crust and the mantle]

Taylor S.R. and McLennan S.M. (1985). *The continental crust: Its composition and evolution*. Oxford: Blackwell Scientific Publishers. [The complete summary of geochemical data on the continental crust]

Zandt G. and Ammon Ch.J. (1995). Continental crust composition constrained by measurements of crustal Poisson's ratio. *Nature,* v. 374, 152-154. [This article reports the most complete study of the Poisson's ratio for different crustal types]