



From Core to Crust: Igneous petrology

A tananyag elkészítését a „A Debreceni Egyetem fejlesztése a felsőfokú oktatás minőségének és hozzáférhetőségének együttes javítása érdekében” az **EFOP-3.4.3-16-2016-00021** számú projekt támogatta. A projekt az Európai Unió támogatásával, az Európai Szociális Alap társfinanszírozásával valósul meg.

SZÉCHENYI 2020



MAGYARORSZÁG
KORMÁNYA

Európai Unió
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BEFEKTETÉS A JÖVŐBE

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Manuscript closed: 27 November 2021

ISBN

Kiadja.....

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1 THE ORIGIN AND STRUCTURE OF THE EARTH

1.1 The origin of the elements

The observable part of the universe is built up by chemical elements, so if we would like to understand the genesis of our planet, our Solar System, and our Galaxy, and of course the meaning of life, the first step is to know this matter. The origin of the chemical elements is closely related to the birth and evolution of the Universe, thus – based on our recent knowledge – it is somehow connected with the **Big Bang Theory** (not the TV series of course). The Big Bang or sometimes called inflation theory is based on the theoretical and experimental studies of *Alexander FRIEDMAN* (born 1888, died 1925), *Edwin Powell HUBBLE* (born 1889, died 1953), *Georges LEMAITRE* (born 1894, died 1966), *Howard Percy ROBERTSON* (born 1903, died 1961), and *Geoffrey WALKER* (born 1909, died 2001). American astronomer, Edwin Powell HUBBLE was the first who showed the relation between distance and the velocity of the galaxies: while the galaxies are receding from our planet at velocities proportional to their distance. It causes a shift in the light emitted by far galaxies toward the red-light waves (called redshift). This phenomenon is described by the **HUBBLE-LEMAITRE Law**, and it can be regarded as the first observation of the expansion of the Universe, which means that the constituents of the Universe are moving away from each other.

The birth of the Universe began 13.8 billion years ago. However, the main part of the elements did not form directly due to the Big Bang, but a long time later, when the stars developed (**Figure 1**). And so, these element-creating processes still work in the stars. Based on the theory, very shortly after the Big Bang event there was no matter. The temperature was so high that it exceeded the 10^{31} K and the rapid expansion increased its volume by a factor of 10^{78} within very little time called **inflationary epoch** ($<10^{-32}$ s). The first particles appeared only after a few microseconds: electrons, positrons, neutrinos; and as the early Universe inflated and cooled, the first protons and neutrons also appeared. That is a question of how the matter still exists because according to the law of symmetry, equal amounts of matter and antimatter (positron, antiproton, antineutron) should have been created. Matter particles have the same mass as their antimatter counterparts, but their electric charge is the opposite. How it could have happened, is one of the greatest challenges in modern particle physics. It seems there is a kind of symmetry-violence or imbalance in nature resulting in the matter being more stable compared to the antimatter.

A few hundred seconds after the Big Bang nuclear reactions may have started resulting in deuterons. Almost all the deuterons finally combined to form helium, however, it should not be forgotten that most of the protons remained uncombined (hydrogen nuclei). After a few minutes, the Universe expanded and cooled down to that point (below 300 million K) when the nuclear reactions

stopped. While the Universe cooled, these physical circumstances could not make it possible to form any additional nuclei and the creation of the primordial elements ended. The final proportion of synthesized elements was about 75% hydrogen and 25% helium by mass, with very small amount of lithium (the Li/H ratio is about 10^{-9}).

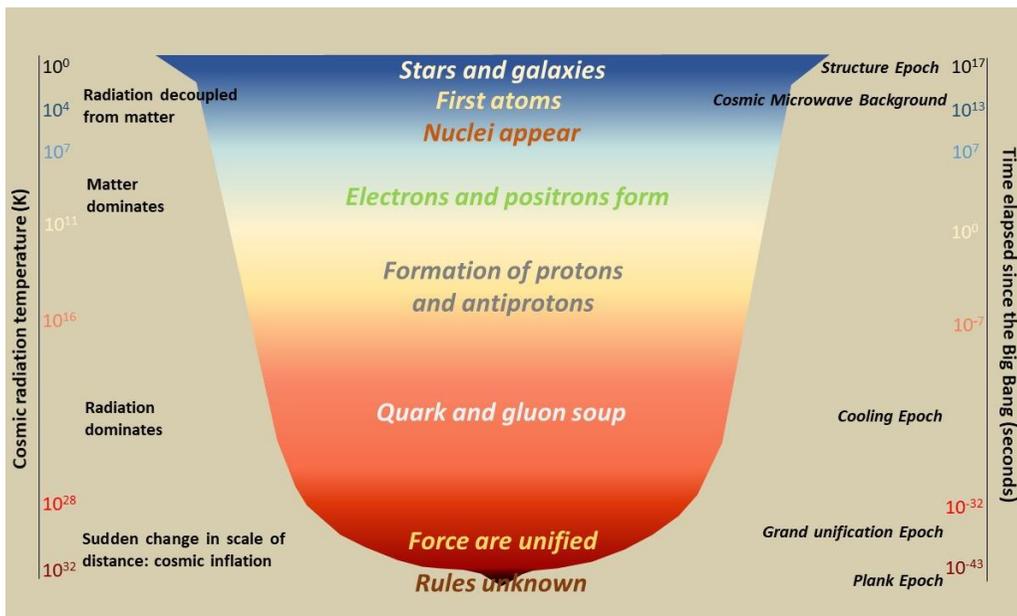
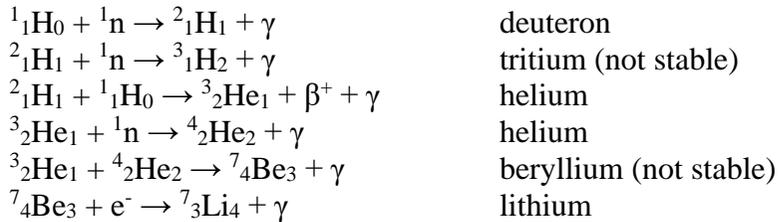


Figure 1 Timeline of the expanding Universe after the Big Bang

1.1.1 Nucleosynthesis and B²FH theory

As we have seen, only a few elements could have formed during the Big Bang at the beginning of the Universe, yet if we look at Mendeleev's periodic table, we can see that the number of natural elements is now close to a hundred. How could this happen? How did the natural elements form, and what kind of processes created them? We are looking for answers to these questions in this chapter, which is in one word: nucleosynthesis.

During the first explanations for the formation of elements in the 1930s and 1950s, researchers tried to find a process that could explain the formation of

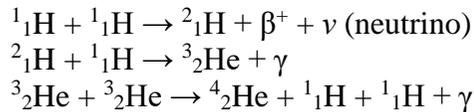
different elements alone. However, it later emerged that instead of a single process, several different mechanisms are responsible for the current abundance of the elements. Moreover, these processes took place in different places and at different times. This theory was first proposed by *Margareth*, and *Geoffrey BURBIDGE*, *William FOWLER*, and *Fred HOYLE* in 1957. This theory is also known as polygenic theory or B²FH. One of the most important pieces of evidence for the theory was the abundance of trace elements and isotopes observed in meteorites.

Briefly, the theory describes the formation of elements in four different stages. First, the **(1) cosmological nucleosynthesis** took place shortly after the Big Bang, and that is when some of the hydrogen, helium, and lithium atoms were also formed. Although helium is the main product of nuclear reactions in main sequence stars, it has not been enough to significantly change the cosmic abundance of helium (and of course hydrogen) over the past billions of years.

During **(2) stellar nucleosynthesis**, additional light elements other than lithium and beryllium, and a small part of the heavy elements are formed in larger stars, in their final stages of life. The next one is **(3) explosive nucleosynthesis** which is related to supernova explosions and creates heavier elements. Finally, **(4) galactic nucleosynthesis** taking place continuously in interstellar space through the interaction of cosmic radiation and matter is responsible for the formation of lithium and beryllium.

1.1.2 From cosmological to stellar nucleosynthesis

380 hundred years after the Big Bang the expanding universe's density and the temperature were much lower (below 3000 K), so the hydrogen, helium, and lithium nuclei were able to catch up free electrons and created neutral atoms. The radiation decoupled from the matter and the universe became "transparent" for the electromagnetic radiation. The well-known **Cosmic Microwave Background (CMB)** can be regarded as a remnant, as well as proof of this event. At that time, the universe was filled up with homogeneous hot gaseous material, however, very small differences observed in the CMB indicate that the matter was not utterly homogeneous. These inhomogeneities formed in the ordinary matter particles were free to react to the effect of gravity, resulting in the formation of the knot of the matter with a denser concentration, and finally, the first stars and galaxies were born. In the next stage of nucleosynthesis, the formation of heavier elements took place and is still happening in the stars. The energy for this process is covered by the thermonuclear fusion of hydrogen to helium atoms in the core of the stars. For the sake of simplicity, only the following reactions should be considered:



As a summary, we can conclude that four hydrogen atoms form only one ${}^4\text{He}$. For this reaction at least 10 million K temperature is required, otherwise, the protons are unable to overcome the electrostatic repulsion and get close enough to each other where the strong nuclear force will act (strong interaction). In our star, the Sun 600 million tons of hydrogen transform to helium in every second. Based on certain calculations the Sun will run out of hydrogen in its core within ca. 10 billion years, and as the Sun's age is roughly 5 billion years, it will work for additional 5 billion years. For a given point the star exists in a **dynamic hydrostatic equilibrium** (the star is not expanding or contracting), but this state will be changed when a major part of the hydrogen exhausts in the core a so-called gravitational collapse will begin because the outward thermal pressure will not balance the gravitational force. Due to its relatively low mass, our Sun will not explode as a supernova, however, a new balance will be established. During these processes, the core continuously contracts, while the hydrogen outside the helium-filled core starts to fuse and the outer part of the star expands and the Sun transforms to a subgiant first, then a **red giant star (Figure 2)**.

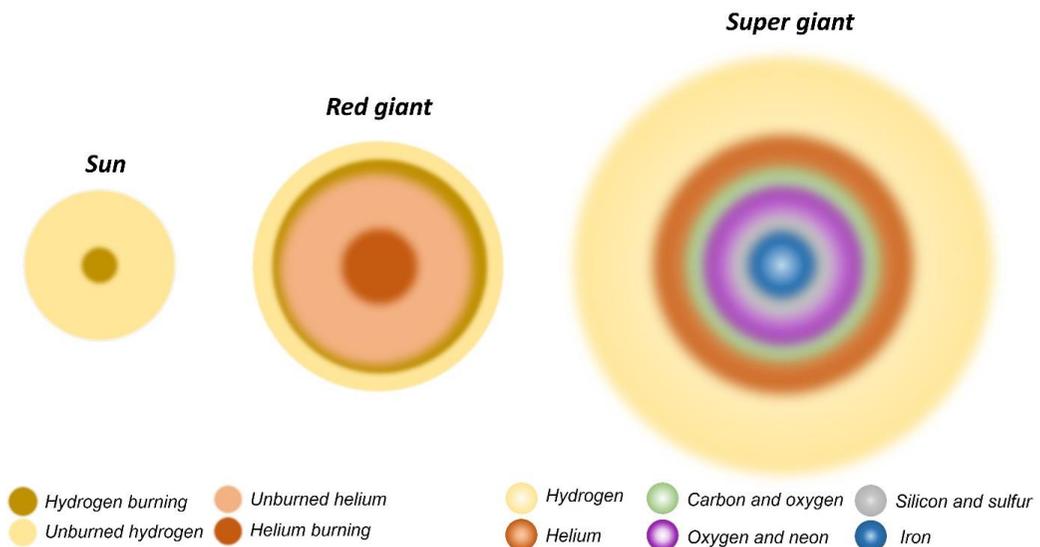
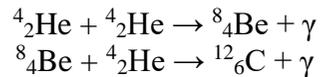


Figure 2 The Sun, a red giant, and a super-giant at various stages of its evolution

Calculations suggest that the diameter of this red giant Sun will be so large, that it will reach and engulf Mercury, Venus, and the Earth itself, too. The temperature and the density in the core will increase and when the temperature exceeds 100 million K the ‘burning’ of the helium will start. Here, the term ‘**burning**’ should be referred to as a **thermonuclear fusion reaction** of the given element. In our case, for the burning of the helium, the reaction can be given as follows:



During this reaction beryllium temporarily forms as it immediately reacts with helium. Additional reactions may also occur resulting in ${}^{16}\text{O}$, ${}^{20}\text{Ne}$, and ${}^{24}\text{Mg}$, however, light elements such as lithium, beryllium, and boron are not stable at that temperature. In Sun-like stars, further reactions do not occur. When helium disappears and the reactions ceases the core gravitationally collapses into a white dwarf, while the outer part dissipates forming a **planetary nebula** (**Figure 3**).



Figure 3 Helix Nebula (NGC 7293) is a planetary nebula in the constellation of Aquarius 650 light-years away from the Earth, which is an unravelled dusty layer of a red giant (credits: NASA, ESA, O'DELL, C.R. (Vanderbilt University), and MEIXNER, M., MCCULLOUGH, P. and BACON, G. (Space Telescope Science Institute, <https://esahubble.org>))

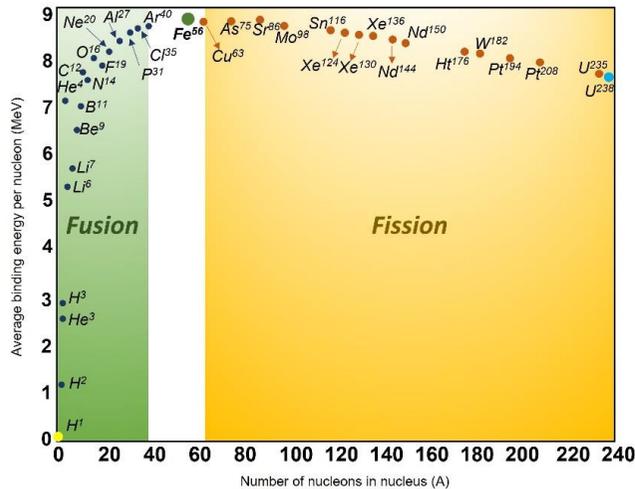
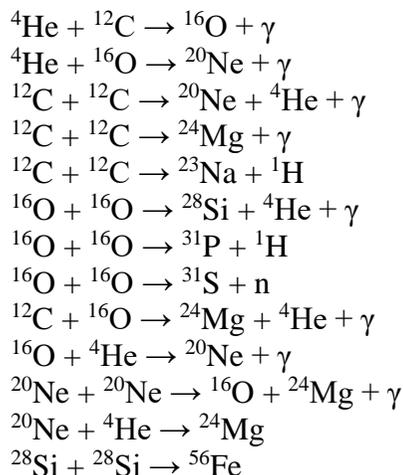


Figure 4 The amount of average binding energy in the function of mass number in MeV. Note that the binding energy is the highest for iron where the curve reaches its peak. It means that iron is the most stable element, neither fission nor fusion is possible. Elements with less mass than iron release energy through nuclear fusion, whereas elements with higher mass than iron release energy through nuclear fission (modified after: <http://www.earth-site.co.uk/Education/binding-energy/>)

In higher-mass stars, thermonuclear fusion reactions keep going, and when the mass is more than four times greater than that of the Sun the burning of carbon starts, as well as of neon, oxygen, magnesium, and silicon. These reactions release less and less energy thus their duration is also less and less. During silicon burning, iron is released. **Iron is the most stable nucleus (Figure 4)**. Some reactions of helium, carbon, neon, oxygen magnesium, and silicon burning:

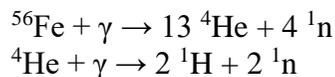


Note that the fundamental element of these nuclear processes is ${}^4\text{He}$, which explains the abundance of the elements with even atomic numbers (Oddo-Harkins rule).

The further fate of the star depends on its mass. If it is smaller than 1.4 solar mass it will end up as a red giant, while its core will become a white dwarf. In a **white dwarf** with a mass greater than this limit (called Chandrasekhar limit which is 1.4 solar mass), the gravity force overcomes the electron degeneracy pressure, and the atoms collapse as the electrons fall into the nuclei. The diameter of the core shrinks to ca. 100 km while its mass may exceed the mass of the Sun many times, and the star turns into a **neutron star** or **black hole**. The energy released during the core collapse is so high that the outer layer of the star explodes forming a supernova.

1.1.3 Explosive nucleosynthesis

As we discussed earlier the thermonuclear fusion or so-called e-process stops at the mass number of 56, thus the last nuclide which may form by fusion is ${}^{56}\text{Fe}$. The average binding energy per nucleon is the highest in this nuclide. In other words, thermonuclear fusion produces energy while the nuclides reach a mass number of 56. Above this number nuclear reactions show endothermic characteristics, namely, they consume energy. If the core of a star consists of iron mainly energy-producing reactions cease, and gravitational core-collapse begins leading to the spectacular death of the high-mass star (supernova explosion). In this case, if the mass of the star exceeds 1.4 solar mass, Coulomb repulsion cannot further prevent the gravitational collapse between the electrons. The supernova explosion begins with the sudden collapse of the star core, which shrinks from an Earth-like size to roughly 100 km in diameter in a matter of tenths of a second. When the density of matter in the core exceeds the density of the nucleus, the core ‘kicks back’ and sends a very strong shock wave outward. As the shock wave travels outward in the core at the extremely high temperatures formed, the nuclei decompose through the following photodisintegration processes:



Thus, a lot of free neutrons and protons are suddenly formed, which initiates another important process, the r-process. At the same time, the core emits a ‘**flood**’ of neutrinos due to electron-positron annihilation reactions. As the shock waves reach the margin of the core, they tear apart the outer layer. At the same time the burning of carbon, oxygen, and neon starts again, and isotopes of argon, calcium, chlorine, chrome, iron, sulphur, and titanium are formed.

Explosive or supernova nucleosynthesis represents the formation and scattering of heavier chemical elements in interstellar space through a supernova explosion. Very shortly before the explosion, the temperature in the outer layer reaches 3 billion K, free neutrons and protons are generated and a wide range of reactions occur, such as **slow** and **rapid neutron capture** (s- and r-processes), as well as **proton capture** (p-process).

During the s- and r-processes the remnant nuclei capture occasionally a hundred neutrons, thus unstable short-lived nuclei may form. Unstable nuclei emit β radiation (electron) through the stabilising process as the competent neutrons transform to protons. Another possible way for nucleosynthesis is when the nuclei capture protons resulting in elements with higher atomic numbers. These processes are responsible for the explosive nucleosynthesis and there are only a few minutes for all of this, just before the supernova explosion. Only the largest stars turn into a supernova. Those kind of stars are very unstable, and the total lifetime of such stars is sometimes only a few million years. The amount of the released energy is unimaginable: the most studied and documented supernova 1987A produced more energy within a minute than our Sun during 10 billion years.

1.1.4 Interstellar nucleosynthesis

It was mentioned earlier that the light elements, such as lithium, beryllium, and boron, are not stable at high temperatures prevailing inside the stars. These elements cannot even be formed in the stars, but rather are used in the nuclear reactions that take place. Lithium, beryllium, and boron are formed in interstellar space through the interaction of gas clouds and cosmic radiation. High atomic number nuclei in gas clouds formed during supernova explosions collide with high-energy particles of cosmic radiation and, as a result, fall apart into smaller nuclei and fragments. This explains the remarkably low cosmic frequency of lithium, beryllium, and boron. This process is called the x-process. And what about the supernova? The exploded outer layer of the star will become an interstellar gas cloud (nebula), and the remaining core will be a neutron star, or a black hole if it is large enough. In the interstellar gas clouds, which already contain the elements formed during the supernova explosion (including heavy elements), new stars and solar systems begin to form. Our solar system also emerged from such a cloud of gas, which is often referred to as solar nebula, or sun dust. Its material comes from at least 2-3 supernova explosions (**Figure 5**).



Figure 5 Composite picture of the three-light-year tall 'Mystic Mountain' in the Carina Nebula. The cloud is made up of dust and gas. Blue colour corresponds to oxygen, green to hydrogen and nitrogen and red to sulphur (credits: NASA, ESA, LIVIO, M. and the Hubble 20th Anniversary Team (STScI), <https://esahubble.org>)

1.2 Formation of stars and solar systems

Thanks to the space-based and surface-based large telescopes (such as Hubble Space Telescope), individual steps in star formation can already be observed directly. Stars form in clouds of matter in interstellar space, from their matter. These clouds are also called **molecular clouds** (suggesting that they are largely made up of molecular hydrogen), or **nebulae**, or **clouds of gas and dust**. These clouds of gas and dust are built up by hydrogen in approximately 70% by weight, while the remainder is helium. They also contain some heavier elements that were synthesized inside stars and scattered in space by a

supernova explosion or ejected by a red giant. Besides the gaseous material, these clouds also contain dust particles, and the spectroscopic study of various interstellar clouds shows that the ratio of dust to gas can be one in a hundred. In such dust clouds, stars are often born. Even an entire star cluster can form from a giant cloud.

Our Solar System also formed from such a cloud of gas and dust about 4.56 billion years ago. This cloud is called differently by various authors, it can be a nebula, a **presolar nebula**, or a **solar nebula**. These terms are largely used as synonyms. The heavy elements of our solar cloud (like the other clouds) derive from the material of earlier supernova explosions and red giant releases. Its composition was close to that of today's Solar System and may have been relatively homogeneous. Besides the gas, it also contained dust particles, the remains of which were found in the matrix of carbonaceous chondrite meteorites (see in the meteorite chapter). These particles consisting mainly of aluminium-oxide, hibonite, nano-diamond, silicon carbide, silicon nitride, spinel, titanium oxide, are the so-called presolar grains, cosmic dust, also called extra-terrestrial dust or space dust. Examination of so-called **calcium-aluminium inclusions** (CAI) in carbonaceous chondrites shows that short-lived radioactive isotopes were also present during the evolution of the early Solar System. The most important such kind of isotope is ^{26}Al , which has a half-life of only 0.7 million years. It means that these isotopes had to be synthesized and added to the material of the solar nebula no more than a few million years just before the Solar System formed.

At that time, the nebula temperature must have only been 10 to 20 K. The gravitational contraction and then collapse of the gas and dust cloud led to the formation of the Solar System. The contraction was probably triggered by a shock wave from a nearby supernova. During the contraction, the solar cloud warms, flattens, becomes more and more disk-shaped (protoplanetary disk), and most of its material gathers in the centre of gravity. The **protoplanetary disk** is an orbiting circumstellar disk consisting of gas and dust, where the initial rotation is enormously increased due to the conservation of angular momentum. The material gathered in the middle becomes the low-mass proto-Sun. As the proto-Sun continued to thicken, in a few million years it became so warm that thermonuclear fusion began in it, and finally, the Sun was born. Meanwhile, the protoplanetary disk rotates faster and faster around the central proto-Sun due to the contraction. The temperature in the inner part rises above 1000-1300 K, however, the outer parts are much colder. At that temperature, most of the dust grains in the cloud completely evaporate, and only the particles in the colder outer region, further from the Sun, may survive that kind of temperature.

Various objects of the Solar System, such as planets, asteroids, form from the rotating protoplanetary disk. The processes, which ultimately lead to the formation of the planets, are called **condensation** and **accretion**. During

condensation, the gaseous material of the disk precipitates in the form of small solid particles, and then the coupling of the precipitated particles into larger objects becomes the accretion.

1.2.1 Condensation

Condensation occurs at a pressure of about 10^{-4} atm, at such a low pressure the liquid state is not stable. The different elements of the gas cloud precipitate at different temperatures, i.e., they crystallise in the form of solid minerals. During condensation, the precipitating elements bind to each other and form compounds, or may react with other earlier precipitated solid phases, however, some elements precipitate in elemental form. To describe the order of condensation, one must become familiar with the cosmochemical classification of the elements.

1.2.2 Cosmochemical classification of the elements

The cosmochemical classification divides the elements into two main groups: 1) **refractory** and 2) **volatile** elements. Refractory elements are those that have a high melting point, so they are among the first to condense during the cooling of the hot nebula gas, or to remain in the solid-state for the longest time when it is heated. In the case of volatile elements, this is just the opposite, i.e., they condense at low temperatures.

In detail, rhenium, osmium, tungsten, zirconium, and hafnium can be distinguished as highly refractory elements (>1700 K). Refractory elements are (1500-1700 K) aluminium, scandium, calcium, titanium, most of the rare earth elements, thorium, and uranium. Moderately refractory elements, such as beryllium, magnesium, silicon, chromium, iron, cobalt, nickel, niobium, cerium, ytterbium, platinum, and palladium, condensate between 1300-1500 K. Copper, barium, manganese, strontium, phosphorous, and gold can be regarded as moderately volatile elements (1100-1500 K). Volatile elements (700-1100 K) are sodium, potassium, boron, sulphur, rubidium, caesium, gallium, tin, and selenium, while lead, zinc, thallium, indium, bismuth, oxygen, carbon, nitrogen, hydrogen, and noble gases represent highly volatile (<700 K) elements.

1.2.3 Condensation sequence

Calculating the condensation temperature of each element and determining the sequence of condensation is one of the greatest successes of thermodynamics. Evidence for condensation processes and the order of condensation can be found

in meteorites. From this point of view, calcium-aluminium inclusions (CAIs) in chondritic meteorites are particularly important.

First, the refractory elements condense (**Figure 6**). At temperatures between 1800 and 1300 K, platinum metals such as osmium, iridium, ruthenium, and rhenium and tungsten, condense. These elements can be found as tiny metal droplets in the CAIs of chondritic meteorites.

The high-temperature metal droplets are followed by calcium, aluminium, and titanium oxides and silicates forming calcium-aluminium inclusions (CAIs) in the chondrites. These elements precipitate as **corundum** (Al_2O_3), **spinel** (MgAl_2O_4), **perovskite** (CaTiO_3), **hibonite** ($(\text{Ca,Ce})(\text{Al,Ti,Mg})_{12}\text{O}_{19}$), **melilite** ($(\text{Ca,Na})_2(\text{Al,Mg,Fe})[(\text{Al,Si})\text{SiO}_7]$), and **anorthite** ($\text{CaAl}_2\text{Si}_2\text{O}_8$), and they can be regarded as the first and oldest condensates of the Solar System. Based on their age, they crystallised in a very narrow time interval, 4567.30 ± 0.16 million years ago. Additional refractory elements may also condense at this stage, such as uranium, thorium, zirconium, barium, and rare-earth elements.

The next step is the condensation of metallic Fe and Ni, which are also found in chondritic meteorites. Simultaneously, magnesium and silicon precipitate in the form of olivine and pyroxenes. In the case of equilibrium, the already condensed components can react with the vapour. For example, reaction with metallic iron may result in iron-rich olivine and pyroxenes.

Between 1200 and 900 K, moderately volatile elements are formed, such as alkali metals, sodium, potassium, rubidium, zinc, copper, and gallium. Sodium reacting with the earlier condensed anorthite results in Na-plagioclase, while sulphur reacts with iron to form iron sulphide called **troilite** (FeS), and at even lower temperatures with oxygen it forms **magnetite** (Fe_3O_4).

Between 800 and 500 K, volatile elements such as lead, thallium, mercury are condensed. Silicates can react with water vapour to form aqueous phyllosilicates. Below 300 K, highly volatile elements, such as nitrogen, hydrogen, carbon, and noble gases also condense.

The temperature in the protoplanetary disk is uneven: it decreases when we move away from the proto-Sun, so the volatile elements do not condense in the inner part of the Solar System. In the outer part, condensation continues at much lower temperatures. The distance beyond which it is already cold enough for the water to condense is called the frost line, or ice line. This border may have been somewhere around 2.77 astronomical units, which corresponds to the distance of today's *Ceres*.

1.2.4 Accretion: from condensates to planets

The question is, how could these tiny, much smaller than a mm, condensed particles stick together into planet-sized objects? This process is called accretion and can be divided into three stages.

A) Accumulation of condensed small particles: The colliding tiny particles clump together under the influence of electrostatic and van der Waals forces and form into larger clusters. This adhesion was faster beyond the frost line as the ice condensed on the pre-existing particles. Through the process, cm-sized objects were created.

B) Formation of multi-kilometre objects: This part of the accretion process is not well understood. We do not know through what kind of interactions cm-sized particles could have assembled into multi-kilometre-sized objects.

C) Formation of planetesimals: Gravity was already affected on the multi-km objects, therefore protoplanets with a size of 100 to 1000 km in diameter, so-called planetesimals, were formed due to these collisions. According to various authors, the Kuiper Belt may consist of such remnant planetesimals or planetary embryos.

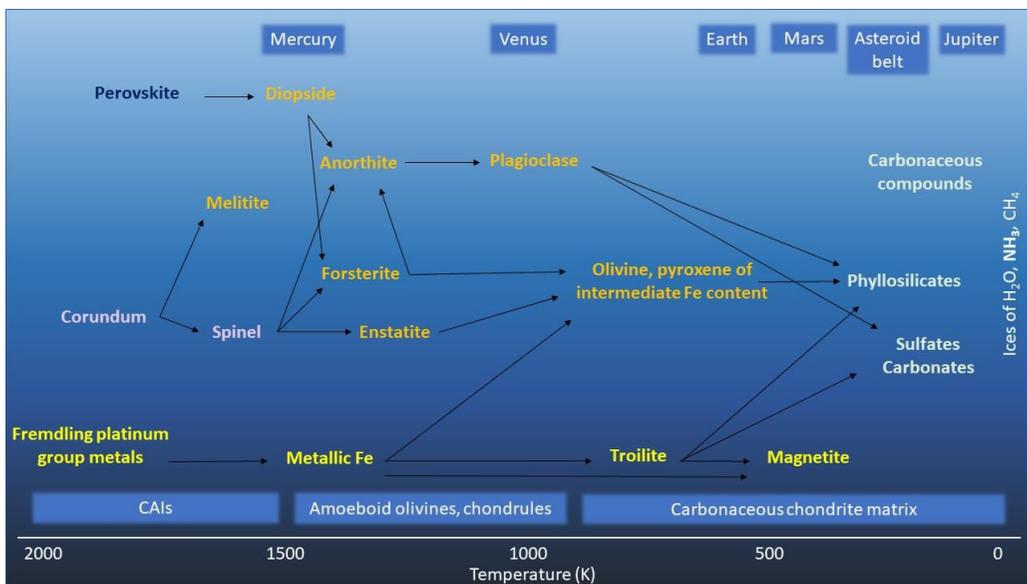


Figure 6 Simplified condensation sequence of minerals from a nebula with a composition of our Solar System. As the temperature is decreasing the minerals react with the gas or vapour to form new minerals at a lower temperature. The arrows indicate the direction of these reactions. Capital letters show the position of the planets and other objects as M: Mercury and Mars, Ve: Venus, E: Earth, A: Asteroid belts, J: Jupiter

The planetesimals formed into planets with further growth. Each growing planet completely cleared its orbital area or growth zone by ‘ingesting’ the planetesimals there. The terms **planetesimal** and **protoplanet** are not consistent in the literature. Some authors refer to objects with a few kilometres

in diameter as planetesimals, while those with hundreds of kilometres in diameter are called *protoplanets*. Other authors and literature, as in this one, use the terms planetesimal and protoplanet as synonyms, with a meaning of a celestial body with the size of hundreds of kilometres.

In the early Solar System, the growing planets may have collided with each other. Traces of these huge collisions can still be seen on the surface of the planets nowadays. For example, the anomalously large iron core of Mercury may be a result of an interplanetary collision that ‘shaved’ off a significant portion from the outer silicate mantle. The extremely slow and retrograde rotation of Venus, as well as Uranus’s almost ‘horizontal’ axis of rotation, can also be the result of collisions. The formation of the Earth-Moon system is also likely to be linked to such a collision.

The main accretion phase of planet formation lasted for about 30 to 40 million years. The solar nebula cleared because the gases condensed on the gas and ice giants in the outer Solar System, and the solar wind swept the volatiles out from the inner Solar System. However, various planetesimals and smaller debris materials were present for a long time, causing heavy bombardment for about another 700 million years forming the cratered surface of the Mercury and the Moon.

1.2.5 Chemical differentiation

As mentioned earlier, volatile elements did not condense in the internal Solar System. During accretion refractory minerals began to grow. Uncondensed volatile elements were swept out of the inner Solar System by the increased radiation of the young Sun called the solar wind. Therefore, **rocky** or **terrestrial planets** formed in the inner Solar System.

In the outer Solar System, accretion was faster due to lower temperature and H₂O condensation. The growth of the **gas** and **ice giants** began with the formation of their core: this is how the inner part of these planets consisting of silicates and ice was formed. When the mass of the core reached approx. ten times the mass of the Earth, the gravity was already strong enough to capture and keep the gas, so the gas and ice giants grew. Thus, the chemical difference between the inner (Earth-like or rocky) and outer (gas and ice) planets is due to the difference in temperature.

1.3 The abundance of chemical elements in the Solar System

Determining the abundance of elements in the universe and even on Earth is by no means an easy task. For the determination of the composition, especially in the case of crust-forming rocks are available directly. However, the Earth has

undergone chemical differentiation, resulting in the formation of spheres with different compositions. Although the determination of their average composition is extremely important for understanding the evolution of the Earth, the abundance of the elements in the Solar System can only be inferred indirectly. Samples from different celestial bodies (Moon, Mars) provide additional knowledge, however, the starting point for determining the average composition of the Solar System is the study of the composition of the Sun and meteorites.

The abundance of the elements in the Solar System was formerly also called **cosmic abundance**. However, from the data available to us from meteorites and the study of the Sun, we can only conclude about our Solar System. The development of astronomical spectroscopy makes it possible to study the composition of other Solar Systems and even galaxies, and from these, we know that there are differences among the chemical compositions of various solar systems and galaxies.

Since the Sun contains 99.8% of the material of the Solar System by mass, its composition will in any case be characteristic of the composition of the entire Solar System. Another source of data is meteorites, including the most primitive types of the Solar System called CI chondrites. In the next, we will deal with the structure and composition of the Sun and the classification and study of meteorites.

1.3.1 The Sun

The Sun is the central star of the Solar System, and Earth and other planets revolve around it. The diameter of the Sun is 1.392 million km, which is 109 times the diameter of the Earth. The Sun-Earth average distance is 150 million km, a distance called an astronomical unit (AU). The estimated lifetime of the Sun is 10 billion years, and it has already reached half of it. It will become a red giant in 5 billion years. The Sun consists of 92.1% hydrogen and 7.8% helium expressed in atomic percentage, while the other elements come out only 0.1%. By weight, the same means 71% hydrogen and 27.1% helium, while 1.9% goes to the other elements.

The internal structure of the Sun is inferred from theoretical calculations and indirect observations. The innermost part of the Sun is the core, where hydrogen fuses to helium, so most of the helium can be found here. The rest of the Sun gets its energy from here. There are two spherical shells around the core, the radiation zone, and the convection zone. The lower part of it is the photosphere. That is the visible surface of the Sun, and above it, the chromosphere can be found. The schematic structure of the Sun is shown in **Figure 7**.

The composition of the Sun can be determined by spectral analysis. The source of light is the photosphere, but this gives a continuous spectrum. Absorption lines that can be used for analysis are formed in the chromosphere above the

photosphere. The gaseous elements in the chromosphere absorb radiation of the wavelength characteristic for them, therefore we see dark lines in the continuous spectrum (**Figure 8**) as spectral absorption lines. The wavelengths of dark lines are characteristic of the elements in the chromosphere, and the degree of absorption is characteristic of their quantity. By measuring the spectral lines, the chemical composition of the Sun (the composition of the chromosphere) can be determined. The thousands of dark lines that appear in the Sun's absorption spectrum are called FRAUNHOFER lines after the German physicist *Joseph von FRAUNHOFER* (born 1787, died 1826) who originally observed these features.

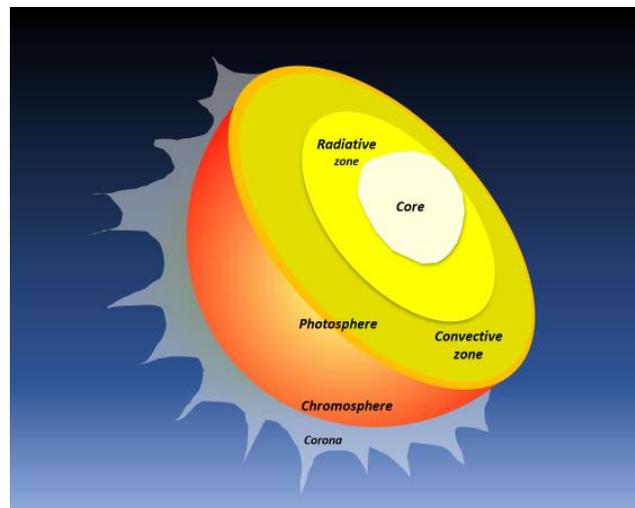


Figure 7 The structure of the Sun

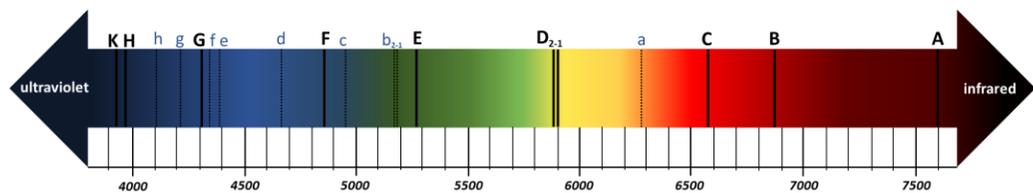


Figure 8 Absorption spectrum of the Sun. The dark lines are called Fraunhofer lines

1.4 Meteorites

To understand what a meteorite is, the first step is to clarify its concept. As we will see, meteorites are in some way related to the so-called meteoroids, meteors, and last but not least asteroids, but according to recent research, surprisingly in some ways, it is related to artificial objects. So, to find out what a meteorite is, we need to know the concepts of meteoroid, meteor, and asteroid.

The other important factor, or rather the set of factors, that can be a significant key in the definition of the meteorite, knowing the circumstances of how the meteorite eventually came to be found. Where did it come from, when did it take its final place, how long did it stay there, etc.? The space object from which the meteorite is most often formed is the so-called **meteoroid** (note that a meteorite can also be formed from a comet or a small asteroid). Meteoroids are solid celestial bodies in the interplanetary space that are smaller in diameter than asteroids (meteor and meteoroid are Greek words with common roots meaning ‘high in the air’). In the past, the lower limit of asteroids, along with the upper limit of meteoroids, ranged from 10 to 50 m, while other authors considered objects less than 100 m orbiting in space to be meteoroids (for more detail see MCSWEEN, 1987).

The name **asteroid** is of Greek origin and means ‘star-like’. Today, thanks to the higher resolution of instruments and advances in technology, the boundary between meteoroids and asteroids have been conventionally reduced to 1 m. The reason for this is that the more sophisticated astronomical telescopes already allow detecting objects of that size, and when the Sun's rays reflect on them it gives an image as if they were stars. Of course, in addition to the fact that this is a conventional and, by no means, a physical boundary, it should also be considered as an approximation, namely 1 m is not ‘carved in stone’. Anyway, if the celestial body is larger than ca. 1 metre, it is already called an asteroid. There is also a lower limit for the size of meteoroids, which is 30 μm because particles smaller than this can radiate heat well enough not to evaporate when they reach the atmosphere. So, the size of meteoroids is mostly between 30 μm and 1 m. Particles less than 30 μm are collectively referred to as interplanetary dust (IPD).

The next term we need to clarify is **meteor**. It is nothing more than a physical phenomenon of light (which may be accompanied by other physical phenomena) caused by solid objects, meteoroids, asteroids, comets entering the atmosphere. If this visual phenomenon is unusually bright, it is called a *fireball*, while if the flight through the atmosphere is accompanied by a sound effect, in the form of the so-called sound explosion, such a meteor is called a *bolide*. When a solid object strikes a celestial body that has no atmosphere and is accompanied by a light phenomenon, it is called an *impact flash*. Impact flashes can sometimes be seen on the surface of the Moon, with a good telescope!

The most common interpretation – and almost all definitions are the same – is that a meteorite is a natural object, a solid piece of rock that fell from outer space to Earth, but we are not far from the truth when we say that many authors have many different definitions. While in the 19th century the question of where the solid body that came to the surface of our planet might come from was neglected, the authors of the 20th century had already made it clear that it did come from space. At that time, however, we were still there by definition that meteorites could only be found on Earth. This kind of “Earth-centricity”

essentially reverts in some definitions, but the breakthrough happened in the 1980s when it was pointed out that meteorites, alien rock fragments from space can occur not only on our planet but also on other celestial bodies. According to the International Astronomical Union (IAU) Commission F1 on Meteors, Meteorites and Interplanetary Dust, a meteorite can only be formed if it was a meteor before. And as it was pointed out earlier for the occurrence of this light phenomenon atmosphere is needed! Thus, according to the IAU, a meteorite is a natural rock of foreign origin that does not destroy and reach its surface when it falls on a celestial body with a sufficiently dense atmosphere. That is, heating up, glowing and ablation are very characteristic for meteorites as they pass through a dense atmosphere. A foreign object, a rock that crashes into a celestial body without atmosphere but is not destroyed, should be called impact debris, according to the IAU. However, agreeing with *Alan RUBIN* and *Jeffrey GROSSMAN* (2010) a meteorite can occur on any celestial body, regardless of whether it has an atmosphere or not. So, in short, meteorites are natural, foreign rocks that come from meteoroids or larger celestial bodies. Approximately 40,000 tons of rock material fall to Earth in the form of meteorites each year.

1.4.1 Taxonomy of meteorites

The nomenclature system of meteorites, as well as the classification of igneous rocks on Earth, is based on their mineral and chemical compositions (whole rock and O isotope), and texture. Nevertheless, the names of individual petrological groups are completely different from those learned for terrestrial rocks, even though some meteorites have a terrestrial equivalent in the petrological sense. As we shall see later, the formation and development of the nomenclature have taken a stand-alone path, which, of course, may provide little consolation to the reader wishing to become acquainted with meteoritics, but we must bear this tiny inconvenience.

The vast majority of meteorites formed under diverse physical and chemical conditions differ from the well-known terrestrial environments. That is the reason why their mineralogical and chemical composition, as well as the texture, may vary significantly from terrestrial rocks. And that is why the nomenclature and grouping of meteorites is different compared to terrestrial rocks. We have already learnt in the case of terrestrial rocks that the classification is established on similar composition and texture. This is also true for meteorites. However, it should be noted that the classification of meteorites has undergone significant changes in recent decades: while previously members of each group did not necessarily belong together in a genetic sense, i.e., meteorites of different origins and evolutionary histories could belong to a group until then, in the last few decades, the pursuit of genetic grouping has gained more and more ground. People have known meteorites for millennia: in many civilisations, we can find

traces of the meteorite itself or meteorite fall. Occasionally, the original meteorite stone was utilised in some form, such as making it a utility tool, while in many cases, the fallen stone itself was not found, but written records about the phenomenon were made. One of the best-known examples of the former is archaeological finds from 3000 BC suggesting the use of meteorite, long before the general spread of iron tools (the so-called Iron Age). No wonder that the ancient Egyptians, according to hieroglyphics, referred to iron meteorites as 'heavenly metal'. The ancient Greeks and Romans also knew meteoric fall as an event, but according to Aristotle's theory, it was considered an atmospheric weather phenomenon. Meteors and meteorites were also known to Islamic, Japanese, and Chinese cultures, as evidenced by individual religious relics (the holy stone of *Suga-Jinja* shrine in Japan, the *Kaaba* Stone of Mecca) or written memoirs (Chinese records). Considered the father of modern chemistry in the second half of the 18th century, *Antoine LAVOISIER* (born 1743, died 1794), together with two colleagues, studied a meteorite that fell in 1768 near *Lucé* (Maine, France) amid heavy thunder and a hissing sound. That was the first known case in the history of modern science when someone tried to determine the chemical composition of a meteorite. Although the *Lucé* meteorite fell on a bright day, in clear weather, and the peasants working in the fields witnessed the incident, a committee found that the rock found was not falling from the sky, but a molten pyrite-containing sandstone struck by a lightning. As can be seen from the above, neither the classical philosophical way based on Aristotelian traditions nor the very young scientific approach could find a correct explanation for the formation of meteorites until the end of the 18th century, when a lawyer who graduated legal and philosophical studies, but had a strong natural scientific vein, came up with a completely unusual theory.

The first known scientific work proposing extra-terrestrial origin for meteorites is linked to *Ernst CHLADNI* (born 1756, died 1827), a German lawyer physicist, musician. The idea of CHLADNI was considered ridiculous or rather a minor sacrilege at the time and harshly criticised as meteorites were thought to be kind of volcanic formations in which lightning struck. In any case, the book itself has attracted the interest of several natural scientists, and events not so distant in time were gracious: meteorites fell at *Wold Cottage* (1795, Yorkshire, England) and in *L'Aigle* (1803, northern France) and the information obtained by examining them in detail confirmed CHLADNI's theory.

The classical and well-known classification classifies meteorites into three major groups: I) stony, II) iron, and III) stony-iron meteorites. The origin of these ancient names dates to the 19th century and is related to studies of *Gustavus ROSE* (born 1798, died 1873), *Nevil STORY-MASKELYNE* (born 1823, died 1911), *Gustav TSHERMAK VON SEYSENEGG* (born 1836, died 1927), and *Aristides BREZINA* (born 1848, died 1909). This classification is widely used in public but, it has now been pushed back in scientific communities because it is no longer sophisticated in a scientific sense. The fundamentals of current

scientific and genetic nomenclature system used here and discussed in detail is related to *George Thurland PRIOR* (born 1862, died 1936) British and *Brian MASON* (born 1917, died 2009) a New Zealand mineralogist. Over the past forty years, the system developed by PRIOR, and MASON was modified with major or minor changes.

Meteorites can traditionally be divided into two large groups in terms of whether we have evidence at the time of their arrival on Earth, but more recently, additional subgroups have been proposed to express and nuance uncertainties around perception. We call **fallen** meteorites those whose entry into the Earth's atmosphere (i.e., the fall) was observed and documented. The earliest meteor fall, for which the date is known, and the material of the meteorite was found, was recorded on 19th May, AD 861, in Fukuoka Prefecture, Japan. The peach-sized L6-type chondrite meteorite is preserved in the *Suga Jinja* Shinto shrine to this day. Among the fallen meteorites we can distinguish various classes. **Confirmed fall**, in which case it can be proved beyond a reasonable doubt that the fallen meteorite has been found. In the case of **probable fall** meteorites, on the other hand, there is a certain, usually small doubt about either the meteorite found or the conditions of the fall, which does not indicate with 100% certainty that the meteorite is related to the fall event in question. If we think about it, this case can easily turn around if a meteorite is found months and years after the fall event.

Any meteorite that cannot be classified as a found meteorite is called a **find**. In such specimens, the connection between the fall event and the meteorite itself cannot be proved in the absence of adequate evidence: thus, in this case, the incident was not witnessed, no photograph, recording, or record of the fall was taken. It may also be the case that, for various reasons, the occurrence of the fall event itself may be called into question. Within the meteorite group 'find', the category of **find, possible fall** may include those specimens that appear to be related to a particular fall event, but there is not enough information to prove this one hundred percent. In short, there is some doubt that the questioned meteorite can be attributed to a particular fall event, but there can also be significant doubt about the occurrence of the fall event itself (e.g., misperception). In the next group of found meteorites (**find, doubtful fall**), the connection to the falling event is doubtful: the specimens belonging to it have a great doubt that they can be linked to a falling event, or even the fact of the falling event itself is highly questionable. Finally, it is a group of actually found (**find**) meteorites for which there is no or no credible evidence of the conditions of the fall.

Meteorites are primarily named after the geographical location where they were found. If the site cannot be accurately identified or geographically associated (most meteorites are found in deserts), the stone will be named after some smaller typical geographic region (e.g., Northwest Africa 156, Arabian Peninsula 001).

1.4.2 *Primary and secondary minerals in meteorites*

The basis for the taxonomy of meteorites is to classify meteorites that can be considered similar according to their individual properties, as well as other natural or artificial materials, living organisms, or even concepts. This makes it easier to express, understand and examine the relationship, similarity, and difference (which can be genetic or non-genetic) between the individual groups or the members. The well-known classical triple division of meteorites dates back to the age when science was busy describing each type and determining its main mineral constituents and chemical composition. But, like mineralogy and petrology, meteoritics began to develop rapidly in the 20th century, which is not surprising since the principles and research methods of these disciplines are common. As we learnt more about the internal structure of crystals and then about the rock formation processes in the Earth's crust and mantle, the related nomenclature systems evolved (e.g., STRECKEISEN's classification of igneous rocks, TAS diagram for volcanic rocks, metamorphic facies defined by Eskola). Eventually, this also happened with meteorites. Although the above-mentioned classical triple division was not suitable for systematising meteorites based on their formation, evolution, and development, due to its simplicity it is certainly suitable for creating some kind of framework for laypeople. Then in the 20th century, as our knowledge became wider and deeper, demand has risen for a scientifically established, but genetically sufficiently detailed classification. This venture then succeeded so well, that the nightmare of university students came true: more nomenclatures were born! It should be added that none of them are better or worse than the other, they are only made with a different approach, they are based on different ideas. WEISBERG et al. (2006) provided a very thorough summary of meteorite classification systems (**Figure 9**) and their evolution.

It is much more expedient than the above-mentioned classical triple division to share the meteorites into a **non-differentiated** and **differentiated** group. The former and large group formed by the so-called chondrites, while the latter is also commonly called achondrites. **Chondrites** are, in fact, meteorites that have spheres of a few tenths of a millimetre to a few millimetres in size in their texture, called **chondrules**. In Greek, *χόνδρος* /k^hóndros/ also means seed; it is interesting that in Greek the same term is used for cartilage at the end of bones. It may be somewhat confusing that not all undifferentiated meteorites contain chondrules, which can be misleading when grouped, but that is a fact that these meteorites are closest to the chemical composition of the Sun (if only non-volatile elements are considered) as they were least affected by transformation and differentiation. For this reason, it is more appropriate to call the meteorite group into which chondrites can be placed undifferentiated or unmelted

meteorites. Note, for ease of understanding, undifferentiated and chondritic meteorites are used in the same sense.

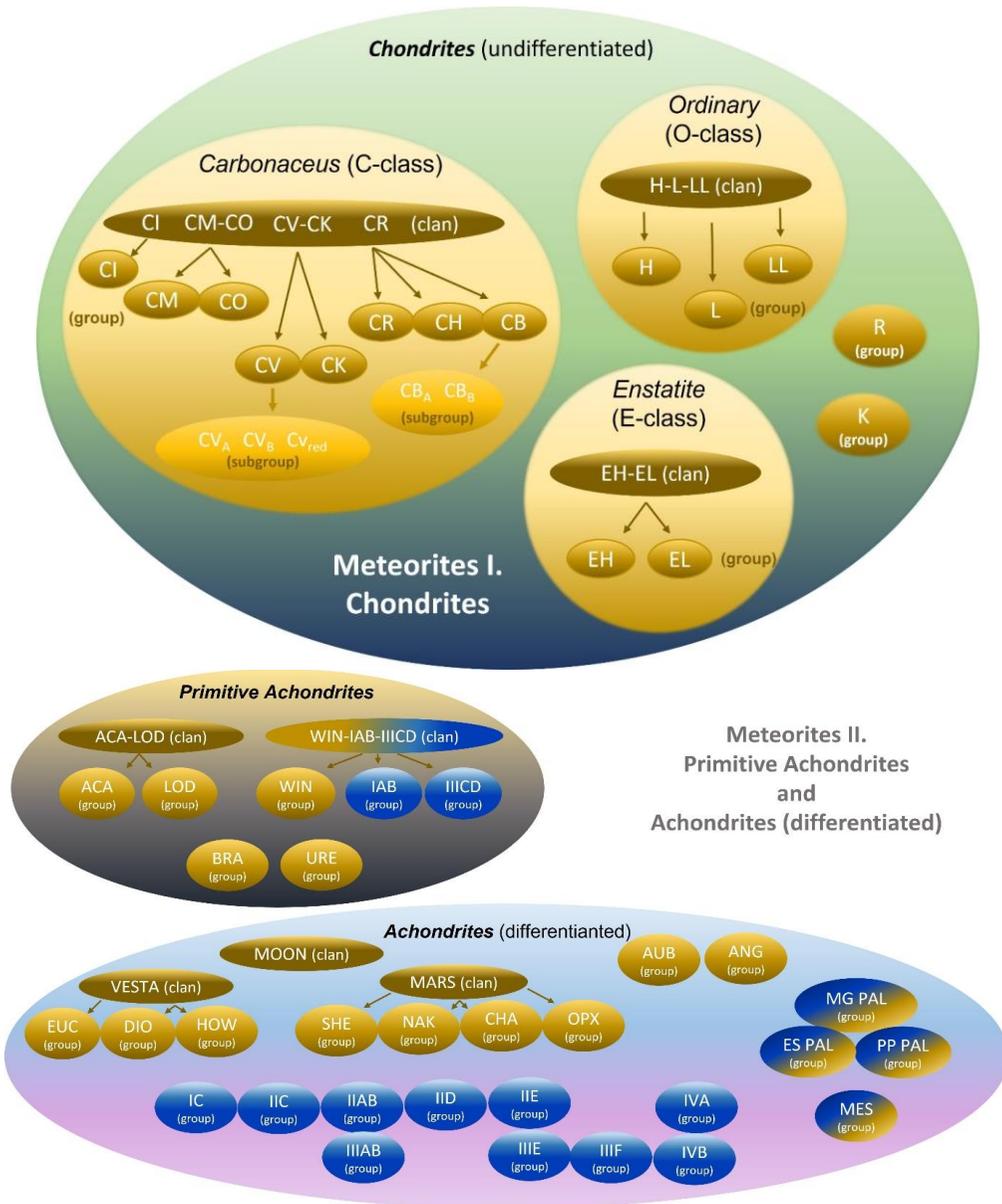


Figure 9 Systematics of the meteorite classification showing the major units, such as classes, clans, etc. Brown colour indicates silicate or stony meteorites, the blue colour refers to iron meteorites, while brown-blue composite shows stony-iron type

Based on terrestrial analogies, meteorites belonging to the **achondrites** (the 'a' is a privative prefix, i.e., achondrite means 'not chondrite') can be considered as igneous rocks, as they are formed by partial or complete melting and then recrystallisation of particles in the celestial body. Similar rock formation processes took place as we learnt in terrestrial igneous systems. This is also called igneous crystallisation or recrystallisation in meteoritics, so such texture is also called achondritic or igneous or recrystallised texture, as the process destroys the original chondritic features. These include meteorites from differentiated asteroids as well as Mars and Moon.

Besides the pure 'pedigree' meteorite types, gradual transition types may also occur in nature, making scientific research much more interesting. In some meteorite types, achondritic recrystallised texture can already be observed, however, their chemical composition still reflects the original, so-called **primitive**, or **primordial composition** ('primitive' is by no means a pejorative marker in this case, so we must mean that it is pristine). We call these meteorites primitive achondrites representing the transition between undifferentiated and differentiated meteorites, or if you like, between chondrites and achondrites.

A further disturbance in the taxonomy of meteorites may be that the nomenclature of chondrites, achondrites, and primitive achondrites have evolved independently for a long time. Nor can it be neglected that although the classification is based on texture, mineralogical, whole-rock composition, and O-isotope ratios, these characteristics are not fully consistent with each other, so there are some overlaps.

Classification based on the chemical composition of the whole rock is particularly prevalent in chondritic meteorites, where the composition is usually expressed normalised to magnesium or silicon and then compared to CI-type chondrites. As chondrites, except for some carbonaceous chondrites, always contain some iron-nickel alloys or iron sulphides, the amount of reduced (metal iron, iron sulphide) or even oxidised iron (iron oxides, iron in silicates) is also decisive. In addition, the proportions of differently volatile elements (e.g., Sb/Ni vs. Ir/Ni) may also be characteristics of each chondritic group. In the case of the classification of achondrites, the chemical composition is less pronounced, but they are particularly well suited for the monitoring of igneous processes, especially for eucrites. In iron meteorites, on the other hand, the emphasis is again on chemical composition. In this case, some trace elements, such as the concentrations of gallium, germanium, iridium, and nickel are considered. In conclusion, the current scientific classification can be used quite well, but let us have no doubts, it is far from satisfactory in all its details.

Another important feature of meteorite classification is hierarchy. Accordingly, in the nomenclature we can talk about classes, clans (also for some authors superclans, see KALLEMEYN et al. (1996), groups, and sometimes subgroups. For example, in the case of chondrites, the letter "C" is used to denote so-called

carbonaceous chondrites, while the letter "O" is used to indicate the nomenclature position of meteorites belonging to each taxonomic group. Carbonaceous chondrites are also characterised by the location of the naming meteorite in the code of each clan and group, for example, in the CI clan and group code, the letter "I" means *Ivuna* in south-western Tanzania, and in the CV group the 'V' *Vigarano* in northern Italy, in group CM 'M' *Mighei* in south-western Ukraine, in group CO 'O' *Ormans* in eastern France, etc.

1.4.3 Undifferentiated meteorites (chondrites)

Chondrites are the most primitive objects in the Solar System. Essentially, they can be regarded as sedimentary rocks: physical aggregates of components with various origins. These components formed in a solar nebula were placed randomly next to each other. Their name refers to their textural characteristic, or component: the presence of so-called chondrules (**Figure 10**). Chondrules are droplets of a few mm that were heated to about 2,000 K for a short time in their history and melted completely or, less frequently, only partially, then rapidly cooled, condensed, and solidified. Chondrules contain crystalline phases and sometimes glass, and their texture can be extremely diverse (**Figure 10**). Primarily crystal phases are olivine and pyroxene, however, they may also contain plagioclase (or plagioclase-like minerals) and, troilite (iron sulphide), and Fe-Ni metal in smaller amounts.

Additional components occur in chondrites, such as calcium-aluminium inclusions (CAIs, Fe-Ni metal and sulphide phases, and the fine-grained matrix. Calcium-aluminium inclusions (CAIs) are the earliest formations in the Solar System. Because they are high-temperature condensates, they are also called refractory inclusions. These are also mm or cm size sets of grains made up of minerals containing calcium and aluminium. Their constituents can be perovskite, melilite, spinel, anorthite, etc.

Metal and sulphide phases may also occur in chondrites forming shiny spots of several mm (**Figure 11**). The metallic phase may be Fe-Ni alloy or sulphide such as troilite (FeS).

The matrix is composed of a set of fine-grained olivine and pyroxene, which can be converted to phyllosilicates by aqueous conversion on the parent asteroid. In some types of chondrites, the matrix also contains organic matter.

Non-differentiated meteorites are classified into three major classes based on their composition. These are **carbonaceous chondrites (C)**, **ordinary chondrites (O)**, and **enstatite chondrites (E)**. Perhaps the biggest difference between these three classes of chondrites is that the meteorites belonging to them were formed at different temperatures. We know that chondrites do not represent wholly the elemental composition of the Sun, since significantly less of the volatile elements, such as oxygen or nitrogen, are not

preserved in chondrites, however they can be found in the Sun (remember the mass of the Solar System, more than 98% is found in the Sun). So, there are some differences in the chondrites compared to the average elemental composition of the Sun, but in this respect, we can also find differences between the chondrites themselves. In this respect, carbonaceous chondrites contain the most volatile elements, i.e., such meteorites are the least depleted of volatiles, but even carbonaceous chondrites do not contain as many volatile elements in proportions as those found in the Sun.

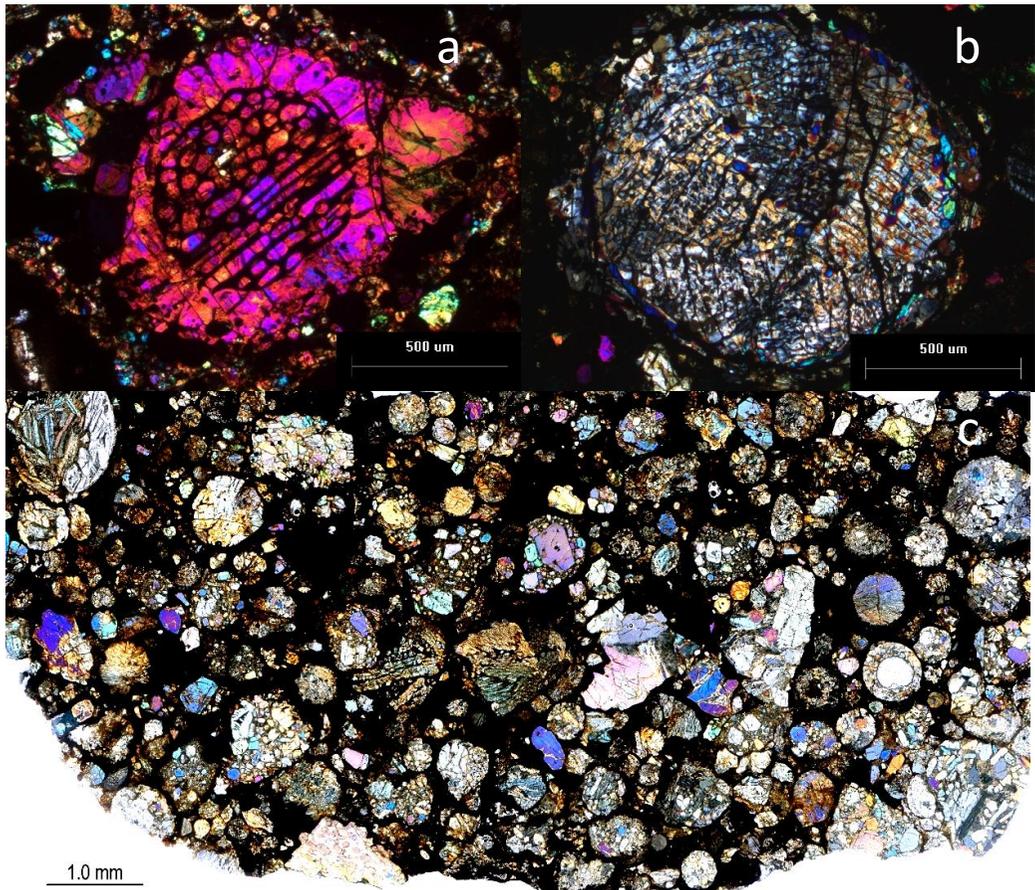


Figure 10 Photomicrographs of various types of spherical chondrules in the Koltsovo (a, b), and NWA 10826 (c) meteorites

Enstatite chondrites already contain fewer volatile elements compared to the carbon class, while ordinary chondrites are even more depleted. So, if meteorites formed at lower temperatures contain more volatile elements, the temperature of carbonaceous chondrites was the lowest at the time of their formation, that of enstatite chondrites may have been higher, while that of ordinary chondrites reached the highest value.

In addition to the concentration of volatile elements, another big difference amongst the three chondritic classes is the oxidation state. The oxidation state exhibits through the valences of the iron in the meteorite, primarily. Iron has variable valence. It is often said that elemental iron is not found in nature. Yes, if nature means the Earth's surface and its near-surface environments (let us say, to the best of our knowledge, the core of our planet is made of an iron-nickel alloy, too). There are also such physicochemical conditions in space where elementary, so-called metal iron can be formed, while in other cases iron forms a bond with oxygen atoms. In carbonaceous chondrites, iron forms a bond with oxygen, or sulphur and essentially can be found only in oxide, silicate, and sulphide minerals. In contrast, in ordinary chondrites, and even more so in enstatite chondrites, some of the iron is present in a reduced, elemental state (metallic iron).



Figure 11 Shiny spots of sulphide minerals in a slice of the Chug Chung 018 H5 ordinary chondrite meteorite

1.4.4 Carbonaceous chondrites

In the following, we will deal only with the most interesting group, the carbonaceous chondrites. As their name suggests, they contain carbon and organic compounds in their matrix. Carbonaceous chondrites have different

groups, named after the most prominent specimen. Let's consider an example through CV chondrites. C in the name refers to carbonaceous, and V refers to the most characteristic meteorite in the group, the *Vigarano* (Italy) meteorite. The famous Kaba meteorite from Debrecen, Hungary also belongs to this group. CI-type carbonaceous chondrites are the most primitive objects in the Solar System. This type is named after the *Ivuna* (Tanzania) meteorite, but its most important representative is the *Orgueil* meteorite that fell in France in 1864. Since about 20 kg of the fallen meteorite was collected (the meteorite fell into pieces), there is enough material for chemical analyses. As we will see later, the composition of CI-type carbonaceous chondrites reflects the composition of the Solar System the most.

1.4.5 *Primitive achondrites*

As it was mentioned earlier, the differentiated meteorites are derived from asteroids that have undergone a complete melting process, so the iron-nickel phase has been separated from the silicates. So those types of meteorites were once part of a large object, too, but the interiors of these objects melted so that the denser metals sank to the core and the lighter silicate components rose to the direction of the surface. This process is called **differentiation**.

Prior to the differentiation, **slow recrystallisation** (metamorphism) occurs due to heating up to 950 °C, and more, resulting in a gradual transformation of the chondritic texture. The chondrules' outlines gradually become blurred and then completely disintegrate upon diffusion. Also, as the temperature increases, chemical differences, primarily between the matrix and the minerals disappear and the meteorite achieves **chemical** and **isotopic equilibrium**. For example, that kind of equilibration process may result in changes in the Fe and Mg distribution of olivine and pyroxene. At the end of the process, the original chondritic chemical composition remains but the granular chondritic texture almost disappears.

Further increase in temperature already initiates partial melting processes. Over time, more and more relatively rare meteorites have been found all around the world, having a primitive chemical composition, or O-isotopic composition, while their original texture was lost completely. These meteorites are called primitive achondrites. Nowadays, these types of meteorites can be considered the starting point in the second phase of the thermal history of the chondritic meteorites. Primitive achondrites represent the last stage heating of the chondritic celestial body. Such meteorites are *acapulcoites*, *lodranites*, *winonaites*, *brachinites*, and *urelites* which may contain relic chondrules. These meteorites are extremely rare, except urelites. Some iron meteorites (IAB and IICD) may contain silicate rock fragments as inclusions, which by their composition (isotopic ratio) related to certain primitive achondrites, such as

winonaite. As these three meteorite groups are thought to be derived from the same parent body, they are classified in a clan together.

1.4.6 Differentiated meteorites

Differentiated meteorites are meteorites from the **asteroid belt**, the **Moon**, and **Mars**. That part of the meteorites is very diverse, as we can find here stony (achondrites), iron, and stony-iron meteorites, respectively. Iron meteorites are derived from the iron core of an asteroid, the stony-iron meteorites probably originated at the core-mantle boundary, and achondrites are derived from the silicate mantle and crust. Achondrites are essentially alkaline or ultrabasic igneous rocks. Some of them are true basalts resulting from melting and recrystallization, and they are analogues of terrestrial basaltic rocks on the surface of asteroids, or the Moon and Mars. '**HED**' meteorites (*howardites*, *eucrites*, *diogenite*) have originated from the crust of the asteroid Vesta, while the '**SNC**' group (*shergottite*, *nakhilite*, and *chassignite*) represent the crust of Mars.

Most of the iron meteorites also belong to differentiated meteorites. They are derived from the central part (iron-nickel core) of asteroids. They consist of iron (80-95%) with 5–20% nickel forming kamacite and taenite minerals mainly. Additional trace elements may also occur such as carbon, gallium, germanium, iridium, and sulphur. They regularly show *Widmannstätten* pattern in their texture, which is a kind of intergrowth of kamacite and taenite lamellae, however, at very low (<6%) or very high (>15%) nickel contents that kind of pattern does not develop. Thus, most iron meteorites have a specific texture and structure based on which three groups can be distinguished: 1) *hexahedrite*, 2) *octahedrite*, 3) *ataxite*. The nickel content of the groups differs. Hexahedrites have low 4-6% nickel content. Octahedrites contain 6-15% of nickel, while ataxites have high (>12%) nickel content. Hexahedrites, instead of the *Widmannstätten* pattern, often display parallel lines crossing each other at various angles (Neumann lines). These parallel lines are oriented by the kamacite hexahedrons. Octahedrites show *Widmannstätten* patterns, where kamacite and taenite lamellae oriented by a crystal structure of octahedron, intersect each other. The gap between the lamellae is filled with fine-grained kamacite-taenite assemblage called plessite. In ataxites (in Greek means without structure) due to the high nickel content *Widmannstätten* patterns are not developed. The Hoba, the largest meteorite in the world was found in Namibia in 1920. The Hoba weighs more than 50 tons, and it belongs to this structural class.

As nickel, germanium, and iridium contents correspond to the different parent bodies iron meteorites are classified by their concentration. Thirteen categories are distinguished based on these elements labelled with letters and numbers such as IAB, IIIAB, etc.

Stony-iron meteorites consist of metallic and silicate components in roughly the same proportion. The silicate phases are thought to represent the achondrite part of a planetesimal or asteroid, while the metallic phases are related to iron meteorites. Stony-iron meteorites are thought to represent the boundary of the core and mantle in differentiated bodies that included both core and mantle (**Figure 12**). They can divide into two primary classes: 1) **pallasites** and 2) **mesosiderites**.

Pallasites are composed mainly of metallic iron-nickel phases (Ni-poor kamacite with a Ni content of 5-10% and Ni-rich taenite with a Ni content of 20-65%), and yellowish-orange, sometimes greenish beautifully shiny, transparent, or opaque, cm-sized olivine crystals. They are accompanied by other silicate phases such as pyroxenes, or phosphates, such as farringtonite ($\text{Mg}_3(\text{PO}_4)_2$). The metallic phase regularly exceeds the proportion of the olivine phase. By etching their polished and polished surface with an acid-alcohol mixture characteristic *Widmanstätten* pattern emerges.

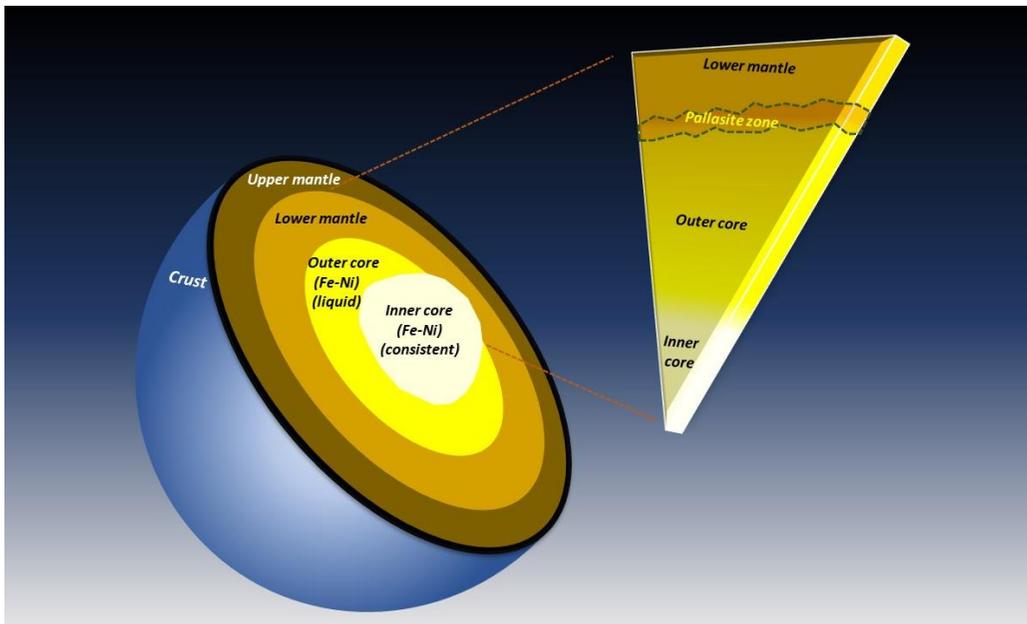


Figure 12 Pallasites may have formed a layer along the boundary of the lower mantle and the upper core of a differentiated parental body

Mesosiderites are composed mainly of metallic Fe-Ni alloys (more than 50%) orthorhombic pyroxene, and calcium-rich plagioclase. Accessory minerals may also occur, such as pigeonite, troilite, chromite, apatite. The brecciated structure is very common.

1.5 Elements in the Solar System

The abundance of elements in the Solar System can thus be approximated from two sides: on the one hand, based on the composition of the Sun's chromosphere, and on the other hand, based on the composition of CI carbonaceous chondrites. If the two compositions are compared, the similarity appears immediately (**Figure 13**). We see a difference only in the case of volatile elements (C, N), of which chondrites contain less, while the amount of lithium is smaller in the Sun due to different nuclear reactions.

The composition of both the Sun and the most primitive meteorites reflects the overall composition of the Solar System, the original composition from which the various objects in the Solar System were derived. Therefore, it is not surprising that objects in the Solar System, such as planets, represent chondritic patterns in their composition.

The relative abundance of the elements can be found in **Figure 14**. As it can be seen, the abundance of the elements shows a 'sawtooth' shape in the function of the atomic number, which is because the elements with even atomic numbers are more common with an order of magnitude than those with odd atomic numbers. This is called the **Oddo-Harkins rule** and shows that even-numbered nuclei are more stable.

The abundance of the elements decreases approximately exponentially as the atomic number increases, but there are very striking fractures in the shape of the curve. The most common elements are hydrogen and helium forming the vast majority of the mass of the Solar System. This is followed by oxygen, carbon, nitrogen, and neon.

Despite their small atomic numbers, lithium, beryllium, and boron are remarkably rare. The nucleus of these elements is not very stable (low binding energy). However, certain elements are remarkably abundant, such as iron, nickel, or lead. They have a stable core with high binding energy. These atomic numbers are also called 'magic numbers' in nuclear physics.

The abundance of some elements in the Solar System differs from the abundance in the Earth's crust (the 'normal' abundance). For example, platinum metal elements are not uncommon at all in the Solar System and even show a small positive anomaly.

Apart from volatile elements (hydrogen, helium, nitrogen, carbon, etc.), carbonaceous chondrites represent well the initial composition of the Solar System. The composition of meteorites can be measured much more accurately than the composition of the Sun.

The elemental abundance of carbonaceous chondrites is often used as a reference thus researchers regularly normalise the composition of other meteorites, as well as terrestrial rocks to carbonaceous chondrites, especially to CI chondrites.

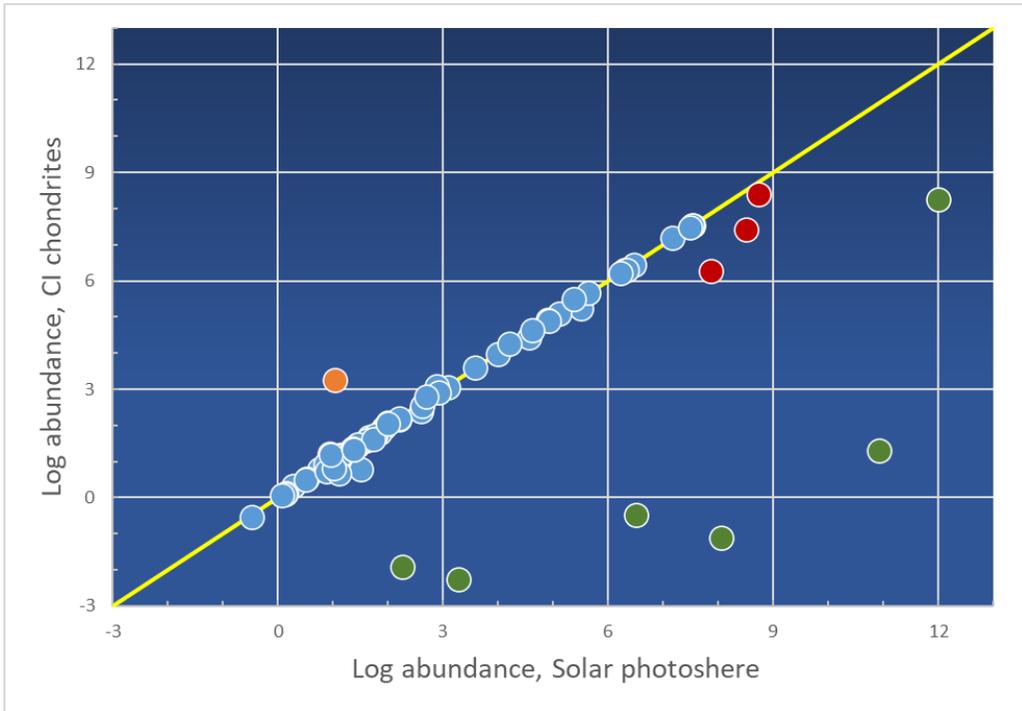


Figure 13 Plot diagram of the abundance of elements in the CI meteorite and at the surface of the Sun

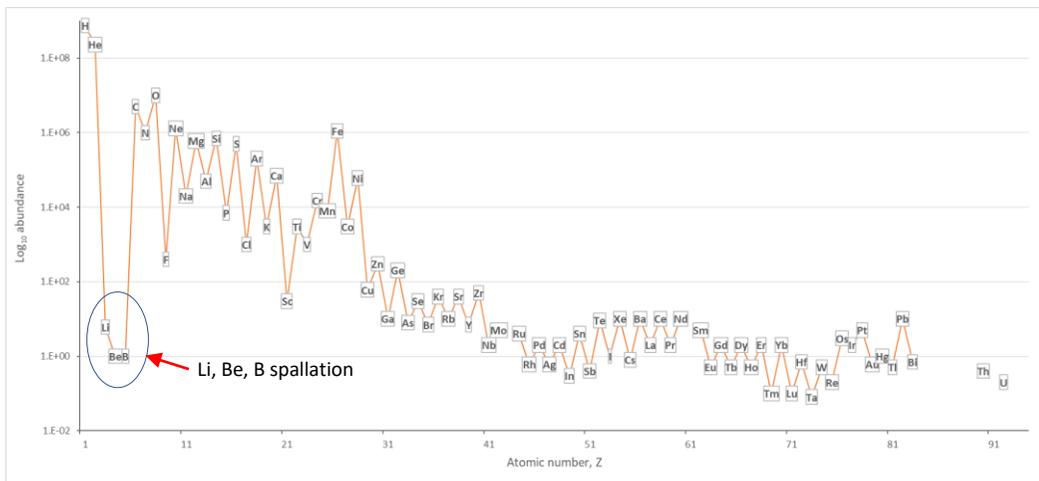


Figure 14 The Oddo-Hartken diagram showing the relative abundance of the elements

This comparison spectacularly shows how the composition changes compared to the ‘primordial material’ of the Solar System. Beyond the major elements, the trace element composition of various rocks and even minerals are regularly normalised to chondrites.

Figure 15 shows a rare earth diagram normalisation to chondrite. **Figure 15 A** (left) shows the rare earth element concentrations in CI chondrites and the average abundance of the particular rare earth element in the upper crust. The unit of concentration is in ppm (parts per million). The sawtooth-like shape of the curve is caused by the fact that the rare earth metals with the odd atomic number are rarer ca. with an order of magnitude than even-numbered ones. **Figure 15 B** shows the abundance of rare earth elements in the upper crust normalised to CI chondrites: for example, the rare earth metal concentrations in the upper crust were divided by the rare earth concentrations of CI chondrites. This normalisation, on the one hand, eliminates the even-odd abundance effect and, on the other hand, illustrates the change from an initial, primordial, or ‘primitive’ material. The figure shows that in the continental crust (more precisely in its upper part), the concentration of rare earth metals is significant. The heavy lanthanides enriched ca. one, while light lanthanides enriched by two orders of magnitude compared to chondritic values.

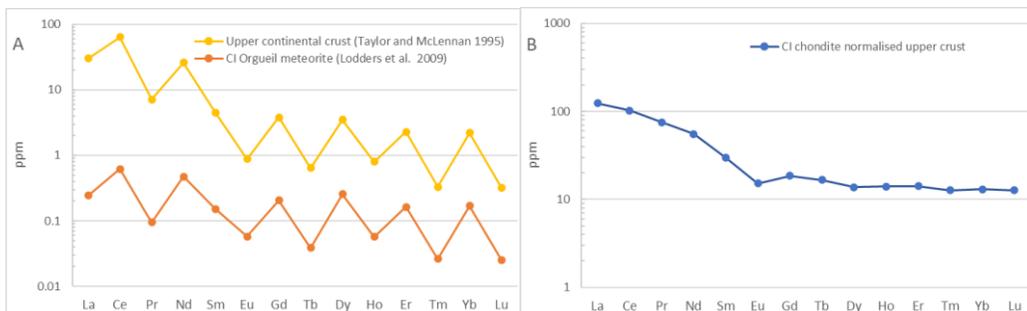


Figure 15 A) The abundance of rare earth elements in chondrites and the continental upper crust in ppm. **B)** Normalised rare earth element composition of the upper crust

1.5.1 Geochemical classification of elements

Geochemistry is a relatively young science, as geology itself is a young disciple. The term itself was first used by *Christian Friedrich SCHÖNBEIN* (born 1799, died 1868), a German-Swiss chemist, the inventor of the fuel cell, in 1843. An asteroid wears his name. The development of geochemistry is closely linked to the discovery of elements (most elements were already known by the end of the 19th century) and the development of analytical methods. Geochemistry is a

data-demanding science, and the discovery of optical emission spectroscopy in 1860 gave a significant thrust to further development.

Frank Wigglesworth CLARKE (born 1847, died 1931), an American scientist was the first major figure of geochemistry, or as he is often referred to: the ‘father of geochemistry’. He analysed rock, mineral, ore, and natural water samples in enormous numbers, and based on the result, examined the terrestrial distribution of the elements. He has significant merit in determining the average composition of the continental crust. His classic masterpiece *The Data of Geochemistry* first appeared in 1908 and then received several editions. In CLARKE’s memory, the mean values of the abundance of elements in the Earth’s crust are called *Clarke number* or *clarke* (although the term is now less commonly used).

The most significant figure in the advance of geochemistry is undoubtedly *Victor Moritz GOLDSCHMIDT* (born 1888, died 1947), a Norwegian mineralogist and founder of modern geochemistry. His name is associated with the X-ray analysis of crystal structures and the recognition of their significance in the distribution of elements, the application of chemical equilibria, the study of meteorites, the geochemical evolution of the Solar

System, the concept of geochemical cycle, element substitutions and the geochemical classification of elements. His masterpiece, *Geochemistry*, which is still in use today, was published in 1954 after his death. The most significant geochemical conference is named after GOLDSCHMIDT. The Goldschmidt conference is held annually in Europe and overseas (especially in America).

The Russian school of geochemistry has focused primarily on the prospecting and exploration of mineral resources. Its most significant figure is the Russian-Ukrainian *Vladimir Ivanovich VERNADSKY* (born 1863, died 1945). His name is associated with the recognition of the role of life in the distribution of elements (**Figure 16**).

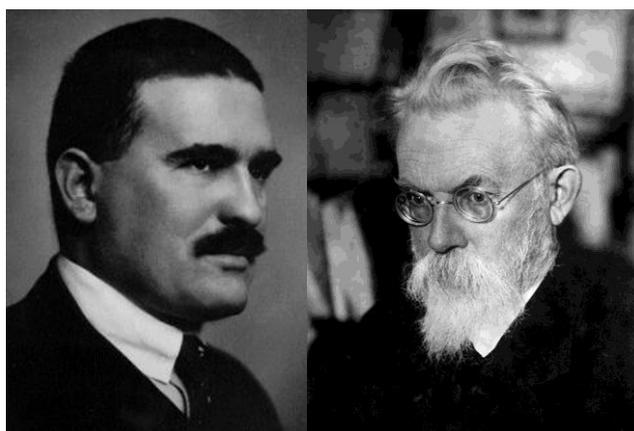


Figure 16 *Victor Moritz Goldschmidt, and Vladimir Ivanovich Vernadsky*

The fast-paced technological progress in analytical methods (e.g., mass spectrometry, neutron activation, etc.) and the quantitative approach played a fundamental role in the advance of modern geochemistry in the 1960s. Modern geochemical research includes the origin and evolution of the Earth and the Solar System, quantification of the geological time scale, determination of PT conditions of igneous and metamorphic processes, convection currents in the mantle and subduction of sediments into the mantle, rate of the orogenesis and erosion, temperature changes (paleoclimatology), formation and evolution of the atmosphere, but the earliest evidence of life on Earth (3.8 billion years ago) is also due to geochemistry.

Geochemistry, as its name suggests, is the connection between geology and chemistry. Various authors define geochemistry as the application of chemistry to solve geological problems, however, geochemistry is more than that.

Geochemistry is a discipline that studies the chemical composition of the Earth as a whole and the Earth's geospheres, such as crust, lithosphere, mantle, core, as well as the outer spheres, such as hydrosphere, and atmosphere. It examines the rules in the distribution and migration of elements in space (in different geological formations) and in time (in different geological ages). It is closely related to mineralogy, petrology, and, of course, chemistry. In a broader sense, geochemistry also includes the study of the chemical composition of the universe and the Solar System. this discipline is also referred to as cosmochemistry.

In nature, certain elements and groups of elements appear together, and their average abundance in different phases also shows a similar change. The joint appearance and similar behaviour in a given environment are based on the similarity of the geochemical properties of the elements. The **geochemical classification** of the elements has been carried out from very different perspectives by various authors. As a result, many systems have emerged that are more of a historical significance and are now obsolete. Goldschmidt's geochemical system is what is still widely used nowadays.

Goldschmidt set up his system based on the results of ore smelting processes and the study of the distribution of elements in different phases, also taking into account the abundance of elements in different meteorites.

1 Siderophile elements ('iron loving'): iron group and the platinum group elements are listed here as follows: gold, germanium, tin, (lead), carbon, phosphorus, molybdenum, rhenium. These elements enrich in the metallic melt during smelting and refinery processes, and predominantly attach to the Ni-Fe metallic phase in meteorites. Their affinity for sulphur and oxygen, respectively, is low. According to Goldschmidt, these elements are mainly enriched in the Fe-Ni core of the Earth. Strictly speaking, only iron groups (iron, cobalt, nickel), platinum metals (ruthenium, rhodium, palladium, osmium, iridium, platinum), and gold and rhenium belong here.

2 Chalcophile elements ('copper loving'): These elements are enriched in the metallurgical sulphide melt and, in the 'chalcosphere' assumed by Goldschmidt, in the oxide-sulphide belt. Some chalcophile elements are enriched in the troilite phase of meteorites. They are characterised by their high affinity for sulphur. These elements are copper, silver, zinc, cadmium, mercury, gallium, indium, thallium, lead, arsenic, antimony, bismuth, sulphur, selenium, tellurium. These elements and their enrichments are found primarily in sulphide ore deposits. It should be noted that in the Goldschmidt Earth model the existence of the oxide-sulphide belt, the 'chalcosphere', was proposed as a separate sulphide zone, but it has long been ruled out.

3 Lithophile elements ('rock loving'): these elements enrich in metallurgical slag and are also characteristic of silicate meteorites and the solid mantle and crust of the Earth (the so-called Silicate Earth). These include in particular alkali metals, alkaline earth metals, some transition metals, rare earth elements, aluminium, and silicon. Lithophilic elements are characterised by a high affinity for oxygen.

4 Atmosphile elements ('vapour loving'): Elements that accumulate in the atmosphere belong to this group.

Some authors distinguish a group of biophilic elements (C, H, O, N, P) playing an important role in the structure of the living organism, however, all of them are already included in one of the groups.

In the Goldschmidt system, the position and classification of many elements are uncertain, some elements are included in several groups, e.g., the elements of the iron group are found in the group of siderophile, chalcophile, and lithophile elements. Goldschmidt's geochemical classification of the elements is shown in the following periodic table (**Figure 17**).

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Figure 17 Goldschmidt's classification of the elements

1.6 Structure and composition of the Earth

The Earth has a radius of 6371 km and a mass of 5.972×10^{24} kg. Our planet shows a well-developed spherical internal structure which is represented in **Figure 18**. After the formation of the Earth, it underwent chemical differentiation, resulting in the formation of spheres with different compositions. The most important spheres are the crust, mantle, and core. The outermost part of a solid Earth is the **crust**. Its thickness varies greatly, ranging from 5 to 70 km, with an average thickness of about 40 km. Beneath the silicate crust, the **mantle** can be found, also of silicate composition, up to a depth of 2900 km. The crust and the mantle together are the so-called **Bulk Silicate Earth**. Beneath the mantle, an **iron-nickel metal core** (iron core for short) is situated, which can be divided into two parts. The outer core is in a liquid state and lasts to a depth of 5100 km, followed by the solid inner core. The core is ca. one-third, and the mantle is the two-thirds of the mass of the Earth, while the proportion of the crust is only 0.4%. The hydrosphere and the atmosphere are located above the solid crust. How did this spherical shell structure develop? As we wrote earlier, the planets, including the Earth, were formed during the collision, and clumping of planetesimals several 100 km in diameter. The larger ones gradually attracted smaller bodies as they circulated, until eventually planet-sized celestial bodies, including the proto-Earth, formed.

The material of the planetesimals may have had a chondritic composition. A significant part of the chondritic meteorites contains metallic iron-nickel phases (**Figure 18**), which were present in the planetesimals in addition to the silicates. As the planets accreted, the gravitational energy released and the heat from radioactive decay warmed the growing planets so much that perhaps the entire celestial body, but at least a significant portion of it, was completely melted, creating a magma ocean on their surface. This may have been the case for the nascent Earth.

In the molten material of the magma ocean, the molten metal and the partially or fully molten silicate phases were segregated, and due to gravity began the sinking and coalesce of metallic iron droplets. The separated iron has accumulated inside the Earth forming the metallic core.

There is debate as to whether the nascent Earth was completely or only partially molten. Even if the inside of the planet did not melt completely, the iron melt could still sink to the core. From the molten mantle, the iron droplets collected in Fe-rich ponds at the bottom of the magma ocean, from where they reached the proto-core either by percolation through an isotropic silicate matrix by the propagation of iron-filled fractures (diking), or by sinking as large iron blobs (diapirs) (**Figure 19**). Crust formation only began later and continues to this day.

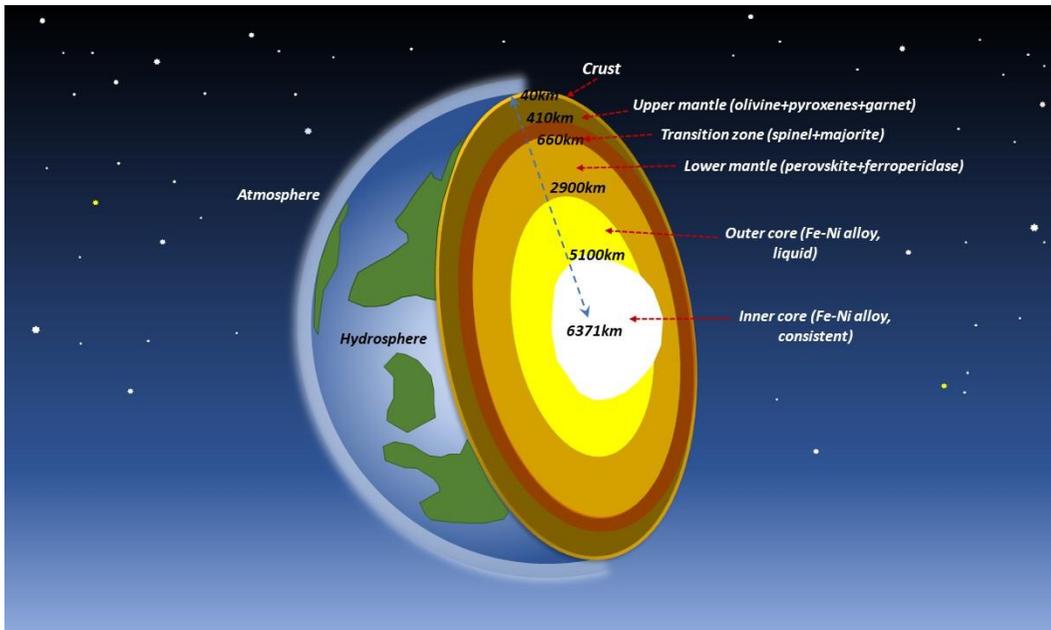


Figure 18 The internal structure of the Earth

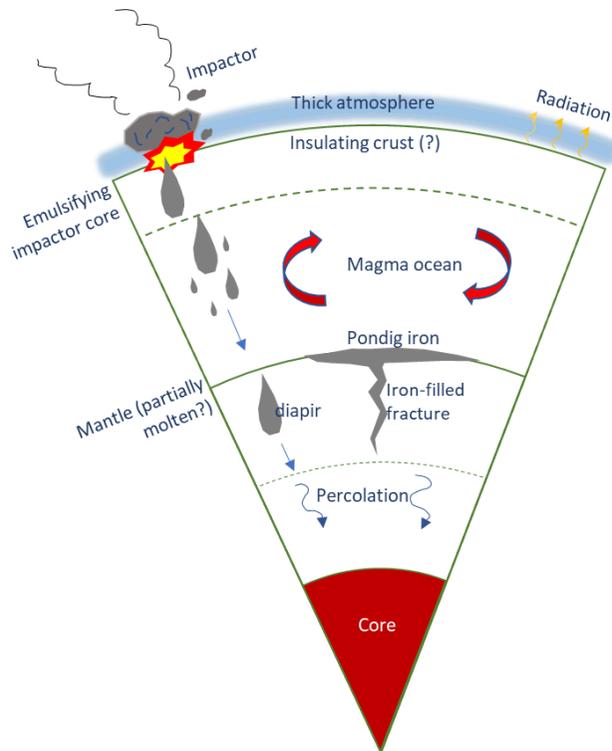


Figure 19 Possible processes of Earth accretion and core formation mechanisms

1.6.1 Composition of the Earth's core

In the molten material of the magma ocean, the molten metal and the partially or fully molten silicate phases were segregated, and due to gravity began the sinking and coalesce of metallic iron. The core constitutes 32.5% of the Earth's mass and consists essentially of Fe-Ni alloy. This composition is consistent with geophysical data, as well as the composition of Fe-Ni meteorites, as those meteorites are traditionally derived from the core of smaller celestial bodies. The Fe/Ni ratio is approximately 16. However, geophysical data and laboratory experiments suggest that the density of the outer core is lower than that of pure Fe-Ni alloys and therefore a component with about 5-15% less atomic weight should be present in the core, i.e., one or more elements with an atomic number less than iron. Such an element may be silicon, carbon, potassium, sulphur, hydrogen, or oxygen.

The core is the main storage site for siderophile elements on Earth, for example, platinum metals, gold, are concentrated here, but there are many other elements with siderophile properties, such as vanadium, chromium, tungsten, cobalt, phosphorus, molybdenum, gold, manganese, gallium, however, their abundance is small.

1.6.2 Age and composition of Bulk Silicate Earth

When the Earth is newly formed and partially melted, the iron-nickel core separated from the silicate mantle. The iron-nickel core constitutes one-third of the total mass. The remaining two-thirds are called Bulk Silicate Earth (BSE), also known as **Primitive Mantle** (PM). The terms Bulk Silicate Earth and the primitive mantle are thus synonymous. They refer to the composition and state of the mantle after the core separation, before the development of the crust. Of course, the elements that form the core were extracted from the Bulk Silicate Earth, so it contains less iron and nickel (and other elements enriched in the core) compared to chondrites.

In the early stages of the Earth's evolution, after a major collision leading to the formation of the Moon, a **magma ocean** formed. Its depth may have been at least 1,000 km, but some authors said the entire mantle may have melted. However, during the cooling and crystallisation of the magma ocean, there was no global magmatic differentiation due to strong convection, and the mantle remained relatively homogeneous. Convection may have been very strong in the low-viscosity ultrabasic magma ocean. The crystallising minerals (olivine and pyroxene at lower pressure, majorite (kind of garnet) deeper location, then Mg-perovskite in great depth) did not sink due to the flow but floated. As the solid/melt ratio increased, so did the viscosity of the magma. The predominantly

crystallised mantle consisted of a high-viscosity crystal suspension this is also called crystal mush in the literature.

Due to its lower specific gravity only, the basaltic melt was able to break through the crystal mesh in the upper 100 km of the mantle and reach the surface. The mantle thus remained relatively homogeneous even after complete solidification, and the first crust may have had a basaltic composition, similar to Mars and Venus. Overall, differentiation occurred only in the upper part of the mantle, from which the basaltic melt was extracted, and the remaining mantle was depleted from basaltic components. This depleted mantle is still present, while the rest of the mantle has remained relatively homogeneous.

1.6.3 The mantle

The largest part of the Earth, or, as it was called before Bulk Silicate Earth, or a primitive mantle, consists of silicate rocks.

There are several possibilities for examining the composition of the mantle. One possibility is the application of geophysical methods, including the study of the propagation velocity of seismic waves, which provides information on the structure and homogeneity of the mantle, and even to some extent on its chemical composition compared to experimental mineralogical data. Based on geophysical studies, we know that the mantle can be divided into two main parts, the **upper** and **lower mantle**, which are separated by a transition zone between 400 and 660 km.

The mantle, although based on the propagation of seismic waves can be regarded as a solid material, appears to be malleable and may perform slow movements, so convective currents may form in it. According to currently accepted models, convection flows in the lower and upper mantle are isolated from each other, suggesting some isolation of the two parts of the mantle. However, there are material movements such as the huge mantle uplift from the mantle-core boundary, the so-called plume, or the subsidence of the oceanic lithosphere penetrating the mantle, which crosses the boundary between the lower and upper mantle. However, geophysical surveys do not provide direct information on the chemical composition, especially the composition of trace elements. The chemical composition of the primitive mantle can be estimated by chemical analysis of the mantle samples and chondritic meteorites.

Direct samples from the mantle, on the one hand, were brought to the surface by tectonic movements (uplifting), often by peridotite massifs of several 10 km in size, or by smaller pieces of the mantle (up to a few 10 cm in diameter) called **xenoliths** (foreign rock) picked up by lava (basalt, kimberlite) derived from a great depth source. **Peridotite**, the characteristic rock of the mantle, has its name from the main constituent mineral olivine (peridot). Peridot is a mineral name that is primarily used by gemmologists and jewellers, while geologists and

mineralogists prefer the olivine name. Mantle-derived peridotite xenoliths can also be collected from some basalts.

Further information on the composition of the mantle can be obtained by examination of rocks crystallised from the high-temperature mantle-derived melt, such as basalts and komatiites, however, the composition of such rocks only indirectly reflects the composition of the mantle source. To estimate the composition of the primitive mantle, geochemists consider the composition of both peridotite xenoliths and basalts, since mantle-derived xenoliths are often post-melting residues that were depleted in basaltic components due to melting. Compositional study of the peridotite xenoliths and basalts, as well as the examination of peridotite-basalt melting system, were firstly performed by *Alfred Edward RINGWOOD* (born 1930, died 1993) an Australian geologist, geophysicist, and geochemist, who provided calculations about the mantle composition in the 1960s. His model is called the pyrolite model, based on the abbreviation of two main components pyroxene and olivine. Later, other authors performed similar calculations, which, although using different data and different calculation methods, gave results very similar to the original pyrolite model. The major element abundance in the primitive mantle is shown in **Table 1**.

Table 1 The major element composition of the mantle by various authors

	Pyrolite model (primitive upper mantle) by RINGWOOD (1989)	Pyrolite model (primitive mantle) by McDONOUGH & SUN (1995)
SiO ₂	44.76	45.00
TiO ₂	0.21	0.20
Al ₂ O ₃	4.46	4.45
FeO	8.43	8.05
MnO	0.14	0.14
MgO	37.23	37.80
CaO	3.60	3.55
Na ₂ O	0.61	0.36
K ₂ O	0.03	0.03
P ₂ O ₅	0.02	0.02

The upper mantle consists of olivine in 60 vol%, in addition, ortho- and clinopyroxene and garnet. Its significance will be in the processes of magma generation, so it will be presented in more detail there. With depth, the pyroxene gradually transforms to tetrahedral **Mg-garnet** (majorite). The **Mantle Transition Zone** between 400 and 660 km depth is built up by **olivines** (α -(Mg,Fe)₂SiO₄) polymorphs, such as **wadsleyite** (β -(Mg,Fe)₂SiO₄) and **ringwoodite** (γ -(Mg,Fe)₂SiO₄), and majoritic garnet in 40 vol%. The top of

the transition zone is marked with the transition of olivine to wadsleyite, while the bottom of this zone is represented by the ringwoodite transition to perovskite.

Below the Mantle Transition Zone with the increasing pressure (above 20 GPa), the ringwoodite and majorite become unstable, thus, the lower mantle is composed of magnesium and calcium perovskites, and oxide minerals, such as **periclase-magnesiowüstite** (Fe,Mg)O, or **stishovite** (SiO₂), respectively.

Although, as mentioned, the mantle does not show large-scale chemical stratification, certain inhomogeneities can still be detected in it. Some of the inhomogeneities detected in the mantle can be attributed to large-scale tectonic processes on Earth, such as the subduction of the oceanic lithosphere and the sediments of partly continental origin. However, for the upper part of the mantle, the depletion of incompatible elements (e.g., uranium, thorium, rubidium, barium, niobium, strontium, light lanthanides) is characteristic, which is associated with the formation of the continental crust.

2 PLATE TECTONICS

The distribution of continents and oceans has not always been the same in the history of the Earth as we can see it today. As we have discussed in the previous chapter the lithosphere, the solid outermost shell of the Earth overlies the asthenosphere, a zone in the upper mantle which, although being solid, is capable of ductile behaviour and thus is able to flow like jam (or jelly). Although jam is considered to be solid we can still pour it out of the jar and spread it over a slice of bread. Scientists believe that the lithosphere is broken into lithospheric plates (**Figure 20**) because of some kinds of motion either in the asthenosphere or in the upper mantle as a whole. As a result, lithospheric plates are in a constant motion relative to each other therefore plates are observed to be diverging, converging or passing each other along active plate margins. **Figure 20** shows the major (Eurasian, North American, South American, Pacific, Indo-Australian, African and Antarctic plates) and medium-sized (Arabian, Caribbean, Cocos, Nazca, Scotia, Fiji and Philippine plates) lithospheric plates. Lithospheric plate movement was proved not too long ago in the history of science. Before, continents were considered fix.

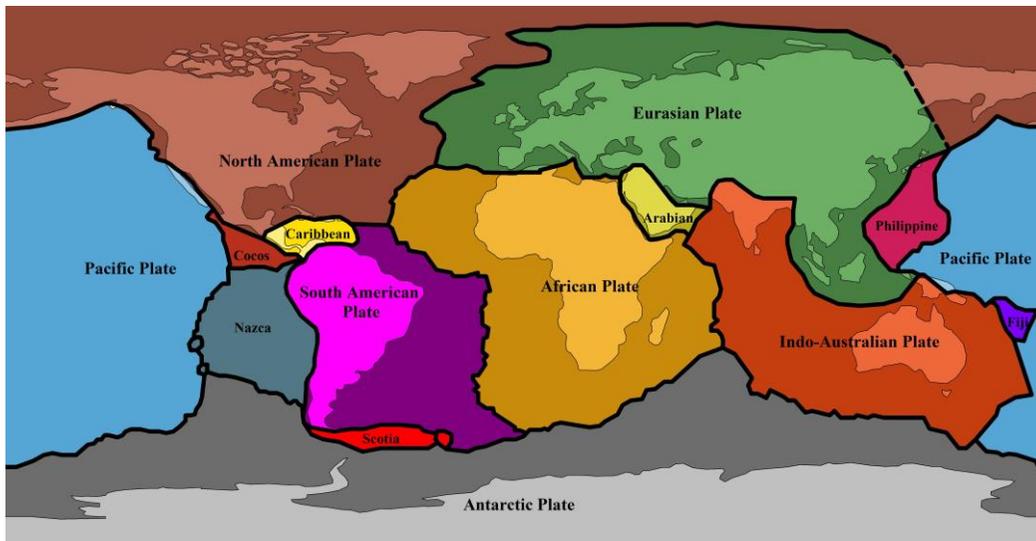
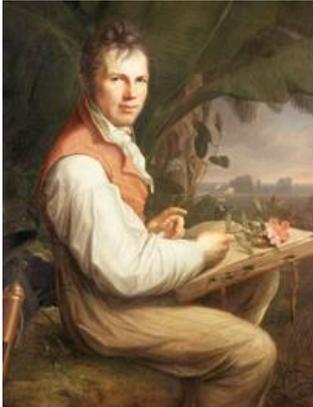


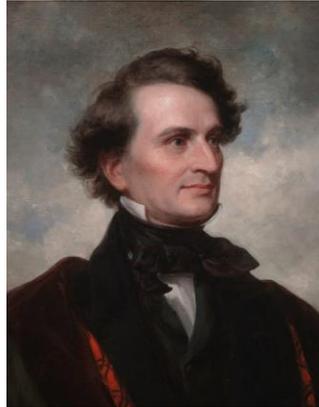
Figure 20 Major and medium-sized lithospheric plates

2.1 A bit of a history

Alexander von HUMBOLDT (Figure 21) realised that the similarities of Africa and South America went beyond that of an apparent ‘matching’ of their coastlines. Several mountain ranges that seemed to end on South America's eastern coast resumed on the western shores of Africa, and there were striking resemblances between the geological strata of the two continents; the mountains of Brazil, for example, were the same as those of the African Congo.



Alexander von Humboldt (1769 – 1859)



James Dwight Dana (1813 – 1895)



Alfred Lothar Wegener (1880 – 1930)



Arthur Holmes (1890 – 1965)



Harry Hammond Hess (1906 – 1969)



Frederick Wine (1939 –)
Drummond Matthews (1931 – 1997)

Figure 21 Some of the famous scientists forming ideas regarding the dynamic Earth and shaping the theory of plate tectonics

James Dwight DANA (**Figure 21**) declared that the Earth was cooling and contracting from a previous molten state. All geological features were caused by this contraction, and even the great mountain ranges were seen to be equivalent to the wrinkles in the skin of a drying apple. *Alfred WEGENER* (**Figure 21**) was basically a meteorologist, teaching undergraduates at the Universities of Marburg and Hamburg. Later he was appointed head of department at the University of Graz. He studied in detail the similarities of the coastline and then the shelf of South America and Africa, the structure of the mountain ranges on the two continents and the distribution of fossils. He wanted to understand the shift of climatic zones in the geological past, establishing the basics of palaeoclimatology. As he was not a geologist originally, his thoughts were not tied by fixist theories (continents are fixed and not moving). When WEGENER read critics on the idea that land bridges once connected the continents through which living creatures could migrate from continents to other continents being far away from each other today, and later these land bridges subsided into the oceans, he began to work out his theory of moving continents. According to the critics, land bridges were impossible to subside into the ocean because the rocks forming the ocean floor have higher density than the rocks forming the continents, therefore the continental crust rocks of land bridges could not subside into the oceanic crust. Opposed to land bridges WEGENER suggested the idea that continents were once joined together and drifted away from each other later. Since he could not give a convincing explanation for the driving force behind the movement of the continents, his theory was not accepted by the scientific public during his life even though he modified the theory several times and gathered a bunch of evidence. He considered forces from the rotation of the Earth, polar wandering, the processes along mid-oceanic ridges or the tidal effects as possible driving forces of continental movement. These were all denied and proved wrong by geologists; therefore, the views of WEGENER were not accepted. He organised three scientific expeditions to Greenland and died during the third one in 1930.

WEGENER's idea of continental drift was rejected by most scientists, but some research yielded results that could be explained only with the development of the plate tectonic concept. Such were the measurements of the Dutch geophysicist, *Felix Andries Vening MEINESZ* who invented a precise method for measuring gravity. The invention made it possible to measure gravity at sea enabling the discovery of gravity anomalies, which were attributed by MEINESZ to continental drift.

The idea of moving continents, as fresh and new, caught the imagination of scientists overseas. *Harry HESS*, a professor of geology at Princeton University in the US (**Figure 21**) had the opportunity to use sonar to map the ocean floor across the North Pacific Ocean while he was a US Navy officer in World War II. His views on the process of moving continents were outlined in the publication 'The History of Ocean Basins' in 1962. The theory is known as 'Sea

Floor Spreading'. HESS identified mid-oceanic ridges where mantle material ascends raising the ridges up to 1.5 km above the surrounding sea floor (abyssal plain). He also found that deep sea trenches are very close to the continental margin in the Pacific. HESS believed that oceans grew from the mid-ocean ridges as the new seafloor spread away from the ridge in both directions. He also considered that the ocean floor was destroyed and recycled along ocean trenches.

Arthur HOLMES (**Figure 21**) was one of the first enthusiastic supporters to WEGENER's concept of continental drift. Holmes presented the idea that convection currents in the mantle could be the driving force of continental drift in a lecture to the Geological Society of Glasgow on 12th January 1928. The Society's Transactions published the paper on mantle convection in 1931. In his 'Principles of Physical Geology' first published in 1944, HOLMES described the process of mantle convection driving seafloor spreading from mid-ocean ridges towards subduction zones.

In 1963, *Frederick VINE* (**Figure 21**) together with his PhD thesis supervisor, *Drummond MATTHEWS* (**Figure 21**) at Cambridge University revealed the magnetic anomaly belts of the oceanic crust and explained them with the change of the magnetic poles of the Earth. In their paper published in Nature they stated that the basalts of the sea floor were the same age at similar distances away from the ridge on each side.

A comprehensive work summarising the plate tectonic concept was written by *William Jason MORGAN* in 1968.

Some scientists considered that **convection currents** cycled throughout the entire mantle. Others believed that relatively "fast" movement is limited only to the upper mantle, however, the idea of simple convection currents is now largely out of favour as modern tomographic imaging methods are unable to identify mantle convection cells that are large enough to drive plate movement. As there is little or no evidence that mantle convection moves the lithospheric plates we are still where Alfred WEGENER was and cannot say undoubtedly what is the driving force of plate motion.

The strongest contender for controlling plate movement is the process known as '**slab pull**' as the coupled system of plates and the mantle is now widely accepted. Newly formed oceanic lithosphere is less dense at mid-ocean ridges than the asthenosphere, however, it becomes denser as it cools and thickens. As a result, it sinks into the mantle along subduction zones pulling the lithosphere apart along divergent plate margins.

Where slab pull is not working (or not alone) '**ridge push**' is another option. As the lithosphere at divergent plate boundaries is hot and less dense than its surroundings, it rises to form oceanic ridges. The plates formed here slide sideways pushing the plate material in front of them, resulting in the mechanism of 'ridge push'.

Nevertheless, plate movement is being revised all the time as new evidence is found by scientists but yet the details remain highly controversial and uncertain for the time being.

2.2 Divergence in plate tectonics

Ascending mantle material accumulates heat energy below the continuous continental lithosphere, which is thus thermally expanded and uplifted resulting in divergence and eventually the formation of **rift system** where the continuous continental lithosphere is torn. Divergence along the rift will eventually separate the two sides of the once continuous continental lithosphere and will result in the birth of typical oceanic crust and later a wide and deep oceanic basin (**Figure 22**). As a result, such boundaries are also known as **constructive plate boundaries**. This process has fundamentally four phases. In the initial phase the continental lithosphere is slightly uplifted and steep, parallel fractures are formed due to tension. These fractures become faults in the second phase along which the rocks subside and a central trench is formed as a result, which is also referred to as rift. In this phase active volcanism may occur on the surface but no typical oceanic crust is born yet. Typical oceanic crust is formed in the third phase when the once continuous continental lithosphere is completely torn. Finally, in phase four, divergence results in the development of a typical ocean basin in the centre of which a rise forms the mid-oceanic ridge where volcanic activity is continuous resulting in the accretion of the oceanic crust (**Figure 22**).

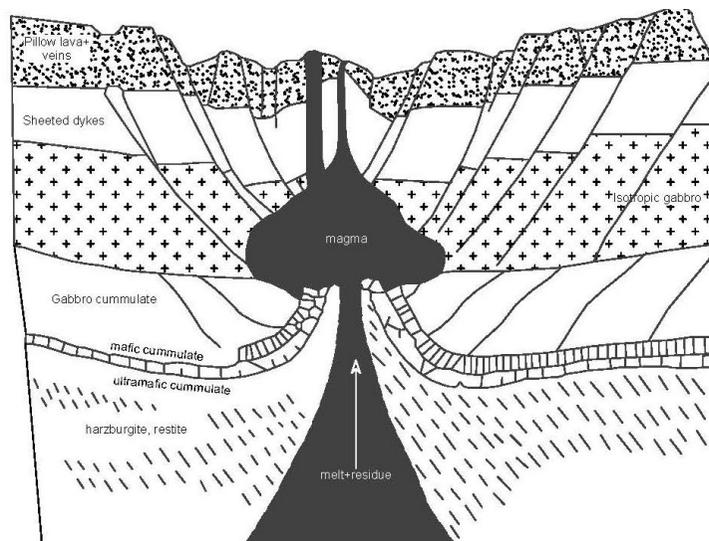


Figure 22 Structure and magmatism of a rift system

Mid-oceanic ridges rise above the surrounding abyssal plain of the ocean floor due to thermal expansion caused by the extreme heat flux supplied by the rising mantle material. This tholeiitic mantle material undergoes **fractional differentiation** while it ascends towards the surface. As a result, a series of ultramafic and mafic igneous rocks are produced, known as the **ophiolite series**. The bottom of the ophiolite series is composed of **ultramafic rocks** like harzburgite and peridotite. Above this zone, **gabbro** dominates. The gabbros are overlain by **dolerites**, which frequently occur in the form of **sheeted dykes**. Finally, basalts represent the top of the ophiolite series. In shallow magma chambers the magma undergoes low-pressure crystal fractionation, which results in the formation of olivine phyric rocks. This porphyritic texture is typical of **ridge basalt**.

Sometimes the basalt lava reaching the surface flows all over the ridge due to its low viscosity forming **flood basalt**. In other cases, the basalt is cooled rapidly when contacting with seawater and drop-like solidification results in pillow basalt structures. It is also possible that the basalt lava flows into the tension joints of the mid-oceanic ridge producing basalt dykes. Occasionally, the magma is trapped in small pockets along the tension joints and undergoes extreme fractional crystallisation and silica acid accumulation resulting in the formation of granite (A-type, anorogenic granite).

The speed of plate motion is rather small, around 6 cm/year on average. Some lithospheric plates move faster while others move slower. Fastest spreading, for example, is measured at Easter islands with 18 cm/year. In contrast Eurasia moves at around only 1 cm/year while rotating in a clockwise direction. The speed of diverging plates at mid-ocean ridges, however, is variable resulting in the breaking and strike-slip faulting of the ridge.

Both cold-water alterations and hot water hydrothermal activities occur along mid-ocean ridges. Cold-water metasomatism occurs in the upper part of the oceanic crust. Since both tension and cooling result in the formation of joints the oceanic crust is highly fragmented near mid-oceanic ridges. Seawater infiltrates along these joints and thus penetrates into the basalt 2-5 km below the ocean floor. Cold-water alteration of the basalt along mid-oceanic ridges include the substitution of olivine and volcanic glass with montmorillonite, zeolitisation and chloritisation. As a result of all the above metasomatic processes, a large amount of oceanic water is stored in the basalt of the oceanic crust. Regarding hot water hydrothermal activities, the most striking features are known as **smokers** (**Figure 23**), which are chimney like structures on mid-oceanic ridges where hot water is released constantly. In the case of black smokers, the water is painted dark by certain sulphides, mainly pyrite (FeS_2), galena (PbS) and sphalerite (ZnS). The temperature of the water flowing out of the 10 m high chimney-like features is around 350°C . However, this hot water does not boil due to extreme pressure. The somewhat cooler white smokers release relatively clean water with a temperature of around 200°C . The heat energy of the hot water released

by the smokers could be utilised for electricity production if the structures to be engineered for a power plant could bear the pressure and temperature conditions prevailing here. Unfortunately, some engineering innovations are still missing for the effective utilisation of certain geothermal potentials.

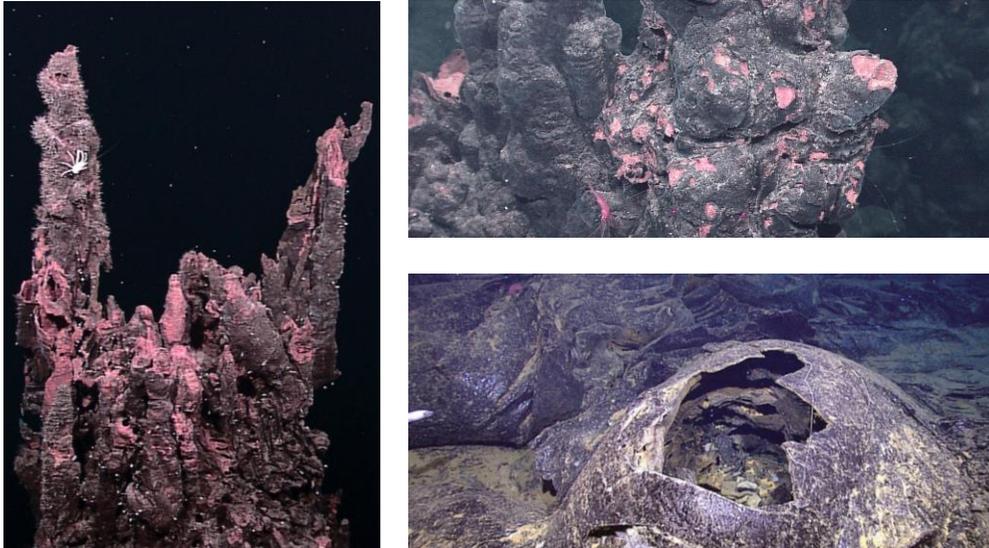


Figure 23 Head (left) and walls (top right) of a black smoker, the broken top of a pillow lava structure (bottom right)

As magma rises along mid-oceanic ridges the magnetite minerals in the magma are oriented into the magnetic directions, indicating magnetic north and south at the time of the formation of the rocks (as found by VINE and MATTHEWS). However, magnetic directions were not always the same in the geological past! Today's magnetic orientation is known as 'normal' but 'reverse' magnetic orientation has been also frequent in the past. As scientists found a period of normal polarity lasts for 480,000 years on average and reverse polarity lasts for 400,000 years on average. As a result, if we can measure the absolute age of oceanic crust rocks at a couple of places, the age distribution of the entire oceanic crust can be calculated (**Figure 24**). Interestingly, the age of the oldest oceanic crust is only around 180 million years while that of the oldest continental crust is around 4.4 billion years. The enormous age difference suggests that older oceanic crust was somehow consumed.

2.3 Seamounts and hot-spot volcanoes

Some active volcanoes are located not along plate boundaries but within lithospheric plates. Hawaii is a fine example of such within plate volcanoes, which are located above so-called **mantle plumes**. Mantle plumes drive heat

and material from very deep in the mantle of Earth. Where this heat flow reaches the crust a so-called **hot-spot** is formed. Hot-spot volcanism builds a composite volcano, which moves as the crust moves above the mantle plume. The location of mantle plumes remains the same for millions of years. However, the lithospheric plate moves above the mantle plume therefore the hot-spot always builds another volcano forming, in this way, an entire chain of volcanic islands. As the volcanoes move away from the active hot-spot erosion wears away the volcanic structures. In time, the height of the volcanic mountain will not reach the sea level and the former volcanoes will form so-called **seamounts**.

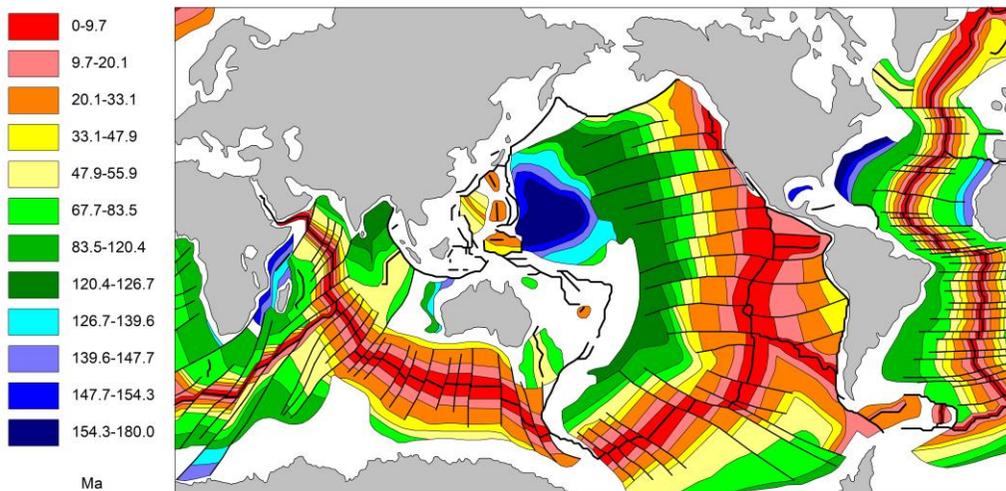


Figure 24 Age distribution of the oceanic crust

Hot-spot volcanism produces basalts that differ from basalts found along mid-ocean ridges in many respects. However, their small viscosity and fast flowing character is similar. Seamounts are built very rapidly resulting in huge composite volcanic structures. In the case of Hawaii, the volcanic structure reaches over 4 thousand metres above sea level and can be followed another 4 thousand metres below sea level. This indicates the rapid growth of the volcanic structure and a huge amount of volcanics produced by the mantle plume. Some scientists believe that the source of most mantle plumes can be traced back to the depth of the boundary of the mantle and the core at around 2900 km.

2.4 Transform plate movement – conservative plate margins

Where lithospheric plates slip past each other, lithosphere is neither created (constructed) nor destroyed (deconstructed), instead the boundary of the lithospheric plates remains mainly unchanged in a global perspective therefore such plate boundaries are known as **conservative plate boundaries**. Some of these plate margins are found between two sections of mid-oceanic ridges, like in the case of the San Andreas Fault where two distant parts of the Pacific Ridge are connected by a transform fault, which, at the same time, is also the boundary between the Pacific plate and the North American plate. Since largely different structures and landscapes are connected via transform faults, conservative plate boundaries are also known as **transform plate boundaries**. Along transform faults extended shear zones occur along which the rocks of the margins of the two lithospheric plates sliding horizontally past each other are strongly deformed. Such shear zones are characterised by a high number of shallow earthquakes, large lateral displacement of rocks without significant vertical displacement and a broad zone of crustal deformation with the occurrence of typical shear zone indicators.

Maybe nowhere else on Earth is such a landscape more spectacularly displayed than along the San Andreas Fault in western California, which is one of the finest textbook examples of the phenomenon. However, many transform plate boundaries are found on the sea floor where they connect segments of mid-ocean ridges like in the Atlantic Basin between South America and Africa or along the Antarctic Ridge between the plates of Indo-Australia and Antarctica and also between the Pacific and Antarctica plates.

2.5 Convergence in plate tectonics – destructive plate margins

As you can see in **Figure 24**, youngest oceanic crust material is found along mid-ocean ridges, as oceanic crust material, i.e., ophiolite series rocks are formed continuously there. Oldest oceanic crust is around 180 million years old and it is found closer to continents (e.g. around Japan, near the Caribbean). The age of the oldest oceanic crust is very young compared to the oldest continental rocks which surprised scientists who started to study these zones. As a result, the zones of subduction were found.

If the movement of the continent is blocked by something (probably another mid-ocean ridge) stress grows in the lithosphere. As a result, compressional stress field is established in which the thinner and more elastic oceanic crust can be deformed easier, bent and pushed underneath the rest of the oceanic lithosphere and the continental lithosphere as well in the process of subduction (**Figure 25**).

The oceanic lithosphere is broken and bent somewhere distant from the continent and the process of **subduction** starts with the descending oceanic slab subducting under another oceanic lithosphere (**Figure 25**). The greater is the continent the further away the oceanic crust will be broken, bent and subducted. It is also possible, however, that subduction starts almost at the edge of the continent. Due to subduction, the oceanic crust is gradually consumed and molten into the material of the asthenosphere and the mantle generating volcanic activity here as well. The plane along which the descending slab moves gradually into the asthenosphere was identified first by Victor Hugo Benioff, an American seismologist based on earthquake hypocentres. As a result, the zone of subduction is named after him and known as **Benioff zone** (the zone was also identified independently by Kiyoo Wadati, a Japanese seismologist at around the same time therefore the zone is also known as the Benioff – Wadati zone).

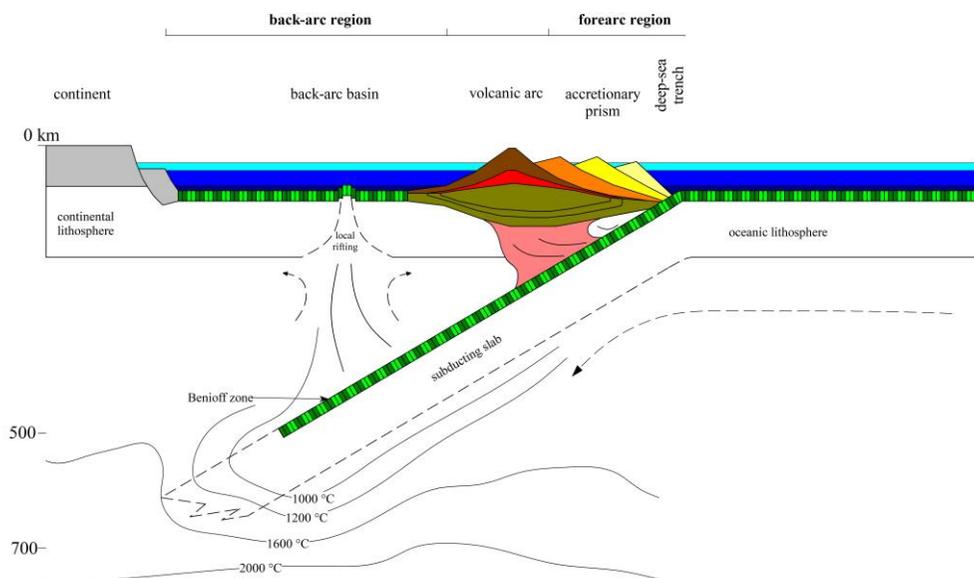


Figure 25 Typical structure of an island arc system

One process, the scientists intended to believe to be one of the probable driving force of plate tectonics is '**slab pull**'. Subducting oceanic slabs are so old and consolidated that their density is very similar to that of the upper mantle. In certain areas (where active subduction zones can be found) the density of the oceanic lithosphere is somewhat greater than that of the underlying mantle therefore the oceanic lithosphere subsides. This subsidence pulls the entire oceanic lithosphere (and even overlying continental lithospheres). Some scientists doubt that the very small density difference between the subducting slab and the mantle would be enough to drive continents and oceanic lithospheres; and the

details of the process are *not yet understood clearly* therefore finding the main driving force of plate tectonics is still one of the hottest research topics in structural geology and geodynamics.

Nevertheless, the subducting slab enters and the asthenosphere the temperature of which is much higher than that of the subducting slab (over 1000°C at the depth of 100 km and almost 2000°C at 600 km). However, this temperature difference is not enough to melt entire, several tens of kilometres thick oceanic lithospheres. For the complete melting of the subducting slab the reduction of the melting point of its rocks is also necessary. This decrease of the melting point is caused by water stored in the material of the subducting slab. The source of this excess water is double. On the one hand, deep sea sediments deposited on the ocean floor and travelling on the top of the oceanic lithosphere are mostly shale and clayey-silty sediments or maybe marls that have high water content. Some of these sediments are scratched off the ocean floor before subduction but the rest is squeezed into the subduction zone. On the other hand, as the oceanic lithosphere travels further away from the mid-ocean ridge it cools, which results in the formation of almost vertical cooling joints. Seawater enters into these cooling joints and travel into the subduction zone as well lowering the melting point of the rocks of the subducting slab. In this way, the material of the subducting slab melts generating magma in the mantle wedge between the subducting slab and the overlying lithosphere.

The igneous activity produces a volcanic arc, which follows more or less the line of the continent margin and is located in a specific distance from the continent depending on the size of the continent and the angle of the Benioff zone. Several structural elements occur in the emerging **island-arc system** (Figure 25); these are the following from the ocean towards the continent: **deep-sea trench, accretionary prism, fore-arc basin, volcanic arc, back-arc basin.**

Subduction, the formation of island-arc systems where the oceanic lithosphere and continental lithosphere converge are typical features along **destructive plate margins**. Here the volcanic-arc is formed by magmatism generated by the subducting oceanic crust but a wide range of other processes like magma mixing for example produces a wide variety of igneous rocks the details of which is discussed in chapter 3.

The deep-sea trench may reach down to 10,000 metres below sea level (the deepest is the Mariana Trench, which is also the deepest point on Earth). Mainly ocean floor sediments are accumulated at the bottom of such trenches. The deep-sea sediments ripped off the subducting slab are pushed and piled into a positive landform, the accretionary wedge or prism. Between the volcanic arc and the accretionary wedge, a negative landform, a basin is developed, known as the fore-arc basin. In this basin a mixture of deep-sea sediments eroded from the accretionary wedge and volcanic rocks eroded from the volcanic arc can be found. All these structural elements are found on oceanic basement rocks. And

even the volcanic arc is likely to be developed above oceanic basement rocks in the case of typical island-arc systems. Sometimes the accretionary wedge may reach the surface of the sea forming small islands like Barbados.

Between the volcanic arc and the continent, a generally wide basin can be found, the **back-arc basin**, which is also known as the marginal basin. Back-arc basins have either continental or oceanic basements, however, oceanic ones are more frequent. Apart from sediments accumulated from both the continent and the volcanic arc, back-arc basins are characterised by basic igneous rocks produced by volcanism related to **local rifting**. Since the descending slab disturbs the material in the asthenosphere, small-scale, local convection occurs resulting in an extensional stress field in the back-arc basin and spreading in the basement of the back-arc basin. This effect of the subducting slab is known as the '**roll-back**' effect. Roll-back is more general in the case of mature subduction, which has a steep Benioff zone and reaches down to around 650 km. Steep subduction, roll-back and spreading in the back-arc basin are typical features along the western margin of the Pacific plate thus subduction here is known as Mariana type subduction. Although the speed of the Pacific plate is relatively great (10 cm/yr) convergence along the western margin of the plate is relatively slow because the relative motion of the converging plates is more or less the same with the Pacific plate moving faster than the Eurasian or the Philippine plate.

In contrast, convergence of the Nazca plate and the South American plate is known as Chilean type subduction where the oceanic lithosphere subducts under a continental lithosphere along a flat dipping subduction zone. Whether subduction is steep or flat dipping is probably determined by the speed of convergence (fast convergence – flat subduction), the age of the subducting oceanic lithosphere (younger, less dense – flat convergence) and the buoyancy of the overriding plate and the thermal properties of the mantle wedge. The flat subduction of the Nazca plate under the South American plate means that active volcanism generated by the melting and consumption of the subducting slab occurs far from the trench. Therefore, the active volcanoes are also found far from the shoreline where active compression forms a high, strongly folded mountain range.

Subduction of the oceanic plate is not always a simple and smooth process. As we have seen before, scattered seamounts can be seen all over the oceanic plates. Such seamounts are several hundred or thousand metres high mountains that sometime occur in series. Such series of seamounts are passively transported by the Pacific plate that is subducting under Eurasia and forms the Japanese islands as an island arc system. Seamounts entering the subduction zone are high and rigid. They cannot be pushed into the subduction zone so easy. They block subduction for years resulting in the accumulation of stress. When the accumulated stress overcomes the resistance of seamounts, the volcanic cones will give way, collapse and will be squeezed into the subduction zone.

Consequently, blocked subduction will move forward several metres or hundreds of metres releasing enormous earthquakes and associated tsunamis. Such collapse of a seamount happened in the case of the Sendai (Tōhoku) earthquake in 2011 when a magnitude 9.0 earthquake shocked Japan and the triggered tsunami caused a catastrophe that will be remembered for long.

When subduction eats up the oceanic crust the island arc systems formed during the process (often several of them in a parallel line) are pressed onto the continent in the compressional stress field forming a very complex system (*melange*). Towards the end of subduction some small oceanic crust material (remnant blocks of the oceanic crust) are not pushed into the asthenosphere but pressed into the emerging high mountains. In this way these small oceanic crust blocks, remnants of the subducting slab are not involved in subduction, they will be involved in **obduction** instead. Eventually, as subduction is completed two continental lithospheres will converge towards each other. Due to the buoyancy of continental crusts, neither of them is capable of subduction only delamination may occur, therefore they will **collide** with each other forming an uplifted, strongly deformed structures in a compressional stress field. One of the most typical, textbook example of continent-continent collision is the Himalaya system where India, following a long journey from Gondwana, south of the Equator, towards north, north-east, collided with Eurasia. The former ocean between the two lithospheric plates (the eastern basin of Tethys) was consumed by subduction while deep sea sediments were thrust together with a part of the colliding and deforming continental lithospheres forming a high mountain range. Magmatism is also generated here as we will see in chapter 3.

The collision of continental lithospheres is known as **orogenesis** and areas experiencing the tectonic and magmatic processes related to orogenesis are known as **orogene zones** (**Figure 26**).

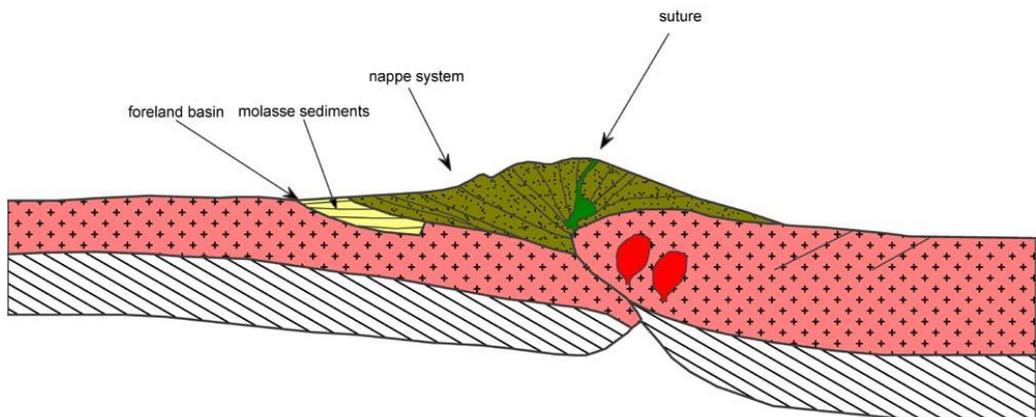


Figure 26 Typical structures in an orogene

As the two continental lithospheric plates compressed into each other increased pressure causes partial melting of the crust forming relatively shallow intrusions (red patches in **Figure 26**). Strongly deformed sediments are thrust into **nappe** systems the **vergence** (direction of movement of the nappes) of which points outward from a central zone with decreasing dip angle of the thrust planes. This central zone is indicated by the obducted remnants of the former oceanic lithosphere in the form of ophiolite series rocks (green zone in **Figure 26**). Generally, the remnants of ophiolite series rocks in orogenic zones are found in a line, more or less, known as the **suture** zone. The suture marks the place of the former ocean and the boundary of the formerly separate continental plates. With the emergence of the high mountain in orogene zones sediments accumulate in the foreland basins forming extended *molasse* sedimentary basins. Numerous orogene zones developed and evolved throughout the history of Earth in orogenies like the Caledonian orogeny in the early Palaeozoic, the Variscan orogeny in the late Palaeozoic or the most recent Alpine orogeny, which is still active in extended areas in the south of Europe and also in Asia.

As the forming mountains in orogene zones are high they will suffer from intense erosion. As a result, such mountains will be denuded within a couple of million years and in the last stage of the orogenic process a denuded, flattened surface will be formed that is more or less similar to the initial state as there will be a thick, continuous continental lithosphere where rifting and the opening of a new oceanic basin may start again, i.e., another orogenic cycle could start all over again.

Orogenic cycles have 5 stages: **early opening**, **late opening**, **early closing**, **late closing** and **cratonisation** (or **consolidation**). Early opening involves only extension in the continental lithosphere without the occurrence of oceanic lithosphere. When oceanic lithosphere occurs the process enters late opening, which will form wide ocean basins with mid-ocean ridges. Opening in the orogenic cycle is also known as rifting or **riftogenesis**. Early closing will see the start of subduction forming island arc systems in a process also known as **tectogenesis**. Late closing starts when the oceanic lithosphere is subducted and high, strongly deformed mountains are formed in the process of **orogenesis** increasing the mass of the continental lithosphere. Finally, cratonisation exposes the root of orogeny zones with igneous intrusions, strongly metamorphosed rocks and related ores. All in all, in the course of an orogenic cycle oceanic crust is formed and then consumed, and the volume of continental crust is increased; and also magmatism is generated in several different settings forming the igneous rocks that we are going to discuss in the next chapter.

3 PETROGENESIS

3.1 Genesis of primary magmas

Magmas that originate from the partial melting of the upper mantle and that have not been modified by fractional crystallization, assimilation, contamination, magma mixing or other additional processes are called as primary magmas.

Earth's 670 km thick outer zone can be divided into two major parts:

1. The crust is the 3-80 km thick part above the MOHO discontinuity;
2. The upper mantle is extending from the MOHO discontinuity to the depth of 670 km.

Prior to the genesis of the magma in the upper mantle, its normal, liquid i.e. partially molten state of the magma has to be discussed.

Studies of lherzolites suggest that the mantle – even though it contains a small amount of volatiles – is practically solid. As a consequence, partial melting must be a complex process which cannot be observed directly, however, surveying simple binary, ternary, and quaternary systems, and extrapolating the conclusions to the mantle, we can get closer to the solution of the problem.

These solutions will not be detailed here.

There are two ideal (i.e., end-member) models for partial melting.

1. According to the equilibrium (batch) melting the generated partial melt continuously reacts with the crystalline residue (and at the same time equilibrates with it) until segregation; in this model the bulk composition of the system remains constant.
2. According to the fractional (Rayleigh) melting after the genesis of the partial melt, it keeps differentiated from the system, so it will not react with the crystalline residue; in this model, the bulk composition varies all the time.

In natural systems the critical factor determining the partial melting process is whether the newly formed magma is able to segregate from the crystalline residual material. It depends on the permeability of the partially molten mantle. Criterion of formation of equilibrium melting is that a partial melting should reach the permeability threshold, and it is followed by segregation and accumulation of the magma. In the case of fractional melting, the magma is segregated after genesis, however, only if the mantle remains permeable even in the case of a very small partial melting (less than 1%). According to certain opinions, partial melting processes in the mantle represent the transition between the equilibrium and fractional models. According to the so-called critical melting model, the mantle becomes permeable after a certain equilibrium melting, and after that, the magma is segregated continuously from the residue and is accumulated.

As previously stated, the mantle is practically solid at its normal state, the question arises: how is it possible to have the amount of partial melting needed for a huge amount of magma genesis? For the spectrum of terrestrial basaltic magma 5-30% of molten material is required. This may occur in several ways:

1. Anomalous thermal perturbation.
2. Decrease of mantle solidus (due to volatiles added to the system, for example)
3. Adiabatic decompression, i.e., pressure-release melting.

It is generally accepted, that all three options are possible, but the genesis of a great volume of magma requires most probably the third mechanism, i.e., adiabatic decompression. It is confirmed by the fact, that a huge amount of hot mantle material originating from the deep occurs at mid-ocean ridges, thus these areas belong to the most significant magma generating areas of Earth.

Figure 27 helps to understand the process. The figure demonstrates the interval of a typical lherzolite melting. The interval between the solidus and liquidus is the field of partial melting. The temperature of the mantle lherzolite at point 'A' is determined by the geothermal gradient, and it is close to the solidus. Let us suppose that this part begins to rise adiabatically (due to a convective cell, for example) towards the surface. The pressure affecting the lherzolite material decreases dramatically, while its temperature practically remains constant. As can be seen in the figure, the rising lherzolite is heading closer to the liquidus field and will partially melt increasingly. At point 'A' the mantle diapir will contain approximately 20% partial melt and 80% residual crystals. Such an amount (20%) of partial melt is already able to be physically separated from the source, therefore at such a depth the magma can be segregated from the residual crystals and may rise rapidly. If segregation does not happen and the diapir continues to rise, the degree of partial melting will be even greater. Presumable pressure-release melting is the main magma producing process in the case of back-arc spreading centres and in the case of many intra-plate volcanic centres as well.

At the initial phase of the partial melting a coherent network of the melt may be formed in the crystalline matrix, that may flow due to its buoyancy. This can be the base of magma segregation. Magma genesis is affected by the density contrast of the silicate melt and the residual crystals, which increases in parallel with the decrease of pressure. This suggests that there must be a natural depth limit, where the basaltic melt comes from. The critical depth, where the difference of density between the melt and residual crystals is close to zero, may be at a depth of 100-200 kilometres (from 30 to 70 kbar) depending on the bulk composition of the partial melt. At greater depths, in the case of a very small difference in density or even in the case of density contrast, the generated melt may remain at the source area without segregating from the residual crystals. This may be the explanation why certain the mantle regions remain undepleted for a long time even in geological terms.

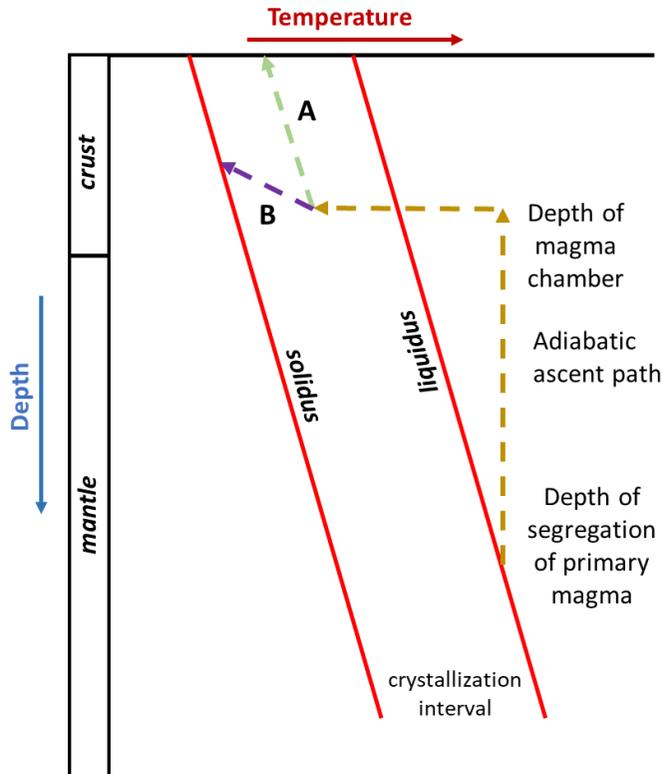


Figure 27 Path of a basaltic magma to the surface and its crystallization interval

Melt extraction is affected by the deformation and permeability of the matrix and the viscosity of the melt: separation of high-viscosity magmas is limited, while low-viscosity magmas are segregated rapidly.

Obviously, equilibrium melting can be the base part of the partial melting process only if the magma is able to stay at its place of origin for a longer time period. This may occur, for example, when the residue is so rigid, that deformation does not take place in it. However, the creep rate of olivine-rich materials is extremely high, therefore the residue will go through compaction, and the liquid phase will be squeezed out. As a result, based on the rheological features of the partial molten mantle, it is more possible, that primary magmas are generated by critical melting and not by equilibrium melting.

Permeability is required for melt extraction. Fluid-assisted shear fracturing may open the interconnected melt network veins. The pressure of the melt may cause fractures in the peridotite above, and finally may create dyke-like conduits reaching the surface. Since the melt can chemically react with the matrix through which it moved, the porous-flow stage is extremely important for the chemical features of the magma.

3.2 Processes modifying the composition of primary magmas

Segregated magma may go through several modifications during its rise, or when it stays in magma chambers in smaller depths, and due to these alterations various rocks of different compositions are finally created. During the rise the magma starts to cool down, and finally solidifies. This solidification occurs in a temperature range rather than at a given temperature. The initial temperature of crystallisation is called liquidus, the temperature of the complete crystallisation is the solidus. Both are pressure dependent. The rising magma – as mentioned already – is almost adiabatic, thus the basaltic magma is barely crystallised before reaching the crust. (It should be noted that significant amounts of olivine may be crystallised in picritic primary magmas during the transport towards the surface.)

When entering the crust of lower density, the magma may get to a state of zero or even negative buoyancy and may be stuck as a result, creating a magma chamber. It starts to cool down in the magma chamber and when the temperature falls below the liquidus, it begins to crystallise. At a temperature between the liquidus and solidus the magma is basically a mixture of crystals and liquid. When the chamber is replenished from below, eruption may occur, or the magma may be completely crystallised resulting in the formation of plutonic or hypabyssal igneous rocks.

Different minerals are not crystallised at the same time. For example, in the case of basalt, at atmospheric pressure (i.e., at 0 kbar) with the decrease of temperature, the crystallisation sequence is the following: olivine – plagioclase – clinopyroxene – ilmenite.

Since composition of the melt and the crystals are not the same, the composition of the residual liquid varies continuously during the crystallisation process. Crystallisation can be either equilibrium or fractional. In the case of equilibrium crystallisation, the crystal continuously re-equilibrates with the melt, thus the system remains constant regarding the bulk composition. In this way, for example, basalt or (crystallised in the deep) gabbro is formed from the basaltic magma. If the crystal and liquid fractions separate from each other within the magma chamber, no equilibrium, but magmatic differentiation will occur. The process is called fractional crystallisation.

It is a general opinion, that the wide variety of igneous rocks that may form from primary magmas is mainly the result of fractional crystallisation. The theory that crystal segregation is mainly gravitational prevailed for a long time. Recently, however, it was revealed, that convective motions in the magma chamber may be enough to keep the crystals in suspension, so in fact gravitational segregation may be secondary. The form of the convection is determined by the size and shape of the chamber, the processes taking place on the wall of the chamber, and the density and viscosity of the magma. During

crystallisation, the heavy element content of the liquid right next to the crystals may increase or decrease, thus its density will be different from the magma, that gives the main mass of the magma chamber, and therefore it convects away from the crystals. As a result, the crystals and the liquid will be separated from each other. The process is called convective fractionation.

However, despite the presence of convections vertical stratification regarding density, composition and temperature may be formed in the magma chambers close to the surface. Gradients may arise directly during fill up or replenishment. Such stratified magmas may fall apart into horizontal layers which are well-mixed and separated from each other by a thin diffusive interface. It is quite unlikely that magma chambers below huge volcanic structures could be presumed closed systems. On the contrary, new, relatively primitive magma is injected from the mantle into the lower part of the magma chamber time after time. The mixing rate depends mainly on the speed and quantity of the „input”, and the difference in density and viscosity between the two magmas. Magma chambers stratified in composition may be changed by crystallisation regarding the density and composition of the layers, that may lead to the mixing of the two layers. When hot, primitive magma is injected into the bottom of the chamber, thermal disturbance may be balanced without mixing, but if the density of the lower layer decreases due to crystallisation, the two layers can mix. When the volatile rich mafic layer gets into the colder silicate rich layer, it can cause a sudden quenching with gas phase exsolution, which may lead to explosion. Ultimately, magma chambers may be considered an open system in general, which can be tapped, replenished time after time and continuously fractionate.

As rising towards the surface, the magma can be contaminated by the assimilation of wall rocks to various grades. Assimilation in the mantle can be hardly proved based on major element data, however it can be indicated by the extreme accumulation of incompatible elements in the primitive basaltic magma. Assimilation in the oceanic crust is also difficult to prove; in contrast, in the continental crust, wall rocks have quite different major and trace element contents (the same applies to Sr-Nd-Pb isotopes), thus contamination here can be easily traced. At the same time, the tholeiite basalt of mid-ocean ridges (MORB) may also assimilate basalts originating from former, already slightly modified effusions.

Crust contamination may play a role in the differences of oceanic and continental tholeiites. The mantle can assimilate crust material in several ways:

1. Mixture of the partial melt material of the mantle and the crust.
2. Melting of the material composing the wall of the magma chamber or the conduit and mixing with the mantle material.
3. Blocks originating from the wall rocks enter the magma and assimilate.

Due to assimilation, the magma cools down and starts crystallisation. Assimilation is restrained by the thermal energy of the magma; therefore, in general, it is not very extensive, rarely exceeding 10-20%. Contamination is a

selective process, apparently rocks with smaller fusion temperature are included, i.e., rocks with a higher alkali feldspar content assimilate easier.

At the same time when the crust **assimilates fractional crystallisation** may also take place (AFC: Assimilation and Fractional Crystallisation). In that case, positive correlation can be found between the indices of differentiation (e.g., silica content and Fe/(Fe+Mg) ratio) and the incompatible trace elements or Sr-Nd-Pb ratios. If the magma chamber is formed in the lower crust or upper mantle, fractional crystallisation may occur without substantial crustal contamination. But if the magma gets into a higher position, it can be contaminated by more fusible rocks present in that environment. However, in both cases, only hot and primitive magmas are able to assimilate the surrounding rocks.

A less significant process still needs to be discussed. Liquid immiscibility may occur in alkali-rich melts, or in basaltic melts, where one of them is iron- and phosphorous-rich and poor in silica, while the other one is vice versa. The immiscibility of sulphide and silicate melts is known, which could be the reason of the formation of some especially important ore deposits. Liquid immiscibility could have also contributed to the formation of certain carbonatite magmas. The aqueous fluid of high temperature can dissolve significant amount of alkali and silica.

3.3 Magmatism at mid-ocean ridges

Magmatism at mid-ocean ridges (MOR) can be regarded as magma generation which occur at constructive plate margins. Deep mantle material reaches the surface here and produces basaltic magma, called as **mid-oceanic ridges basalt (MORB)**. The lithosphere forming along mid-oceanic ridges is symmetrical, i.e., growing on both sides, and its thickness is increasing further away from the ridge, reaching its maximum at a distance of around 60-80 kilometres. Chemical composition of MORB is influenced by several factors, such as the composition of the original mantle material; the degree of partial melting; the depth of magma segregation; the degree of fractional crystallisation and mixing processes in the magma chambers under the mid-oceanic ridges.

Petrography of mid-oceanic ridge basalts (MORB) is depending on the chemical composition and the cooling history. The crystal composition and the texture reflect if the melt, with a temperature close to liquidus, was crystallised in a cold submarine environment. Thus, the grain size distribution is variable (from vitreous to a 20-30% phenocryst content). Regarding mineral content, olivine (with or without Mg-Cr spinel), plagioclase + olivine (with or without Mg-Cr spinel), and plagioclase + olivine + augite MORB types are the most common. Most of the time, phenocrysts of MORB are obviously not in equilibrium with the host magma, suggesting the important role of magma mixing. Normal and

hot spot (elevated) ridge basalts usually show different petrographic features. In the case of a normal MORB, plagioclase and olivine are the dominant phenocryst phases, while pyroxenes may be missing. Both olivine and pyroxenes are present in MORB from elevated ridges.

The composition of olivine corresponds to the composition of the host magma: it may be Fo₇₃ in *ferro basalts*, Fo₉₁ in *picrites*. Spinel phase is common in picrites and in olivine-rich basalts, rare in plagioclase-rich basalts, its composition varies (Al₂O₃: 12-30 wt%; Cr₂O₃: 25-40 wt%). The composition of plagioclase is ranging from An₄₀ to An₈₈.; fast-spreading ridges (e.g., East Pacific Rise) are intermediary, while the slow-spreading ridges (e.g., Mid-Atlantic Ridge) have more calcic plagioclase phenocryst features. It refers to important role of magma mixing, that – similar to olivines – those are out of balance with the composition of bulk rocks.

Considering MORB major elements geochemistry, it is important, that SiO₂ (since it is varying in a very small range) is not a suitable index of differentiation, therefore, instead of silica content, MgO content or the so-called M value ($100\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$) is used. Maximum of M value of MORB glasses is between 55 and 65. Since the magma would be in balance with the mantle olivine at M=70, and the maximums are at M=55-65, consequently most of the MORBs do not represent the composition of primitive glasses, which refers to the importance of fractional crystallisation following segregation in the upper regions. Though the major element spectra are quite uniform, trace element concentrations of MORB coming from sites of topographic and structural features show significant variations. According to this, MORB is divided into three groups, such as **N-type** (normal, depleted), **P-type** (plume, enriched) and **T-type** (transitional) groups. N-type basalts are typical mainly in the Pacific and Atlantic regions south of 30°N, while P-type is typical for the Atlantic north of 30°N and the Galapagos. MORB is generally depleted in large low-valency cations (Cs, Rb, K, Ba, Pb, Sr), whereas K/Rb, K/Ba and Sr/Rb ratios differ in the composition of N-type and P-type MORB (**Table 2**). Concerning rare-earth element (REE) patterns, there are significant differences between N-type and P-type MORB compositions (**Figure 28**). N-type MORB is strongly depleted in light REE, whereas P-type MORB may show light-REE enrichment also. Accordingly, La/Sm_N is smaller than 1 for N-type and higher than 1 for P-type MORB, respectively. Concerning Sr, Nd, and Pb isotopic compositions, it can be stated that these component values range in a much smaller interval for Pacific MORB than for Atlantic MORB, which can possibly be explained with the greater isotopic heterogeneity of its source areas.

Table 2 Major and trace element geochemistry of normal and plume MORB

	Normal MORB (28-34°N)	Plume MORB (Azores)
(%)		
SiO ₂	48.77	49.72
Al ₂ O ₃	15.90	15.81
Fe ₂ O ₃	1.33	1.66
FeO	8.62	7.62
MgO	9.67	7.90
CaO	11.16	11.84
Na ₂ O	2.43	2.35
K ₂ O	0.08	0.50
TiO ₂	1.15	1.46
P ₂ O ₅	0.09	0.22
MnO	0.17	0.16
H ₂ O	0.30	0.42
M value	66.5	64.9
(ppm)		
La	2.10	13.39
Sm	2.74	3.93
Eu	1.06	1.30
Yb	3.20	2.37
K	691	4443
Rb	0.56	9.57
Cs	0.007	0.123
Sr	88.7	243.6
Ba	4.2	149.6
Sc	40.02	36.15
V	262	250
Cr	528	318
Co	49.78	44.78
Ni	214	104
K/Rb	1547	475

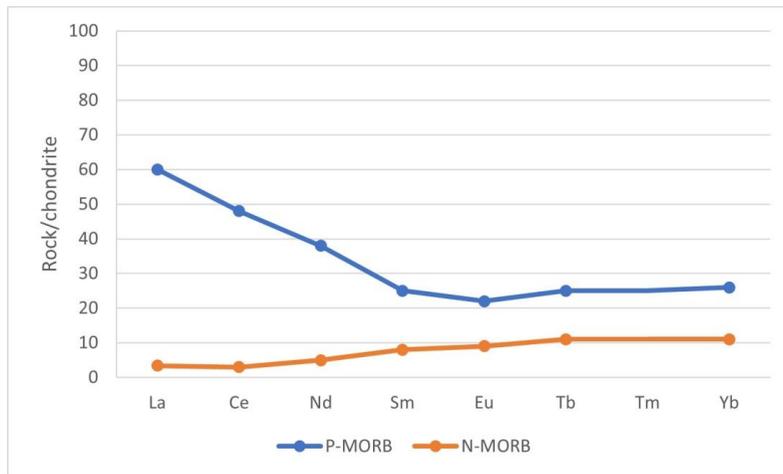


Figure 28 Chondrite-normalized REE patterns of normal and plume ridge basalts from Mid-Atlantic Ridge

Source heterogeneity, fractional crystallisation, and magma mixing played a role in the development and genesis of MORB in the same way. According to our recent knowledge, it is supposed that N-type MORB derives from depleted asthenospheric upper-mantle source, while P-type MORB is from enriched upper-mantle source (see **Figure 29**).

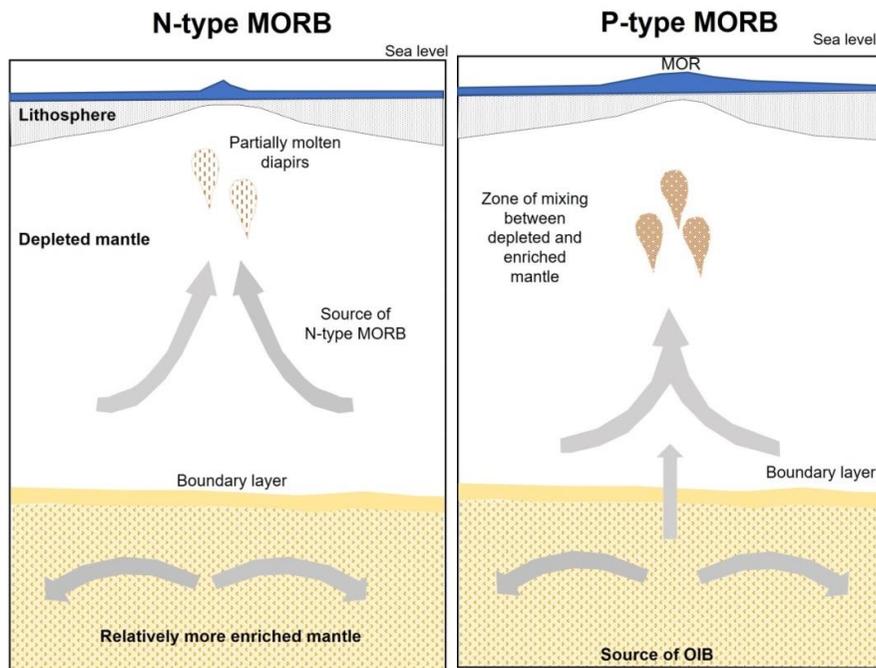


Figure 29 Models of origin of N-type and P-type mid-ocean ridge basalt

Magma may accumulate, fractionate, mix or even ultramafic cumulates may form under the mid-ocean ridges. In the beginning magma has a high density, therefore it remains at the bottom of the magma chamber and layers may be formed until fractionation reaches the point when rollover happens. Due to the position of the block, that composes the top of the chamber, more primitive magma material will emerge at the rift axis, and more fractionated magma material at the edges (**Figure 30**).

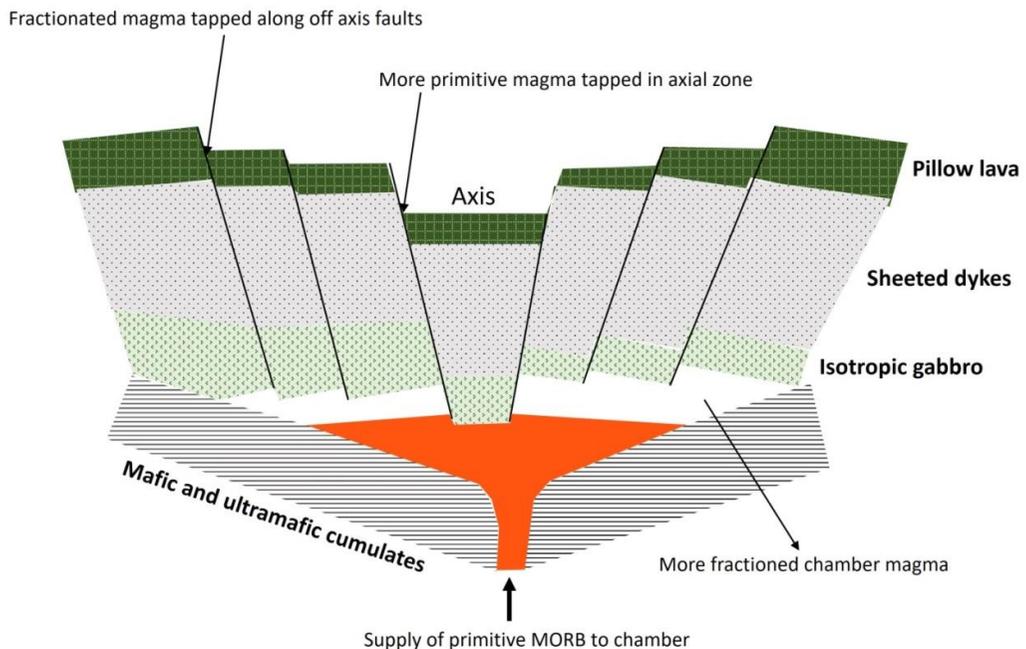


Figure 30 Scheme of hypabyssal environment of a rapid spreading mid-ocean ridge maintaining a dynamic magma chamber below the axis

3.4 Magmatism at island arcs

The area of subduction is one of the most complex tectonic regions, however, within this, ocean island arcs are considered to have the simplest magmatism, since the contamination effect of the continental crust can be ignored here. The subducting slab consists of the following materials: (1) **mantle derived lherzolite of the oceanic lithosphere**; (2) **oceanic crust composed of basalt and gabbro**; (3) **serpentinite**; and (4) **sediments of the oceanic basins**. With the increase of temperature and pressure, prograde metamorphism takes place during subduction, resulting in the metamorphism of the basaltic material into eclogite through greenschist and amphibolite facies, while hydrous minerals dehydrate; the realised H_2O will separate as a fluid

phase. Island-arc magmas may derive from several potential sources. These are the following:

1. The mantle wedge composed by oceanic lithosphere of 40-70 km thickness and varying thick asthenospheric upper mantle.
2. The oceanic crust composed of oceanic sediments and variably metamorphosed basalt, gabbro, and dolerite.
3. Sea water.

Island-arc magmas could be generated from metamorphosed ocean floor (amphibolite, eclogite, etc), lherzolite, and modified lherzolite coming from the reaction of lherzolite and slab-derived hydrous siliceous magma. The slab-derived fluids play particularly fundamental role in magma generation by lowering the mantle solidus, i.e. promote partial melting. Magma generation occur where the solidus of the source material is exceeded by temperature.

It seems, that beneath an island arc magma reservoir system occurs where primary magmas may fractionate. These magma chambers can be found usually up to a depth of 20-30 km below the surface, moreover the shallowest ones are no more than at a depth of a few hundred metres. However, it is possible, that this magma chamber system is refilled from the zone of magma generation (Figure 31).

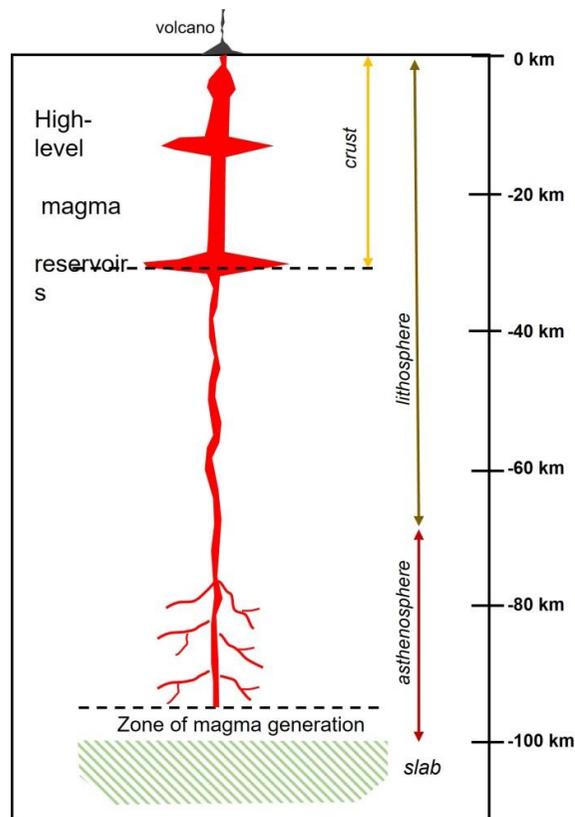


Figure 31 System of magma reservoirs beneath an island arc

Early concepts, mainly based on the Japan arc, suggest that increasing alkalinity is typical of island-arc volcanism when getting further from the trench, however, progressive studies clearly revealed that, concerning many island arcs, the situation is more complicated, since the composition of the magma is significantly influenced by source heterogeneity, degree and depth of partial melting, and variety of crystal fractionation. Nevertheless, tectonic setting and magmatism are undoubtedly in connection. In the early stage of island-arc evolution, basaltic volcanism dominates. With the development of the arc, the thickness of the crust may reach up to 20-25 km and then the crust may act as “density filter” for primary magma; then it will pond in a magma chamber of high-level, and, due to the subsequent crystal fractionation, andesitic magma of low density will be generated. This model is in accordance with the fact that intermediate (andesitic) volcanism is dominant in matured island arcs.

Although, according to the various petrographic spectrum, features of island-arc volcanism are also quite diverse, volcanoes can be divided into two major groups: (1) **basalt and basaltic andesites**, and (2) **andesite and dacite**. Volcanoes producing different materials are different in morphology as well: volcanoes producing more basic lavas are generally flatter, having shield-like shapes, while andesitic dacitic ones are conic composite volcanoes, and the different types often appear near each other. High volatile content is quite typical for island-arc magmatism, therefore explosive eruptions are also characteristic for these volcanic areas. Consequently, calderas formed by the collapse of high-level magma chambers are common volcanic forms in island-arc volcanic fields.

From petrographic point of view porphyritic nature is quite typical of the volcanic rocks of island-arcs. In general, ferromagnesian minerals (olivine, pyroxene, amphibole, biotite) are relatively magnesium-rich. The most common mafic mineral constituent is augite; plagioclase is the most frequent component of phenocrysts, usually Ca-rich. Quartz is also significant in acidic rocks, while alkali feldspars and feldspathoids only appear in high-K series rocks. The most frequent iron-titanium oxide is titanomagnetite that occurs in both basic and acidic volcanic rocks.

Based on their geochemical character, island-arc volcanic rocks belong to three major series: (1) **tholeiitic, calc-alkaline**, and **alkaline**. Since K_2O shows proportionally greater increase, it is used for further classification of the volcanic series in island-arcs. In this way four series can be formed, namely: (1) **low-K**, (2) **calc-alkaline**, (3) **high-K calc-alkaline**, and (4) **shoshonitic series**. Low-K series can be regarded to be corresponding to the island-arc tholeiitic series, while shoshonitic series can be synonymous with the alkaline series. Moreover, shoshonitic series is dominated by basaltic rocks. The basalt is typically enriched in incompatible elements of low ionic potential (such as Sr, K, Rb, Ba), while poor in elements of high ionic potential (such as Ta, Nb, Ce, P, Zr, Hf, Sm, Ti, Y, Cr, etc.). Nickel content of island-arc basalts is

generally low, which may indicate olivine fraction suggesting that they are not derived directly from primary magma. Rare earth element (REE) content of island-arc basalts seems to correspond to K_2O content; tholeiitic island-arc basalts are depleted in light-REE, while calc-alkaline basalts are enriched in light-REE.

Recently, it is widely accepted that basaltic magmas of the island arcs are generated by the partial melting of the mantle wedge located above the subducting slab and beneath the crust. It seems that island-arc magmas are in connection with the partial melt of mantle lherzolite metasomatised by slab-derived fluids. Although evidence shows that sediments are limitedly involved, some cases (Sunda-Banda arc, Lesser Antilles) indicate that terrigenous sediments are more considerably involved. Moreover, sediment may contaminate magmas in the crustal part of island-arcs. The variety in composition of island-arc magmas suggests that, beside crystal fractionation as a major process, magma generation beneath island arcs is quite complex (**Figure 32**).

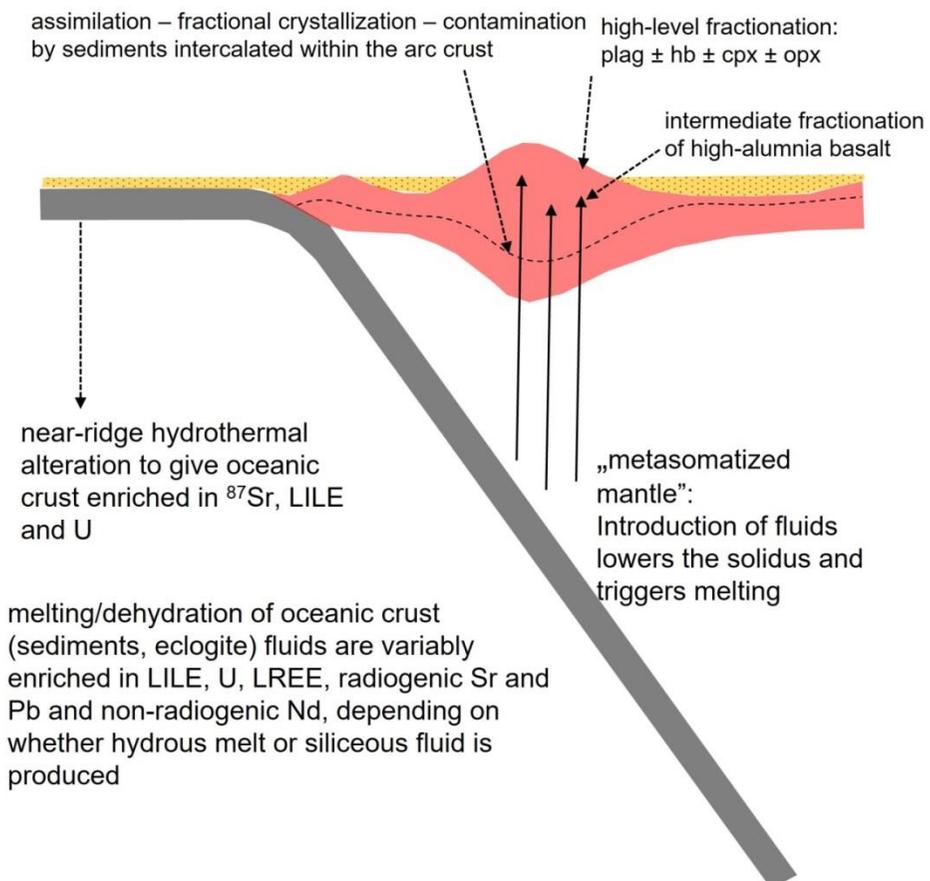


Figure 32 Magma generation processes in ocean-ocean collision zone

3.5 Magmatism at active continental margins

Magmatism at active continental margins represents a more complex pattern than island-arc magmatism, the other subduction-related one. Andesite associations characterising island-arcs are also characteristic igneous rocks for active continental margins, and the four major types of volcanic rocks (low-K, calc-alkaline, high-K calc-alkaline, shoshonitic) are also present here, the thick continental crust causes complexities in magma generation. Maybe, the most important difference between the calc-alkaline series of island-arcs and that of continental margins is the greater abundance of dacitic and rhyolitic and low representation of low-K magmas in the continental margins (**Figure 33**). Consequently, it is quite possible that partial melting of the continental crust plays a significant role in the generation of acidic magmas. The major rock-forming mineral in island-arc plutonic and volcanic rocks is plagioclase feldspar with a composition ranging from sodic to calcic, alkali feldspar (dominantly orthoclase), quartz as felsic, and pyroxene (mainly augite), amphibole (hornblende), biotite, and magnetite as mafic constituents.

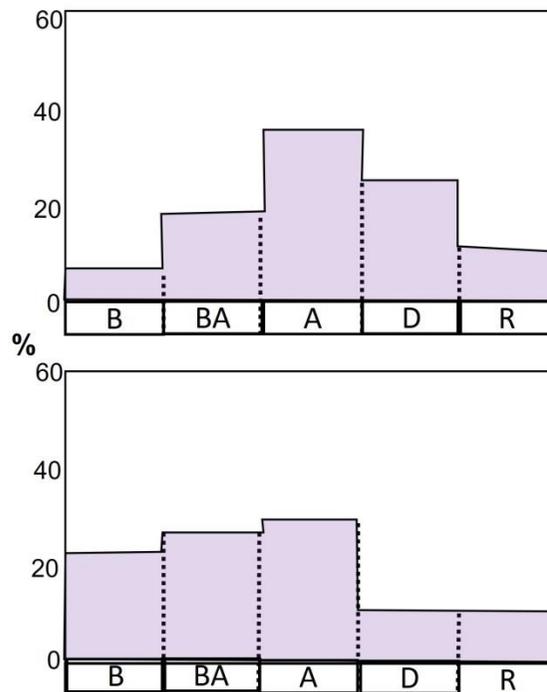


Figure 33 Frequency of basalt (B), basaltic andesite (BA), andesite (A), dacite (D) and rhyolite (the upper diagram refers to the Andes, the lower one refers to the island arcs of the SW Pacific)

Magmatism at active continental margins is a very important mechanism for the vertical growth of continental crust by adding extrusive (mainly andesitic lavas) and intrusive (dominantly granitic batholits) rocks to the crust. Processes are the same as those in the case of island-arc magmatism: The subducting lithosphere is heated going under metamorphic transformation involving dehydration; the released fluid will promote partial melting by lowering the solidus, and partial melting may happen in the so-called subcontinental lithosphere. Moreover, the magma has to pass through the continental crust with a thickness of 50 km. In the magma chamber located in the continental crust the magma is fractionated and is contaminated by assimilating some crustal material (**Figure 34**). Comparing to oceanic island-arc basalts enrichment in incompatible trace elements is greater which, beside an enriched mantle origin, also suggests the crucial importance of crustal contamination.

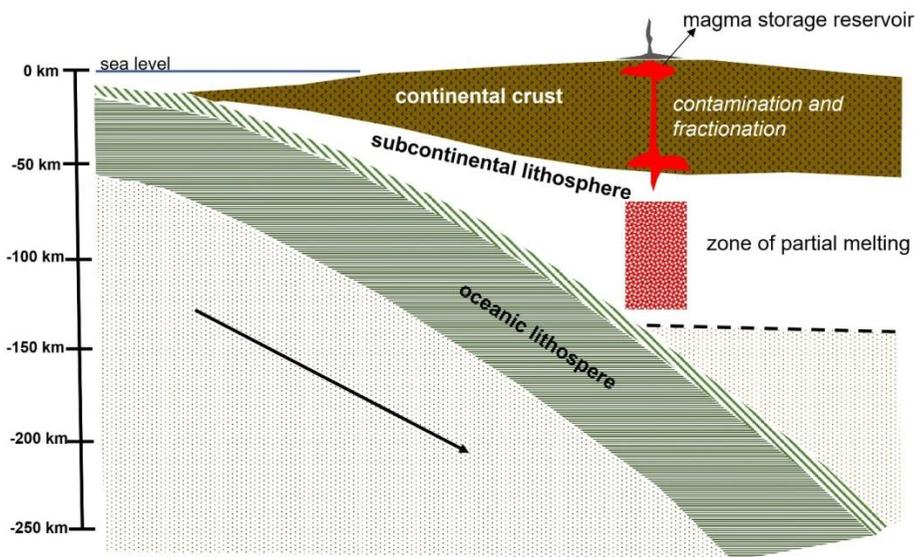


Figure 34 Schematic cross-section of an ideal active continental margin

For the separation of subduction related components from mantle ones, bivariate diagrams using trace element ratios seem to be useful. In the Th/Yb vs Ta/Yb diagram, for example, active continental and island-arc basalts can be separated from mid-ocean ridge and intra-plate ones; moreover, the fields of basalts from oceanic island arcs and active continental margins can be distinguished (**Figure 35**).

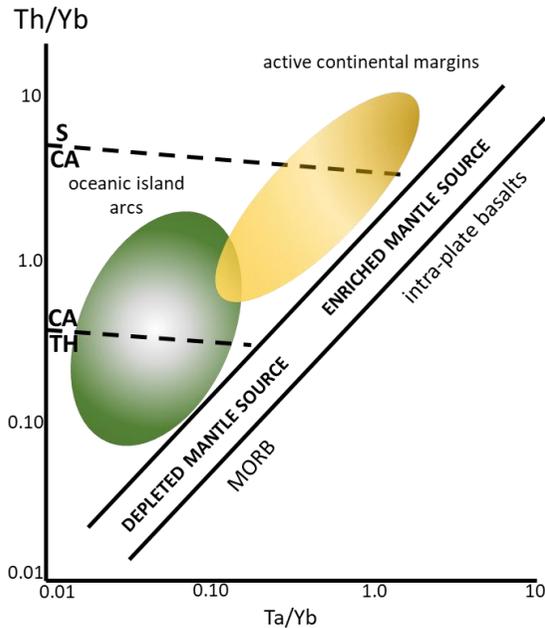


Figure 35 *Th/Yb vs Ta/Yb diagram for discriminating subduction-related and intra-plate basalts*

Although, there are important links between subduction related magmatism (e.g., transportation of incompatible elements to the mantle wedge), geochemical composition of magmas at active continental margins are modified by the assimilation of different (upper, lower, young, or ancient) crustal components. Considering volcanism in the **Cenozoic Volcanic Zone (CVZ)** of the Andes, there is good correlation between crustal thickness and the chemistry of the volcanic rocks.

3.6 Magmatism at back-arc basins

Back-arc basins are considered to be formed by processes quite similar to sea floor spreading occurring at mid-ocean ridges, however, spreading here is associated with subduction. Hence, fluids from the subducting slab can take part in magma generation (although their involvement is not necessary), therefore, back-arc basin basalts may show a transitional character between basalts produced at mid-ocean ridges and active continental margins.

Back-arc basins may develop in an oceanic island-arc. First, ascending fluids and/or partial melts coming from the subducted oceanic lithosphere may metasomatise and partially melt the mantle wedge, and arc basalt can be produced; then an embryonic back-arc basin can be formed by an upwelling of

deep asthenospheric mantle diapir; finally, the arc crust is separated into two parts by the developed back-arc basin (**Figure 36**).

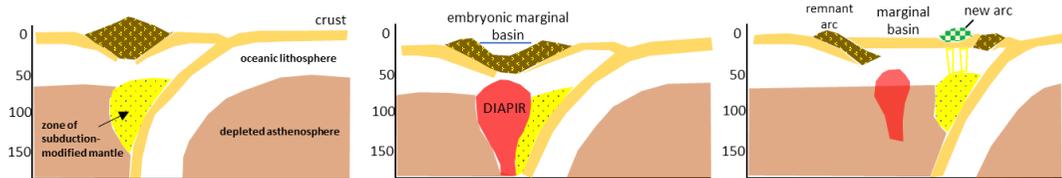


Figure 36 Models for back-arc basin development. Normal magmatism (left) resulted by partially melted asthenospheric mantle wedge; ascending diapir causing cease of magmatism and formation of an embryonic marginal basin (middle); development of the marginal basin re-establishing subduction-related magmatism (right)

Back-arc basin basalts are porphyritic rocks containing olivine, Ca-rich clinopyroxene, plagioclase, chromite and titanomagnetite microcrysts, and large forsteritic olivine and anortite-rich plagioclase phenocrysts; the groundmass is generally fine-grained, or even glassy. **Boninites** containing glassy olivine, and pyroxene and characterised by the absence of feldspars are the typical rocks in the fore-arc region.

Regarding major element compositional pattern of back-arc basalt, it can be stated that some of the rocks show low-K character similar to mid-ocean ridge basalt, however, others have higher potassium content. Similarly, back-arc basalts with high alumina character are more K-rich, while tholeiitic ones have less K_2O . This may suggest that a mantle source, which had been modified by subduction, could be involved in their evolution. Boninites have relatively high (more than 55%) SiO_2 , and high (more than 9%) MgO , and quite low TiO_2 content. It is generally accepted that boninitic magma is not fractionated, but originated from near-primary partial melts, possibly depleted harzburgite sources. In general, the trace element pattern of back-arc basalts shows features characteristic for both mid-ocean ridge and ocean-arc basalts. It seems that in the early stage of the development of back-arc basins subduction-zone fluids have the greatest influence on petrogenesis, however, this influence decreases in the later stages.

3.7 Magmatism of oceanic islands

Oceanic-island volcanism produces one of the most surprising natural phenomena of the Earth's crust: from the base the height of a volcanic oceanic island may reach 10 km, so it is more than the height of Mount Everest from sea level. Unfortunately, our direct knowledge about their foundation is quite poor.

In the Atlantic there are single volcanic islands or, at most, small groups of islands, while in the Pacific they occur in linear chains. It seems that this is in connection with the different spreading rates (low in the Atlantic and fast in the Pacific). According to the generally accepted model, fixed sources in the mantle (hot spots or mantle plumes) feed the volcanoes. The magma supply will be cut because of the motion of the plate, however, new volcanism will begin near the old one. In this way, chains of extinct and active volcanoes have been created. Up till now, more than fifty hot spots have been identified within the oceanic plates most of which can be found near mid-oceanic ridges.

The above mentioned model seems to explain the main features of origin of oceanic islands, however, some important concerns are quite unclear. First of all, geochemical studies of oceanic island basalt revealed that primordial mantle plumes cannot be the single source of oceanic-island basalt, but components originated from the subducted lithospheric slabs should be also involved. The hypothetical mantle plumes are also problematic. There is no direct evidence for their existence, and there are neither geophysical nor geological methods, to prove that. Consequently, we do not have any information about their ascent, shape, size, etc. Concerning the depth of their origin more assumptions exist: some scientists suggest that they rise from the seismic discontinuity at the depth of 670 km; others believe that most of the components of the mantle plumes derived from near the core/mantle boundary.

Chemical composition of the primary basaltic magma of the mantle plume is basically determined by the geochemical and mineralogical composition of the mantle source. On the basis of the geochemistry of oceanic-island basalts, it seems that multicomponent sources are involved, that is near primordial mantle, depleted asthenosphere and lithosphere, recycled ancient oceanic crust should be considered.

The degree and mechanism of the partial melting of the source are also determining factors, however, there are controversial data concerning this point. For example, estimations based on trace element data suggest 2-10% for the degree of partial melting which seem to be quite unreasonably low values; on the basis of melting experiments, however, a degree of partial melting of 20-30% for generating tholeiitic oceanic-islands basalts could be assumed, which seems to be more reasonable; alkalic oceanic-islands basalt generation requires smaller, 5-15% degree of partial melting.

Geochemistry of primary magmas largely depends on the depth of the segregation of magmas. According to a hypothetical thermal model for mantle plumes, the segregation of oceanic-island basalts may occur at depths of 100 km or less, i.e., near the base of the lithosphere; a study on an intense earthquake beneath the **Kilauea volcano** (Hawaii) suggests the depth of 50-60 km as an upper limit. If, however, the depth of segregation is near 100 km the magma should re-equilibrate with the lithosphere during its transportation. Indeed, all of the larger oceanic-island volcanic structures have high-level magma chamber

systems, where further magma modifying processes, (for example fractional crystallisation, crustal contamination and magma mixing) may occur. Among the magmas of oceanic-island volcanism both tholeiitic and alkali magmas can be found (**Figure 37**); the essential differences between them are listed in **Table 3**).

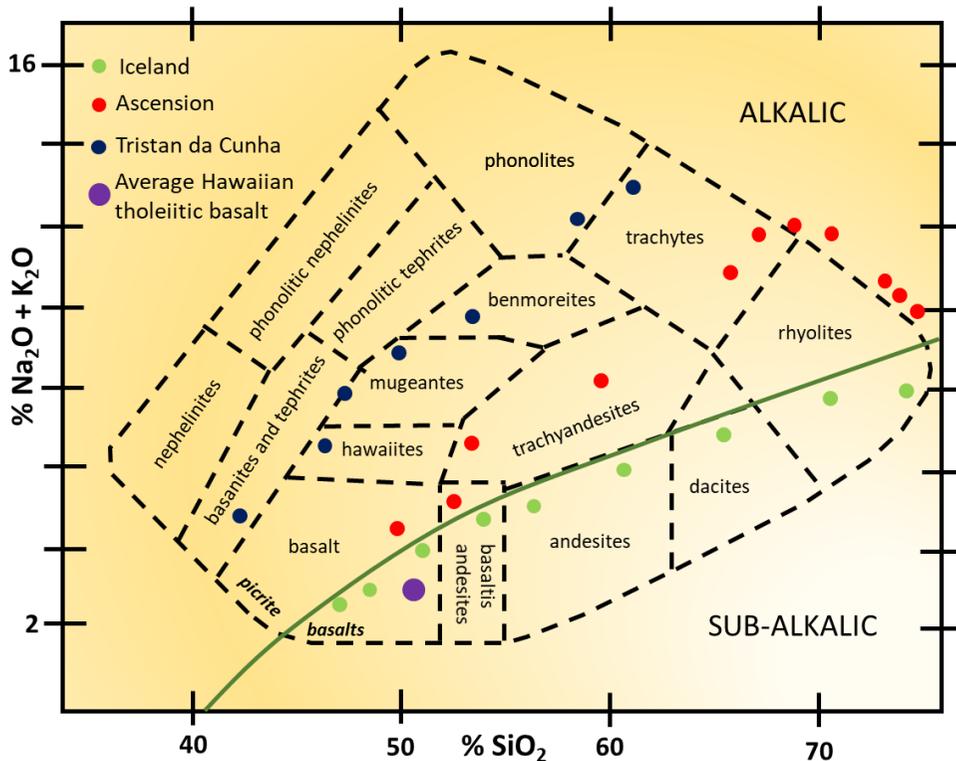


Figure 37 Alkali versus SiO_2 diagram showing oceanic island alkalic and oceanic island tholeiitic series (below the green line)

Spinel is a frequent mineralogical component in both series of oceanic-islands basalts, however, higher Cr_2O_3 contents are characteristic for tholeiitic ones. Olivine occurs in both series, as well, however, tholeiitic basalts contain olivine only as a phenocryst phase; in alkali basalts olivine can be found both as groundmass constituent and as phenocryst. Tholeiitic basalts contain both clinopyroxenes and orthopyroxenes, while alkali basalts contain only titaniferous augite. Plagioclase is present in both series, although it is less common in the phenocryst phase of the alkali basalts. Groundmass of the oceanic island alkalic basalt shows high mineralogical complexity; beside the minerals of the phenocryst phase (olivine, titaniferous augite) alkali feldspars and feldspathoids may also occur. Interstitial glass is rare or absent in alkali basalts, however, it can be usually found in tholeiitic basalts.

Table 3 Major and trace elements spectra of tholeiitic and alkali basalts from Hawaii

	Tholeiitic basalt (Kilauea)	Alkali basalt (Hualalai)
(%)		
SiO ₂	50.51	46.37
Al ₂ O ₃	13.45	14.18
Fe ₂ O ₃	1.78	4.09
FeO	9.59	8.91
MgO	7.41	9.47
CaO	11.18	10.33
Na ₂ O	2.28	2.85
K ₂ O	0.49	0.93
MnO	0.17	0.19
TiO ₂	2.63	2.40
P ₂ O ₅	0.28	0.28
H ₂ O	--	--
(ppm)		
La	13.4	18.8
Ce	35.5	43.0
Sm	6.14	5.35
Eu	1.88	1.76
Yb	1.98	1.88
Rb	9.2	22
Sr	371	500
Ba	150	300
Hf	4.39	3.0
Zr	115	166
Nb	17	16
Y	25	21
Th	1.27	1.2
Pb	5	1

Considering the alkali basalt of oceanic islands, two differentiation trends can be distinguished: (1) the most common undersaturated trend where the final differentiation products are phonolites with nepheline, and (2) the less common oversaturated trend, where the ultimate products are quartz-bearing alkali rhyolites.

Compared to mid-oceanic ridge basalts, oceanic-island basalts are generally enriched in large low-valence cations, such as Cs, Rb, K, Ba, Pb, and Sr. However, Sr and Ba can be regarded incompatible, thus variation of their quantity is in connection with the low-pressure fractional crystallization. Low K/Ba ratios for oceanic-island basalts compared to K/Ba ratios for mid-ocean ridge basalt obviously indicate that the partial melting of different mantle sources played a key role in their origin. Large cations of high-valence, such as Th, U, Ce, Zr, Hf, Nb, Ta, Ti are incompatible, and tend to concentrate rather in oceanic-island basalts than in mid-oceanic ridge basalts. Consequently, Zr/Nb ratio for oceanic-island basalts is characteristically lower than for N-type mid-oceanic ridge basalts; therefore, it can be used for demonstrating the mixing of a mantle source in the petrogenesis of oceanic-island basalts. Oceanic-island alkali basalts generally contain significantly less Ni and Cr comparing to either oceanic-island tholeiites or mid-ocean ridge basalts suggesting that significant fractional crystallisation occur in the ascending magma. Compared to N-type mid-ocean ridge basalts, tholeiitic ocean-island basalts are enriched in light-REE referring to a relatively undepleted mantle source; for alkali basalt even greater enrichment in light-REE and depletion in heavy-REE may indicate that they derive from a source with residual garnet, and which suffered relatively small degree of partial melting (**Figure 38**).

Isotopic ratios of Sr, Nb, and P are highly variable indicating that the plume source of the oceanic-island basalts contains components from both the enriched mantle and the subducted (and then recycled) oceanic lithosphere.

Figure 39 summarizes the processes which play key role in the petrogenesis of oceanic-island basalts. Partial melting is induced by the upwelling mantle diapir in the upward flow containing both plume and depleted asthenospheric components.

The magma derived from the mixing of these partial melts segregates at a depth of 50-60 km. These magmas, which can be regarded as primary ones, will be transported through the oceanic lithosphere, and then they reach the network of high-level magma reservoirs beneath the volcanic island generally at a depth of less than 15 km. In these reservoirs, crystal fractionation and magma mixing occur, and, additionally, may be contaminated by oceanic sediments, altered oceanic crust, altered oceanic volcanics, etc. In this way, magma is chemically further modified.

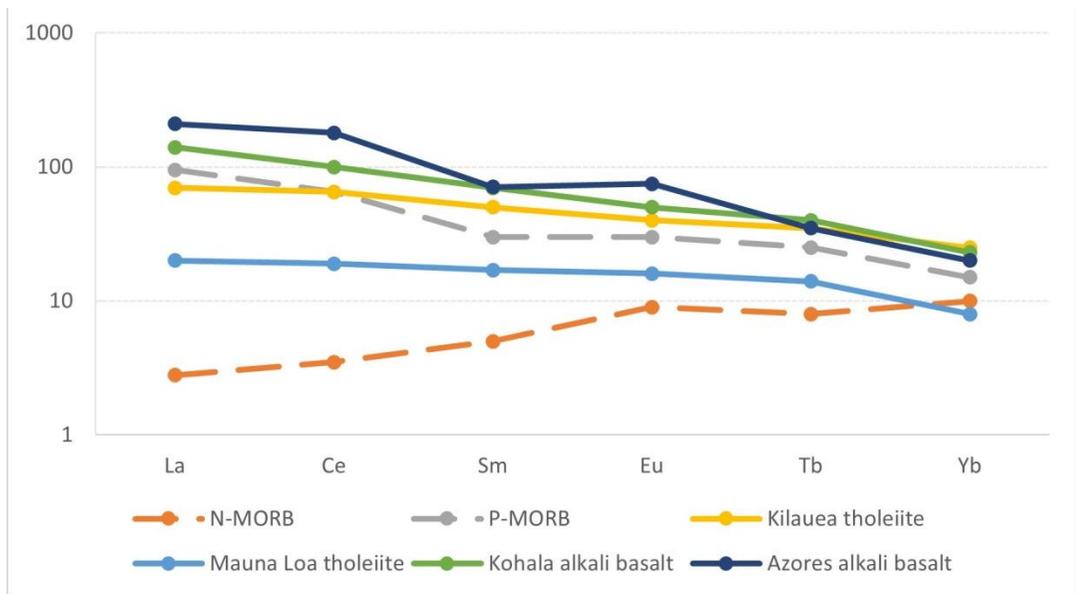


Figure 38 Variations of chondrite-normalized REE abundances in oceanic island tholeiitic and oceanic island tholeiitic basalts

3.8 Magmatism of continental tholeiitic flood basalts

Continental flood basalt provinces (CFBs) are extended areas covered by basalt flows of considerable thickness. The basalts are dominantly tholeiitic, mineralogically and geochemically similar to mid-ocean-ridge basalts, although subordinate (less than 10%) acidic volcanics also occur in some provinces. Regarding their age, it ranges from Precambrian to Quaternary.

The most extensive continental flood basalt provinces are listed in **Table 4**. Some of them (Keweenawan, Siberian Platform) associated with rift zones, while others (Karoo, Paraná, Etendeka, etc.) seem to be associated with ocean basin generation; the youngest one (Snake River Plain) is thought to be produced by hot spot volcanism. It seems to be clear, therefore, that the genesis of continental flood basalts is associated with extensional tectonic activities. One of the main problems is that there is no recent example for continental flood basalt activity; consequently, their evolution has been in focus of scientific debates for decades. The main problems are (1) to characterise the primary magmas, and (2) to estimate the extent of continental crustal contamination.

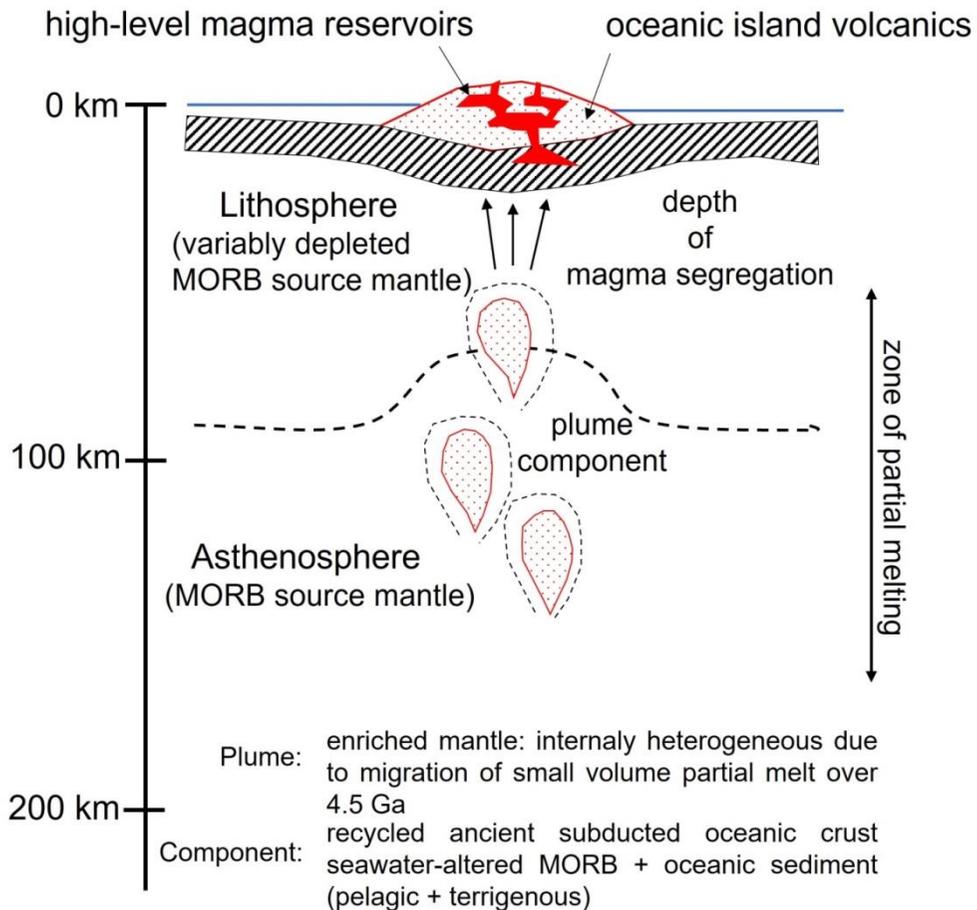


Figure 39 Petrogenetic model for oceanic island basalts.

In general, flood basalts are porphyritic, subordinately aphyric or sub-aphyric. The most common and frequent mineral component is plagioclase, moreover, augite, pigeonite, more rarely olivine and Ti-magnetite can be found as phenocrysts. In the groundmass, plagioclase, augite, pigeonite, Ti-magnetite and olivine occur. The presence of Ca-rich (augite) and Ca-poor (pigeonite) clinopyroxenes as well as the rarity of olivine is typical for most tholeiitic series; Ca-poor pyroxene may indicate continental crust contamination during the ascent. Tholeiitic andesites and dacites are associated with continental flood basalts, olivine is absent, while quartz may occur in the groundmass. Concerning major element content, continental flood basalts are similar to mid-oceanic ridge and oceanic island-arc tholeiites. According to their potassium content, continental flood basalts are ranging from low-K to alkalic, in general, however, most of them fall into the sub-alkalic field (**Figure 40**).

Table 4 Ages, thickness and area of the major continental flood-basalt provinces

Province	Age (Ma)	Maximum thickness (m)	Area (km ²)
Keweenawan (Lake Superior)	Late Precambrian (1100-1200)	12000	>1000000
Siberian Platform	Permo-Triassic (248-216)	3500	>1500000
Karoo /South Africa)	Jurassic (206-166)	9000	140000
Kirkpatrick Basalts, Ferrar Dolerites (Antarctica)	Jurassic (170-210)	900	7800
Paraná (Brazil, Etendeka (Namibia)	Late Jurassic-Early Cretaceous (140-110)	1800	1200000
North Atlantic Igneous Province	Late Cretaceous-Eocene (65-50)	2000	1000000
Deccan Traps (India)	Cretaceous-Tertiary boundary	>2000	>500000
Columbia River (NW USA)	Miocene (17-6)	>1500	200000
Snake River Plain	Quaternary	-	50000

Due to restricted silica content, MgO is used as a differentiation index. Variation of CaO versus MgO for continental flood basalts shows positive correlation suggesting crystal-liquid control by plagioclase and calcium-rich clinopyroxenes during crystal fractionation and partial melting, respectively. In general, Mg' values [Mg/(Mg+total-Fe)] for flood basalts are relatively low (less than 0.7), and their nickel content is also low. These data, as well as CaO-MgO correlations may suggest low-pressure fractional crystallisation with plagioclase and clinopyroxene. Unfortunately, however, major element geochemistry cannot be used for indicating crustal contamination.

Continental flood basalts are generally characterised by low compatible element content suggesting olivine fractionation. Consequently, they are enriched in incompatible elements (except of Y and Yb) compared to N-type mid-oceanic ridge basalts which indicate that they are hardly derived from the mantle source of normal mid-oceanic ridge basalts; rather, similarly to oceanic-island tholeiites, they were contaminated by continental crust in various degrees.

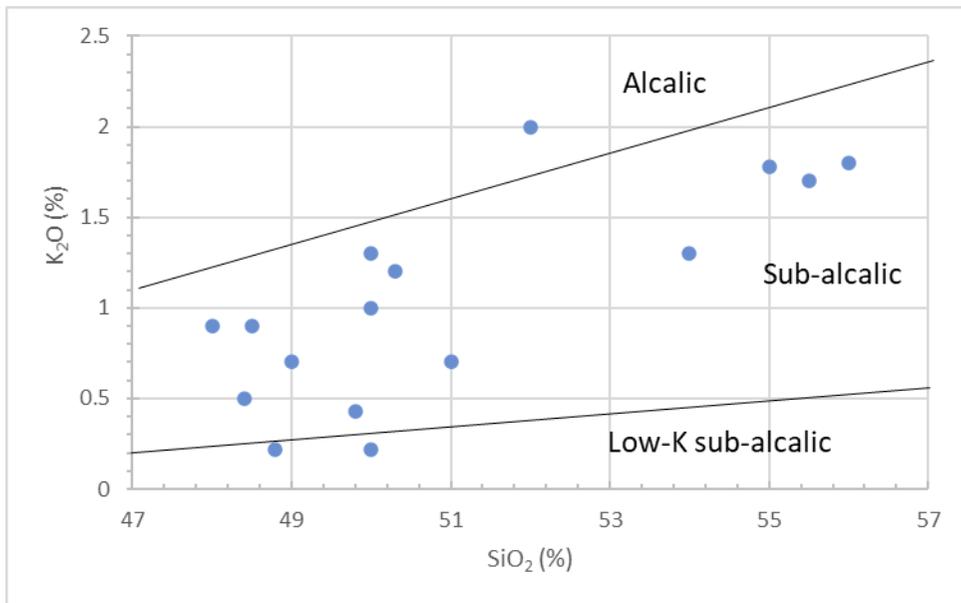


Figure 40 Plot of K_2O vs SiO_2 diagram for Columbia River Basalts

Oceanic-island tholeiite normalised trace element variation diagram for continental flood basalt reveals similarities and differences between them. Their Nb, Ta, P, Zr, Hf, Y and middle REE contents are similar, however, continental flood basalts have more K, Rb, Ba, Th and light rare earth elements. Comparing to mid-oceanic ridge and oceanic-island basalts quantity of mobile incompatible elements (Sr, K, Rb, Ba, Th) in continental flood basalts is increased (**Figures 41, 42**) In general, it can be stated that continental flood basalts can be characterised by higher elemental diversity than oceanic tholeiitic basalts, which can be the result of different processes, such as:

- (1) crustal contamination;
- (2) enriched subcontinental mantle source;
- (3) mixing of depleted and enriched mantle sources;
- (4) combination of the above-mentioned processes.

Moreover, more siliceous magma may be produced by the contribution of crystal fractionation processes and assimilation of crustal rocks. Bimodality of some provinces (that is the presence of rhyolite and basalt without intermediate rocks), however, may indicate that the acid magmas are derived from the base of the continental crust by partial melting of basic rocks.



Figure 41 Diagram showing MORB normalized trace element variations in different continental flood basalts. 1: Snake River Plain; 2: Deccan; 3: Paraná



Figure 42 Diagram showing oceanic island tholeiitic normalized trace element variations in different continental flood basalts. 1: Snake River Plain; 2: Deccan; 3: Paraná

3.9 Magmatism at continental rift zones

Continental rift zones are areas with a central depression and uplifted flanks. Their width, in general, is some tens of kilometres (although the Basin and Range province in the United States reaches a few hundred kilometres), and their length is some hundreds of kilometres. These zones are characterised by lithospheric extension, and, consequently, high heat flow and thinning of the crust and, as geophysical data suggest, asthenospheric upwelling occurs at all active continental rifting areas. Active and ancient rift valleys occur in every continent. Some of them are associated with the evolution of new ocean basins, while others are so-called ‘failed rifts’ aborted after narrow horizontal extension. Considering the processes fundamental for the development of continental rifting, first, the upwelling of lithosphere/asthenosphere boundary should be focused on. According to a simplified model, in the active zones updoming and cracking of the crust is caused by upwelling mantle plumes, thus the tectonomagmatic sequence is the following: (1) doming, (2) volcanism, and then (3) rifting. In the passive case, however, formation of asthenospheric diapirs is caused by the stress triggered cracking of the crust, and the expected sequence is (1) rifting, (2) doming, and, finally, (3) volcanism (**Figure 43**). Indeed, this simplified model is not uniformly and directly fit for the rift development process observed in the field studies. Although most of the recent continental rift zones can be regarded as active ones (probably the Basin and Range Province is the only exception), to make a clear distinction between them is quite uncertain. It seems that lithospheric stretching is characteristic during the initial stages of rifting, and, this way, it may trigger asthenospheric upwelling; in this way, passive rifts may become active ones. Moreover, it is also possible that intra-continental hotspot tracks may weaken the lithosphere, consequently, that areas later may become continental rift zones.

Another crucial point for the magmatism of continental rift zones is the extreme diversity of volcanic products including melilitites, basanites, nephelinites, carbonatites, ultrapotassic rocks, alkalic basalts, etc. In general, magmas are alkaline, volatile-rich and enriched in large-ion lithophile elements (LILE); this pattern suggests an enriched asthenospheric mantle source. It is possible that the old continental lithosphere is the source and the asthenospheric mantle is involved only at the most active rift segments.

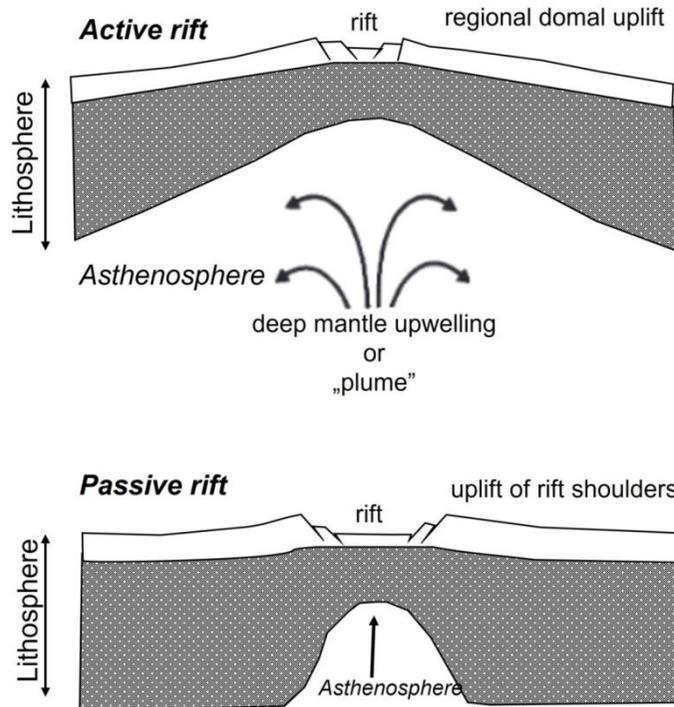


Figure 43 Models of active and passive rifting

Active continental rift zones are characterised by high heat flow (as it was mentioned above) indicating high temperature of the crust, therefore acid magmas may be generated by crustal melting and mixing of partial melts of different chemical composition. The basaltic magmas derived from the mantle may pond in the crust, and may cause partial melting. As a result, silicic melts of low density may form and create extensive silicic volcanism (**Figure 44**). Studies on rocks coming from the East African Rift Zone suggest that three suites can be distinguished from petrographic point of view, such as (1) **basanite-phonolite**, (2) **alkali basalt – trachyte**, (3) **transitional basalt – rhyolite**. Texture of the rocks may vary from aphyric to porphyritic, and minerals of the phenocryst phase dominate the groundmass mineral assemblages, as well. Olivine can be found in each suite with a composition ranging from Fe_{80} (in the basic rocks) to the high fayalitic crystals (in the acid rocks). In phonolites, olivine is not so frequent mineral, and its iron content is extremely high. Ca-rich clinopyroxene also occurs in the wide range of the rocks.

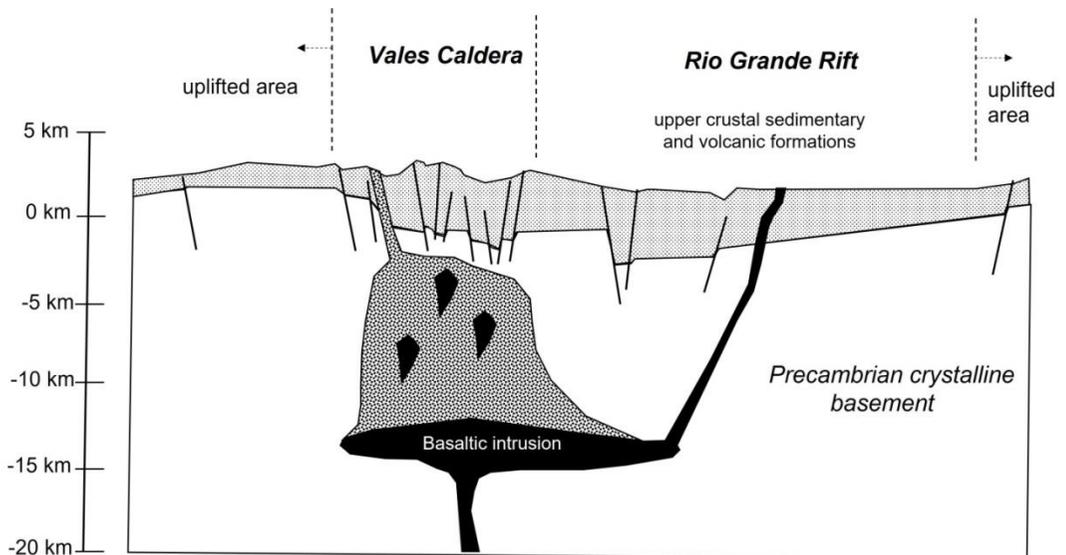


Figure 44 Schematic transection of a crustal magma chamber producing zone of melting. Acidic volcanic rocks feed the caldera above the chamber, while the occasional basaltic flows come from the lower basaltic zone

In basic rocks augite is characteristic, while in trachyte and rhyolite hedenbergite is the most common clinopyroxene. In phonolites aegirine crystallises. Fe-Ti oxides can be found in each of the above-mentioned suites. Amphibole and biotite are characteristic only in the acid rocks of the basanite-phonolite suite (phonolitic tephrite, phonolite). Plagioclase is also common in each suite; its composition is ranging from An_{85-75} (in the basic rocks) to An_{17} (in the acidic rocks). As a phenocryst nepheline and quartz can be found in phonolite and rhyolite, respectively. The ultrabasic ultrapotassic rocks (which occur in the western part of the East African Rift) contain leucite and kalsilite as feldspathoidal minerals.

The basic volcanic rocks of continental rift zones have SiO_2 contents ranging from 37% to 45%, and alkali contents ranging from <4 to <7 , and, except for ultrapotassic lavas, Na_2O is generally higher than K_2O . Silica and alkali content of more acid rocks, however, may exceed 70% and 13%, respectively. Continuous variations of composition from basic to acid rocks as well as bimodal patterns can be observed, as well. Supposedly, a continuous range suggests that fractional crystallisation is a fundamental process in the genesis of acid rocks; however, bimodality may be the result of the varying combination of source heterogeneity, fractional crystallisation, crustal contamination, and degree and depth of partial melting.

The basic end-members of the suites of continental rift volcanism have, in general, quite low compatible trace element concentrations suggesting that olivine fractionation occurred during their genesis. Constant ratios of some incompatible trace elements (such as, Nb/Zr , Ce/Zr , Rb/Zr , etc.) for rocks of

different silica content are also characteristic for certain continental rift zones, especially in East Africa, which proves that fractional crystallisation could be regarded as a dominant process in their genesis. Contrary to this pattern, bimodal basalt-rhyolite suites represent obviously different trends for basalts and rhyolites indicating that other components also contribute to their genesis. Chondrite normalised spider diagram patterns indicate that it is not necessary to assume enriched subcontinental mantle sources for the petrogenesis of continental basalts.

In the areas of continental rift zones, where basaltic cinder-cone fields are characteristic fractional crystallisation and crustal contamination are subordinate in petrogenesis; therefore, rapid uprising and the lack of high-level magma chambers can be assumed. In the area, however, where the central volcanic structure is built up by lavas ranging from intermediate to acid chemistry, fractional crystallisation and assimilation in high-level reservoirs were key processes in the development of the magmas.

3.10 Intra-continental potassic magmatism

Potassium-rich igneous rocks rarely occur in subduction-related and in some oceanic-island series, however, ultrabasic and basic rocks in which K_2O content is higher than Na_2O content, and K_2O/Na_2O ratio is higher than 3 are restricted to within-continental plate-tectonic settings. These igneous rocks are termed as ultrapotassic rocks. Although other criteria can also be considered, some of them, e.g., K_2O content higher than 3 wt. %, would exclude most **kimberlites** and **lamprophyres** which have low total alkali content. Although magmas may generate in different tectonic settings, these rocks can be considered together since they are possibly derived from metasomatized or enriched mantle sources.

Ultrapotassic rocks can be subdivided into three groups: (I) **lamproites**, which are ultrapotassic rocks with high Mg-content, and high K_2O/Na_2O ratio (higher than 5), and, in general, are related to regions where previously continental collision occurred; regarding major element composition, lamproites overlap both kimberlites and lamprophyres; (II) **ultrapotassic rocks of continental rift zones** including *leucitites*, *melilitites*, etc. (see the previous 3.9 chapter on magmatism at the continental rift zones); (III) **ultrapotassic rocks of active orogenic zones** including *leucite-bearing igneous rocks* (mainly in the Mediterranean region) which are related to the complex convergent motion of the African and European plates.

Kimberlites have inequigranular textures composed of macrocrysts and fine-grained matrix. The matrix contains forsteritic olivine, clinopyroxene, monticellite, apatite, calcite, spinel, phlogopite, perovskite, serpentine and ilmenite. The macrocrysts may include olivine (the most abundant) as well as

phlogopite, spinel, garnet, clino- and orthopyroxene. As a very rare additional phase, kimberlite may contain diamond, as well. Kimberlites are subdivided into two groups: the first group has considerable mica in the matrix, while kimberlites belonging to the second group have little or no mica. Lamproites characteristically contain forsteritic olivine, diopside, titanium-rich Al-poor phlogopite, richterite, leucite and sanidine and. Diamond may also occur as an accessory mineral. Analcime, carbonate, zeolite, serpentine are common secondary minerals. Ultrapotassic rocks from the active orogenic zones typically contain clinopyroxenes, leucite and plagioclase as phenocrysts. The groundmass is composed of the same mineral phases, and nepheline and, sometimes, melilite.

In general, these rocks have high K_2O content (higher than 3%) and quite high content of incompatible elements and volatiles (H_2O , CO_2 , F, Cl, SO_2). It is also characteristic that K_2O/Na_2O ratios are higher than 3; Mg values as well as Ni and Cr contents are also high. Lamproites can be characterised by low Al_2O_3 , CaO and Na_2O as well as varying (from 36% to 60%) silica contents suggesting that they derived from a depleted mantle source which later became enriched in potassium and incompatible elements; in general, Mg values are the highest compared to other ultrapotassic rocks. Lamprophyres and kimberlites represent quite similar REE patterns suggesting that both of them were generated from similar sources of the upper mantle. Their different mineralogy, however, obviously indicates that their sources have different geochemical compositions. Ultrapotassic rocks from the Mediterranean region are characterised by high Al_2O_3 content and low Mg values; moreover, these rocks show the lowest enrichments in most of the incompatible elements, however, are enriched in Rb, Th and K. This pattern is quite similar to that of igneous rocks coming from island-arcs and active margins suggesting that these rocks are related to subduction systems.

Regarding the diversity of tectonic settings of the areas where ultrapotassic magmas occur, it seems to be obvious that components of different source components contribute in their petrogenesis (**Figure 45**). Kimberlites belonging to lamproites (group I) have asthenospheric upper mantle sources as it is suggested by Nd-Sr isotopic data. Kimberlites divided into group II derive from the subcontinental lithosphere. Petrogenesis of ultrapotassic magmas at active continental margins involves both asthenospheric and lithospheric mantle sources. Chemical characters of every group of ultrapotassic magmas are basically determined by partial melting processes, however, crustal contamination and fractional crystallisation are also significant in the case of group III volcanic rocks as suggested by Nd-Sr isotopic data.

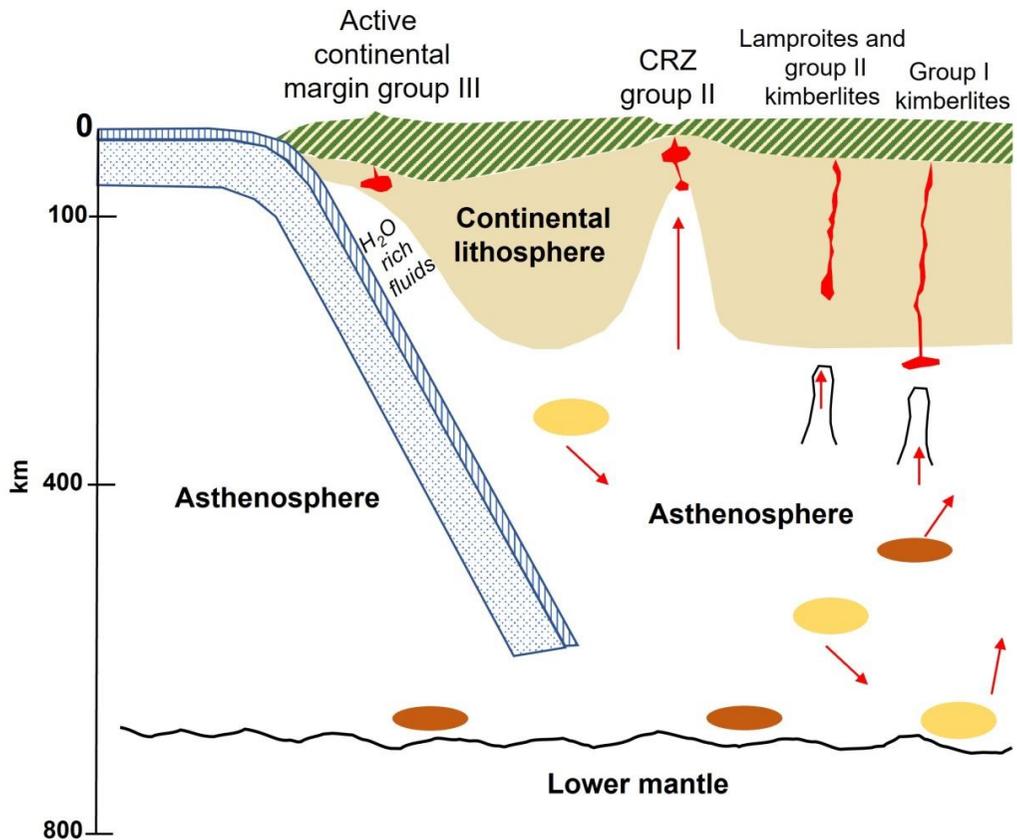


Figure 45 Schematic illustration of tectonic environments and source components of potassic intracontinental plate magmas

3.11 Identifying the tectonic environment of ancient volcanic sequences

Based on the above-mentioned tectonic settings of magma generation and volcanism, four main groups of plate tectonic environments can be defined:

1. Constructive plate margins (mid-ocean ridges and back-arc basins).
2. Destructive plate margins (island arcs and active continental margins).
3. Oceanic intra-plate settings represented by oceanic islands
4. Continental intra-plate settings represented by continental rift zones, continental flood basalt provinces, and intra-continental potassium magmatism not related to rift zones.

The magma series of different classifications seems to be associated with these tectonic environments (**Table 5**). The most common volcanic rock types are the sub-alkalic basalts; alkali basalts (together with their differentiates) occur in all tectonic settings except for the divergent one. Tholeiitic basalts are the dominant

rock type produced by magmatism at divergent (constructive) plate margins and within plate settings. Calk-alkaline rocks are almost completely restricted to magmatism at convergent plate margins, i.e., island-arcs (particularly, more mature arcs) and active continental margins.

Table 5 *Tectonic settings and their characteristic magma series*

Tectonic setting	Plate margin		Within plate	
	Convergent	Divergent	Intra-oceanic	Intra-continental
volcanic feature	island arcs, active continental margins	mid-oceanic ridges, back-arc spreading centers	oceanic islands	continental rift zones, continental flood-basalt provinces
characteristic magma series	tholeiitic, calk-alkaline, alkaline	tholeiitic	tholeiitic, alkaline	tholeiitic, alkaline
SiO ₂ range	basalts and differentiates	basalts	basalts and differentiates	basalts and differentiates

Since, as we can see, there is correlation between particular geochemical parameters of recent igneous rocks and the plate tectonic settings where they occur, it seems to be obvious that the geochemical patterns of ancient igneous sequences could be used for identifying their former tectonic environments. Several so-called ‘tectonomagmatic discrimination diagrams’ have been developed, most of them for discriminating the tectonic setting of different basaltic, and some of them for granitic rocks. In general, major elements are less suitable for applying to this kind of discrimination since there is significant overlap between major element distribution patterns of basalts from mid-ocean ridges, back-arc basins, and volcanic arcs; rather immobile trace and some major (such as Fe, and Ti) and minor elements (for example Ni, Cr, V, Zr, Nb, Ta, Hf) are useful for this purpose. Tectonomagmatic discrimination diagrams can be grouped as it follows:

1. Diagrams using major and minor elements, for example TiO₂–K₂O–P₂O₅, MgO–FeO–Al₂O₃ and TiO₂–MnO–P₂O₅ ternary diagrams (see **Figure 46**); diagrams which use discriminant function calculated from major element data belong to this group, as well.
2. Diagrams using minor elements and immobile trace elements; for example, ternary diagrams Ti/100–Zr–0.3Y belong to this group (see **Figure 47**).

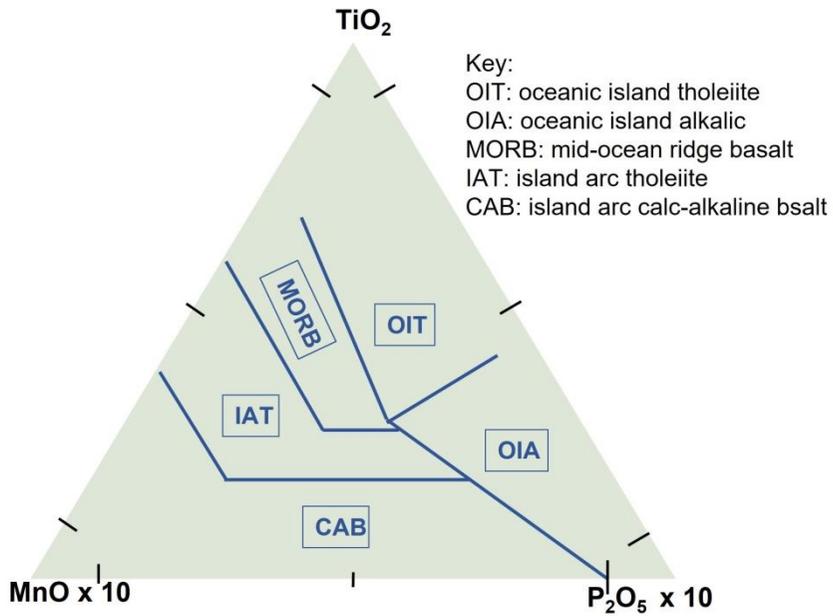


Figure 46 TiO_2 - $MnOx10$ - P_2O_5 triangle diagram for discriminating oceanic basaltic rocks

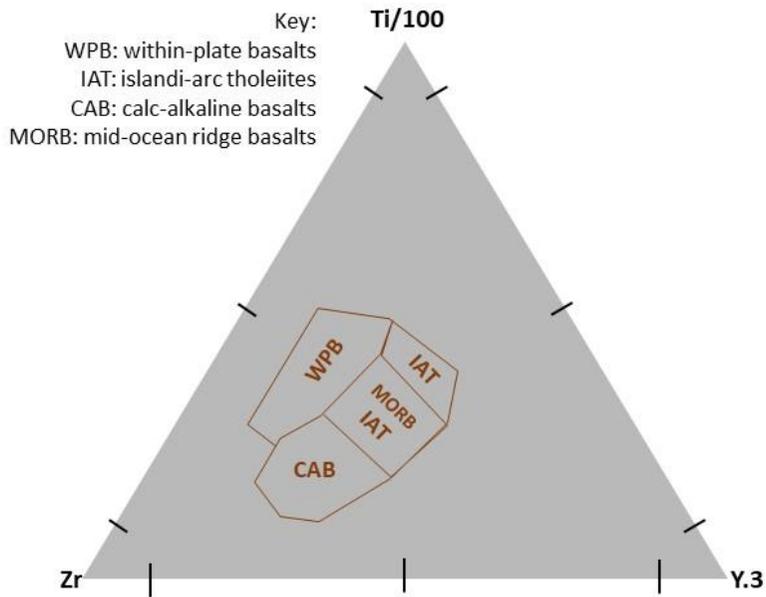


Figure 47 $Ti/100$ - Zr - $0.3xY$ triangle diagram for discriminating basaltic rocks

3. Diagrams using immobile trace elements; for example, Th–Hf/3–Ta ternary diagram, Th/Yb vs Ta/Yb diagram (see Figure 3.9 in Chapter 3.5).

Although magma chemistry is controlled by tectonic settings, too, some considerations should be taken into account. From geochemical point of view there could be so-called ‘mixed signals’; for example, some volcanic rocks may occur in different tectonic settings; similarly, geochemical interaction resulting in similar trace element distribution patterns may take place in different tectonic environments. Consequently, trace element concentrations of volcanic rocks are in closer relation to processes by which than the tectonic environment where they were formed. If processes are in connection with tectonic settings, tectonomagmatic discrimination diagrams are useful tools. Moreover, in the case of very old rocks, discrimination diagrams should be carefully used, too, because it can be assumed that the trace elements spectra of the mantle source have changed, therefore field boundaries compiled based on the geochemistry of recent rocks are not valid for them.

4 FURTHER READINGS

- Agee, C., Bullock, E., Bouvier, A., Dunn, T., Gattacceca, J., Grossman, J., Herd, C., Ireland, T., Metzler, K., Mikouchi, T., Ruzicka, A., Smith, C., Welten, K. and Welzenbach, L. 2015: Categorization of Finds and Falls. *The Meteoritical Bulletin*, 24 February 2015, <https://www.lpi.usra.edu/meteor/docs/falls-finds.pdf>
- Bergman, S.C. 1987: Lamproites and other potassium-rich igneous rocks: a review of their occurrence, mineralogy and geochemistry. In: Fitton, J.G. & Upton, B.G.J. (eds): *Alkaline igneous rocks*. 103–89. Geol. Soc. Spec. Publ. 30. Blackwell, London.
- Best, M.G. 1982: *Igneous and metamorphic petrology*. Freeman, New York.
- Burbidge, E.M., Burbidge, G. R., Fowler, W.A. and Hoyle, F. 1957: Synthesis of the Elements in Stars. *Reviews of Modern Physics* 29 (4), pp. 547-650.
- Cas, R. and Wright, J. 1988: *Volcanic successions modern and ancient. A geological approach to processes, products and successions*. Springer-Verlag, Berlin.
- Cox, K.G., Bell, J.D., and Pankhurst, R.J. 1979: *The interpretation of igneous rocks*. Allen and Unwin, London.
- Elsasser, W.M. 1963: Early history of the Earth. in Geiss, J. and Goldberg, E. (eds.): *Earth Science and Meteorites*, North Holland, pp. 1-30.
- Firsch, W., Meschede, M. and Blakey, R.C. 2011: *Plate Tectonics*, Springer, Heidelberg, Dordrecht, London, New York, 212 p.
- Gill, J.B. 1981: *Orogenic andesites and plate tectonics*. Springer-Verlag, Berlin.
- Gomes, C.B. and Keil, K. 1980. *Brazilian stone meteorites*. Albuquerque, University of New Mexico, 161 p.
- Johnson, D., Tyldesley, J., Lowe, T., Withers, P.J. and Grady, M.M. 2013: Analysis of a prehistoric Egyptian iron bead with implications for the use and perception of meteorite iron in ancient Egypt. *Meteoritics & Planetary Science* 48 (6), pp. 997-1006.
- Kallemeyn, G.W., Rubin, A.E. and Wasson, J.T. 1996: The compositional classification of chondrites: VII. The R chondrite group. *Geochimica et Cosmochimica Acta* 60 (12), pp. 2243-2256.
- Kearey, Ph., Klepeis, K.A. and Vine, F.J. 2009: *Global Tectonics*, Wiley-Blackwell, 496 p.

- Lee C.T. 2018: Geochemical Classification of Elements. In: White W.M. (eds) *Encyclopedia of Geochemistry*. Encyclopedia of Earth Sciences Series. Springer, Cham., https://doi.org/10.1007/978-3-319-39312-4_255
- Lodders, K. and Fegley, J.B. 1998: *The planetary Scientist's companion*. Oxford University Press, Oxford, 370 p.
- Mason, B. 1967: Meteorites. *American Scientist* 55(4), pp. 429-455
- McCall, G. J. H., Bowden, A. J. and Howarth, R.J., 2006: *The History of Meteoritics and Key Meteorite Collections: Fireballs, Falls and Finds*. Geological Society of London, 513 p.
- McDonough, W.F. and Sun, S. 1995: The composition of the Earth. *Chemical Geology* 120 (3-4), pp. 223-253.
- McSween H.Y. 1987: *Meteorites and their parent planets*. Cambridge: Cambridge University, 237 p.
- Norton, O.R. and Chitwood, L.A. 2008: *Field Guide to Meteors and Meteorites*. Springer-Verlag London, 287 p.
- Palmason, G. 1982: *Continental and oceanic rifts*. American Geophysical Union, Washington DC.
- Park, G. 2021: *Introducing tectonics, rock structures and mountain belts*. Dunedin Academic Press, 137 p.
- Prior, G.T. 1920: The classification of meteorites. *Mineralogical Magazine* 19, pp. 51-63.
- Reynolds, M.D. 2001: *Falling stars: a guide to meteors and meteorites*. Stackpole Books, 148 p.
- Ringwood, A.E. 1989: Significance of the terrestrial Mg/Si ratio. *Earth and Planetary Science Letters*, 95 (1-2), pp. 1-7.
- Rollinson, H. 1998: *Using geochemical data: evaluation, presentation, interpretation*. Longman, Harlow.
- Rubin, A.E. and Grossman, J.N. 2010: Meteorite and meteoroid: New comprehensive definitions. *Meteoritics & Planetary Science* 45(1), pp. 114-122.
- Solomatov, V.S. 2000: Fluid dynamics of a terrestrial magma ocean. In: Canup, R.M. and Righter, K. (eds.) *Origin of the Earth and Moon*, Tucson: University of Arizona Press, pp. 323-338.
- Thorpe, R.S. (ed.) 1982: *Andesites: orogenic and related rocks*. Wiley, Chichester.

Weisberg, M. K., McCoy, Timothy J., and Krot, A. N. 2006: Systematics and evaluation of meteorite classification. in: Laurretta, D.S. and Sween, H.Y., Jr., (eds.) *Meteorites and the Early Solar System II*, University of Arizona Press, pp. 19–52.

White, W.M. 2013: *Geochemistry*. John Wiley & Sons Inc, New York, 668 p.

Wilson, M. 1993: *Igneous petrogenesis. A global tectonic approach*. Chapman & Hall, London.

Yang, J., Goldstein, J.I. and Scott E.R.D. 2010: Main-group pallasites: Thermal history, relationship to IIIAB irons, and origin. *Geochimica et Cosmochimica Acta* 74, pp. 4471–4492.

<http://www.earth-site.co.uk/Education/binding-energy/>

<https://esahubble.org>

<https://my.nsta.org/resource/105025/plate-tectonics>

<https://www.webelements.com/>

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