

## Composition of the mantle

These notes are modified from some originally written by Jason Phipps Morgan

Knowing the bulk composition of the mantle is important for several reasons. If we know the mean abundances of the elements that compose prevalent mantle minerals as well as the chemistry of upper mantle rocks that are sometimes exposed (usually in slightly modified form) at the Earth's surface, then we can use mineral physics measurements of likely mantle phases to determine whether the observed depths of seismic velocity discontinuities correspond to phase transitions or, alternatively, to changes in mantle composition. This has important implications for the style of mantle convection, i.e., whether the mantle deforms as stratified layers, or convects in a more top-to-bottom flow pattern. The latter mode of convection implies cooler lower-mantle temperatures and more efficient heat-extraction during the thermal evolution of the Earth.

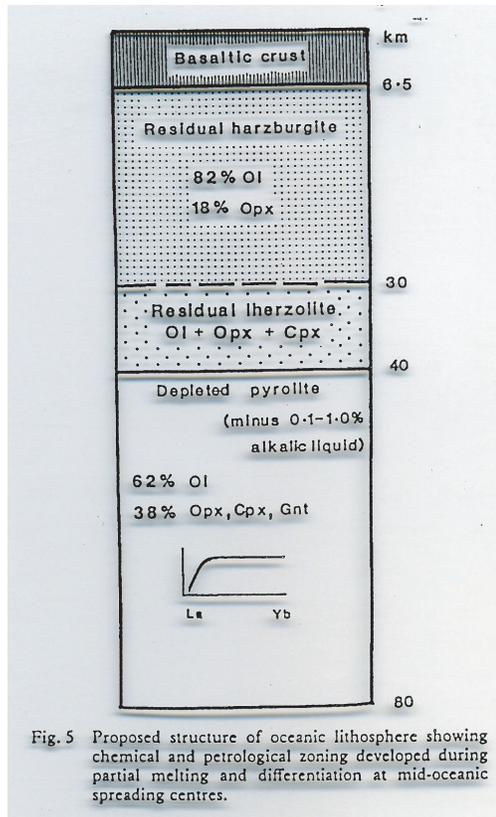
### 1. Mantle composition

Several approaches have been used to try to determine the bulk composition of the Earth, or more specifically, the composition of its crust, mantle, and core. They correspond crudely to: a) looking at (the right mix of) rocks; b) looking at (the right mix of) meteorites; or c) looking at the Sun. The fact that these approaches (crudely) agree is encouraging, but somewhat circular, as the "right mix" cannot be judged without additional assumptions.

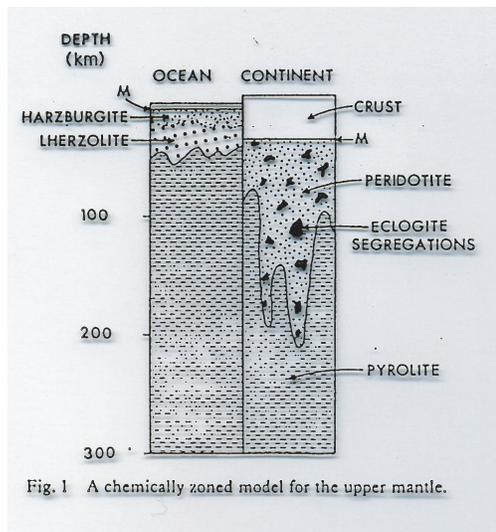
Perhaps the simplest way for a geochemist to determine the mantle's chemistry is to find mantle rocks at the surface and analyze them. There appear to be two different ways to bring mantle rocks to the surface: these expose different kinds of rocks. One way is to bring up slivers of mantle by faulting, usually associated with orogenesis and continental collision/accretion. These rock assemblages, mostly peridotite (a rock made up of 75% olivine and 25% pyroxene minerals) are known as ophiolites. The second way is to bring up mantle xenoliths (foreign rocks) in kimberlites, which are explosive emplacement events found in many continents. Both ophiolite and kimberlite mantle samples are rocks with magnesium-rich olivine, pyroxenes (one Mg-rich, the other more Ca-rich), often plagioclase (a feldspar), and, rarely, garnets in lieu of plagioclase.

Broadly speaking, there are two types of ophiolites. One type has an overlying sequence of basalt (gabbro) rocks, topped by 2km of sheeted dikes, a 1/2 km pillow basalt section (this is the form of submarine basaltic extrusion), and marine sediments (e.g. Troodos Ophiolite in Cyprus, Samail Ophiolite in Oman, Lewis Hills Ophiolite in Newfoundland). These are thought to be created by the obduction (tectonic emplacement) of oceanic crust and lithosphere. Nicolas (Nicolas, 1989) provides a good summary of these ophiolites. The mantle rocks in these harzburgite ophiolites most likely melted beneath a mid-ocean ridge to produce the overlying oceanic crust in the ophiolite section. (Note that these rocks, which lie at the top of the melting column, are likely to have melted more than most rocks would melt beneath a spreading center.) Thus these 'mantle' rocks are not thought to have preserved a typical mantle chemistry, but rather have the chemistry of a mantle residue from partial melting. To get at the mantle chemistry, we need to 'put back in' the melt. This is the basis for Ringwood's pyrolite model of mantle chemistry (Ringwood, 1975). The trick is to convince people that you can constrain how much, and what kind, of melt phase to add to harzburgite to get 'mantle'. A different, but related conceptual approach to correcting mantle rocks for melt extraction was proposed by Jagoutz et al; (Jagoutz et al, 1979) and later reassessed by Hart and Zindler (Hart and Zindler, 1986, Zindler and Hart, 1986). The fact that Ringwood and Jagoutz et al, and Hart and Zindler approaches lead to similar major element chemistries is encouraging as they use different meteorite/mantle assumptions to correct their (almost-) mantle samples.

Harzburgite ophiolite sections show evidence that they have experienced 20% melt extraction. They are barren: further melting of these rocks does not produce mid-ocean ridge basalts. There is another class of ophiolites, Alpine peridotites, which have a poorly developed, often absent, overlying basaltic



A petrological model for oceanic lithosphere based on ophiolites



A petrological model for continental and oceanic lithosphere

sequence. These rocks are usually more fertile (i.e. have more Ca+Al, as well as higher concentrations of other elements that preferentially partition into the melt phase, as well as a lower melting temperature). When melted in the lab, Alpine lherzolites produce a basaltic melt like that of the oceanic crust, but with somewhat lower concentrations of incompatible elements (elements that prefer the melt phase). The technical difference between an Alpine peridotite and an ophiolite is the presence or absence of the basaltic

sequence: ophiolites have both the mantle sequence as well as at least hints of past overlying crust. However, since all are tectonically emplaced, it seems that a minority of workers view Alpine peridotites as mantle rocks (subcontinental mantle) that never had overlying basaltic crust. Alpine peridotites and ophiolites occur fairly frequently in regions of continental collision. For example, roughly 30 of these terrains have been found in the well-surveyed Mediterranean orogenic belts. However, the average outcrop scale is small, usually less than 10 km<sup>2</sup> in extent. Because the melts obtained by melting lherzolites at pressures corresponding to 25–75km depths by 5–20% have a MORB (Mid-Ocean-Ridge-Basalt)-like chemistry, the most fertile lherzolites are viewed as proto-mantle samples that have experienced the least amount of melting during ascent/emplacement. Some kimberlite ultramafic nodules (Ol+Mg pyroxene+Ca pyroxene+Garnet) have similar, usually even more fertile compositions.

**Table 5-2 PYROLITE COMPOSITIONS**

Model compositions	Model compositions					Limiting compositions		Average mantle pyrolite (8)
	(1)	(2)	(3)	(4)	(5)	(6)	(7)	
SiO <sub>2</sub>	45.2	44.9	46.1	45.6	42.9	44.9	46.5	45.1
TiO <sub>2</sub>	0.7	0.24	0.2	0.2	0.2	0.1	0.2	0.2
Al <sub>2</sub> O <sub>3</sub>	3.5	4.3	4.3	3.9	5.8	3.2	3.6	4.6
Cr <sub>2</sub> O <sub>3</sub>	0.4	0.4	—	0.4	0.2	0.5	0.4	0.3
Fe <sub>2</sub> O <sub>3</sub>	0.5	—	—	—	0.3	—	1.0	0.3
FeO	8.0	8.2	8.2	5.2	8.9	7.6	9.4	7.6
MnO	0.14	0.1	—	0.1	0.14	0.1	0.2	0.1
NiO	0.2	0.2	—	0.3	0.2	0.3	—	0.2
MgO	37.5	38.9	37.6	37.7	37.2	40.0	33.0	38.1
CaO	3.1	2.5	3.1	3.0	3.7	3.0	5.1	3.1
Na <sub>2</sub> O	0.6	0.23	0.4	0.5	0.4	0.2	0.5	0.4
K <sub>2</sub> O	0.13	0.02	0.03	0.1	0.003	0.0006	0.2	0.02
P <sub>2</sub> O <sub>5</sub>	0.06	0.02	—	0.03	—	—	0.01	0.02
								100.0

**Explanation:**

- (1) Pyrolite model composition (Ringwood, 1966a).
- (2) Pyrolite model composition: 99% Lizard peridotite (Green, 1964) + 1% nephelinite (VV3-1 Kay, 1971).
- (3) Pyrolite model composition: 83% residual harzburgite + 17% primitive oceanic tholeiite (Table 5-5).
- (4) Bulk composition of Vourinos ophiolite complex (Table 3-3, average of A and B).
- (5) Upper mantle model composition from Carter (1970, table 4) corresponding to a parental mantle olivine composition of Fo<sub>44</sub>.
- (6) Least fractionated alpine ultramafic, representing smallest degree of partial melting of pyrolite (Tinaquillo peridotite, Table 3-3).
- (7) Ultramafic liquid representing highest degree of partial melting of pyrolite (Table 5-2).
- (8) Mean of columns 2,3,4, and 5. Alumina content is probably too high because of column 5 (see text).

Table 1. (from Ringwood, 1975)

The problem with using any of these samples to infer the composition of the mantle is that almost all of these samples show clear evidence for prior melting and melt extraction. This is also extremely likely to happen during ascent and emplacement. The solidus temperature of lherzolitic rocks increases with pressure, roughly 2–3 degrees/km of overburden. Thus the solidus temperature at the surface is 100–150 degrees lower than at 50km depths. Since conduction is much less efficient than advection, mantle ascent can be regarded as adiabatic at ascent rates even as low as 1mm/yr. The time scale for conduction is  $L_a^2/\kappa$  where  $\kappa = 1\text{mm}^2/\text{s}$  for mantle silicates, the time scale for advection =  $L_a/U$  where  $U$  is the advective speed and  $L$  is the length scale. Thus, during 50km ascent in 50my (an extremely slow orogenic emplacement rate of 1mm/yr or  $\pi \times 10^{-11}$  m/s, the length scale of conductive transport during this time is only 40 km, i.e., the rock is not emplaced at conductive equilibrium. Only if emplacement rates are lower than this (i.e. 0.1–1mm/yr) is the rock able to be emplaced without emplacement-related melting.

The corrections for this 'missing' melt take two forms. The first is to compare REE (rare-earth-element) abundance profiles in the mantle rocks with those in what are presumed to be low-degree basaltic melts.

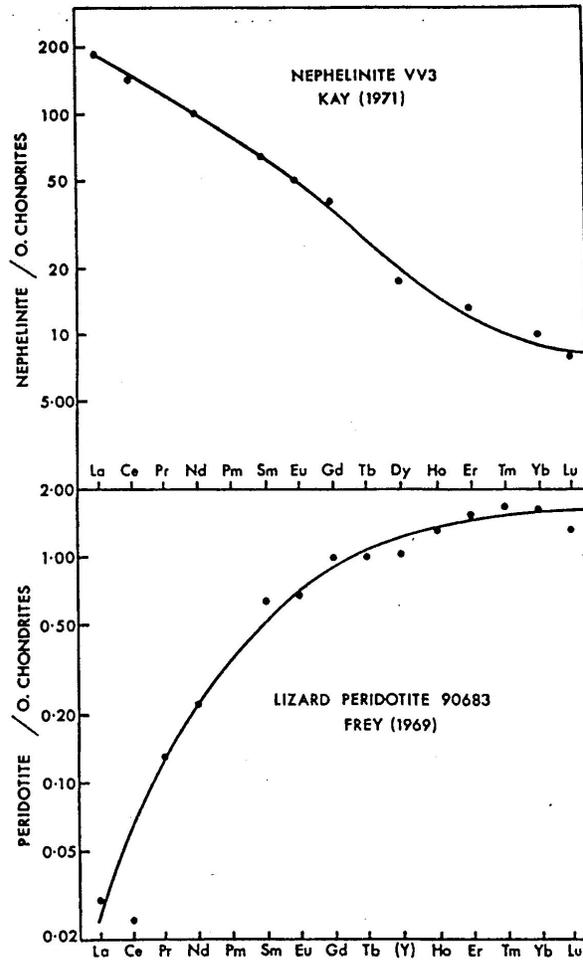


FIGURE 5-2  
Rare earth abundances in the Lizard high-temperature peridotite (after Frey, 1970a) and in a Cape Verdes nephelinite (VV3-1, after Kay, 1971). Rare earth abundances have been divided, element by element, by the corresponding abundances in ordinary chondrites.

Figure 1. (from Ringwood, 1975)

REE are Lanthanide-series trace elements that are strongly partitioned into the melt phase of a basalt melt + lherzolite assemblage. Since REE have chemical properties that smoothly (and only slightly) vary with atomic number, they differentiate in a very regular and systematic fashion during melting. The idea is to assume that the Earth's mantle originally had a relative REE abundance distribution like that of chondrites (normalizing REE abundances by chondritic abundances makes for a smoothly varying REE pattern in basalts and lherzolites). Then by comparing the REE patterns of low-degree melts with lherzolites (see fig 1), we add the right amount of low-degree melt to make the combined lherzolite + melt REE pattern look like that of chondrites. Other element abundances are then inferred from the bulk chemistry of this mixture. Variants on this scheme assume the mantle has a particular (non-chondritic) REE pattern and follow the same recipe. Ringwood was the original proponent of this technique (see column 2 of table 1)

Another approach is to adjust the mantle chemistry for melting by comparing proto-mantle sample Mg/Si and Al/Si ratios with those of chondrites, and using this information to back-correct the presumed mantle rocks for the effect of melting. This approach (e.g. Jagoutz et al, 1979) uses a completely different major element comparison of peridotites and meteorite chemistry to back-correct presumed mantle rocks for the effects of melting. This method assumes that the Earth is made of a mix of chondrites. Thus Earth's major element chemistry lies somewhere along the chondrite trend. The Earth trend will not be the same

Table 2 LOSIMAG CI Earth model

	B <sup>b</sup>	C	D	E	F	G
SiO <sub>2</sub>	18.58	18.58	31.96	18.58	31.48	45.96
Al <sub>2</sub> O <sub>3</sub>	1.640	1.64	2.82	1.64	2.78	4.06
FeO	24.49	3.05	5.25	3.05	5.17	7.55
MgO	15.27	15.27	26.26	15.27	25.87	37.77
CaO	1.298	1.298	2.23	1.298	2.20	3.21
Na <sub>2</sub> O	0.134	0.134	0.230	0.134	0.227	0.331
K <sub>2</sub> O	0.013	0.013	0.022	0.013	0.022	0.032
Cr <sub>2</sub> O <sub>3</sub>	0.189	0.189	0.325	0.189	0.320	0.467
MnO	0.053	0.053	0.091	0.053	0.090	0.131
TiO <sub>2</sub>	0.073	0.073	0.126	0.073	0.124	0.181
NiO	1.39	0.112	0.193	0.112	0.190	0.277
CoO	0.064	0.0052	0.009	0.0052	0.009	0.013
P <sub>2</sub> O <sub>5</sub>	0.008	0.008	0.014	0.008	0.014	0.020
Silicate sum	63.202	40.425	69.530	40.425	68.50	100
HVE <sup>a</sup>	30.21	30.21	—	30.21	—	—
MVE <sup>a</sup>	1.258	1.258	—	1.258	—	—
SVE <sup>a</sup>	5.33	5.33	—	5.33	—	—
Oxygen <sup>a</sup>	—	5.06	—	4.185	—	—
	36.798	41.858	—	40.983	—	—
Fe	—	16.665	28.664	16.116	28.239	89.648
Ni	—	1.004	1.727	1.004	1.700	5.397
Co	—	0.046	0.079	0.046	0.078	0.248
Oxygen <sup>a</sup>	—	—	—	0.875	1.483	4.708
Core sum		17.715	30.470	18.590	31.50	100
Sum all	100	100	100	100	100	

<sup>a</sup> HVE = highly volatile elements (H<sub>2</sub>O, S, C, organics); MVE = moderately volatile elements (Na, K, P, Cr, Mn); SVE = slightly volatile elements (Mg, Si); oxygen is that obtained from FeO, NiO, CoO reduction.  
<sup>b</sup> Column headings are as follows:  
 B—Major components of CI chondrites from Anders & Ebihara (1982) all HVE separated from silicate components; partial separation of MVE, to leave MVE/Al ratios in silicate similar to those of Jagoutz et al (1979) mantle; partial separation of SVE, to leave SVE/Al ratios as in PUM, Table 1.  
 C—Separation of metallic core, leaving enough FeO to provide MgO/(MgO + FeO) = 0.90 (molar) and enough NiO and CoO to provide ratios to Al similar to those of Jagoutz et al (1979) in mantle.  
 D—Removal of HVE, MVE, and SVE and oxygen from C; remainder renormalized to 100%.  
 E—Column C, with partial oxygen budget left in core.  
 F—Removal of HVE, MVE, SVE and part of oxygen from E, remainder renormalized to 100%.  
 G—Silicate mantle and oxide core renormalized separately to 100%.

Table 2. (from Zindler and Hart, 1986)

as the chondrite trend since the chemistry of peridotites has been modified by post-accretion melting, melt extraction, and possible additions of intrusive melts (this last process is called metasomatism). If both Earth and chondrite fractionation processes produce linear trends, and Earth is made from a subset (any subset) of chondrites, this approach should work correctly.

If we take the mantle trend for Mg/Si vs. Al/Si and plot it against the meteorite trend for these elements, they intersect at a point which is inferred as the original mantle ratio of these elements. Once we know these ratios, then we use relationships between the ratios of other elements to these elements (e.g. most mantle melts have chondritic Ca/Al ratios, etc.) to constrain other element abundances. This approach (see fig 2), as slightly adjusted by Hart and Zindler (1986), leads to the mantle composition known as Losimag shown in table 2.

Ringwood also explores the consequences of taking mantle harzburgites (which experience 10–20% melt extraction during melting beneath a mid-ocean ridge), and adding 10–20% MORB to these to make 'original' mantle pyrolite. Finally, Ringwood also explores the consequences of using the chemistry of the least-depleted mantle xenoliths or of using ultramafic melts that are thought to be the highest degree melts from the mantle. These latter techniques are probably less reliable since it is hard to see how the mantle can melt by more than about 25% during pressure-release melting. All of these estimates are shown in table 1.

Other approaches to determining the Earth's chemistry try to use meteorite or stellar abundances (with

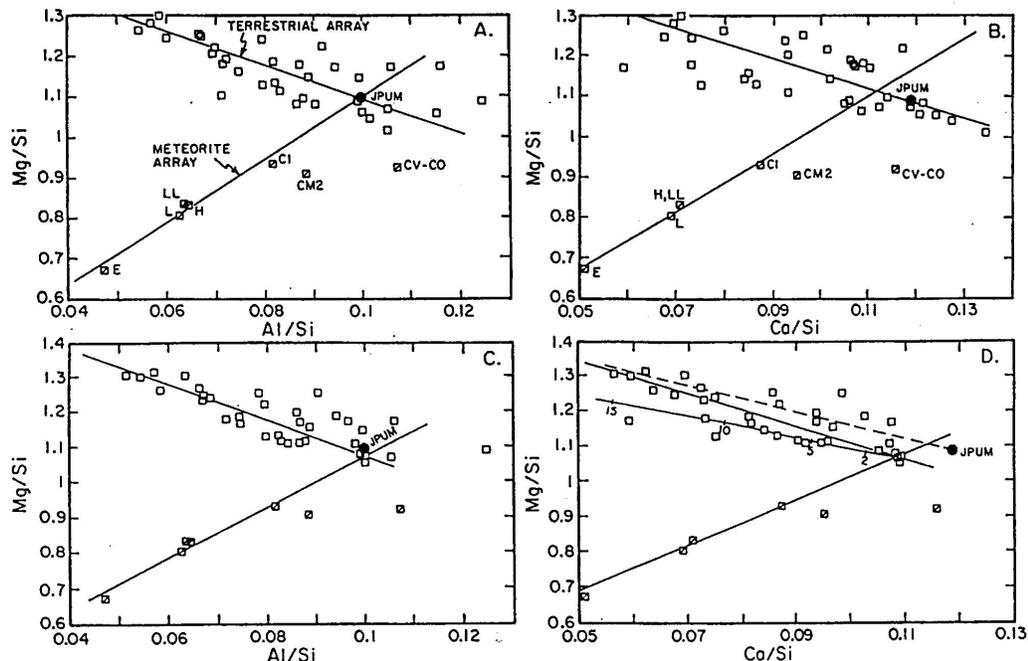


Figure 21 Mg/Si-Al/Si-Ca/Si relationships of meteorites and lherzolites. Panels A and B display data as published; panels C and D show lherzolites after subtraction of clinopyroxene (cpx) to force Ca/Al = 1.09 [=chondrites; see Hart & Zindler (1986) for complete discussion]. Solid circle labeled "JPUM" is the estimate of primitive mantle composition from Jagoutz et al (1979). Solid lines are eyeball fits to the data (forced to go through "JPUM" in A and B). Dashed line in D is the best-fit terrestrial line from B (included for reference). Solid line with the tick marks in D is trajectory of residual compositions resulting from melt removal; the percentage of melt removed is indicated by the numbers [see Hart and Zindler (1986) for description of this model]. CM2 and CV-CO chondrite data have been ignored in drawing the best-fit line to the meteorite array. Note that all concentration ratios are given as metal weight ratios.

Figure 2 (Zindler and Hart, 1986)

constraints from theories for nucleosynthesis of the elements during stellar evolution) as estimates for the abundances of these elements in the stuff which accreted to form the Earth. The primary difficulty in this exercise lies in determining what proportions of the various meteorite classes (iron, stony, chondrite, etc) should be brewed to make the Earth. For example, we know that practically all He either escaped from the Earth or never accreted in this higher-temperature region of the nebula. It is believed that water, sulfur, and carbon were all highly volatile and largely not accreted into the Earth, while Na, K, P, Cr, and Mn were accreted at less than nebular abundance ratios. Even Mg and Si, two of the primary elements in silicate Earth, are thought to have been slightly volatile during initial accretion (see notes at bottom of table 3)

Table 3 shows a comparison of the mantle chemistries (or Bulk Earth compositions) that are inferred from pyrolite and Losimig chemical abundances as well as meteorite and stellar-based approaches to constraining the composition of the Earth. Most geochemists typically show element abundances as weight-% of component materials, e.g. wt=% SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, FeO,...). This is largely done for historical reasons. Most geophysicists seem to prefer to visualize element abundances in terms of atomic abundances. Table 3 converts a typical pyrolite and Losimig mantle composition to numbers of atoms of each element per unit Si. These different estimates of mantle composition crudely agree. This agreement means that we can take the pyrolite model (determined from a cocktail of observed rocks at the surface) as our preferred composition.

The pyrolite and Losimig compositions, when expressed as a chemical formula, are quite similar:

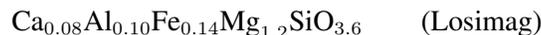
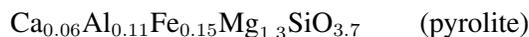


Table 3: Comparison of Inferred Abundances of Mantle Elements

Major Mantle Element	Terrestrial Mantle Abundance-Pyrolite model (atoms/total) (Col. 2 of Tab. 1)	Atoms/Si (Pyrolite) (Ringwood, 1976) (Col. 2 of Tab. 1)	Atoms/Si (LOSIMAG) (Hart&Zindler, 1986, Tab 2)	Atoms/Si (Solar) (Ross and Aller, 1976)	Atoms/Si (Nucl.+ Chondrite) (Cameron, 1973)
O (Oxygen)	0.583 (~58%)	3.713	3.63	15.5 C?V	21.5 C? V
Mg (Magnesium)	0.204 (~20%)	1.299	1.24	0.89	1.01
Si (Silicon)	0.157 (~16%)	1.0	1.0	1 C?	1 C?
Fe (Iron)	0.0240 (~2.4%)	0.153	0.138	0.71 C	0.83 C
Al (Aluminum)	0.0176 (~1.8%)	0.112	0.104	0.074	0.085
Ca (Calcium)	0.0094 (~1%)	0.0599	0.075	0.050	0.072
Na (Sodium)	0.0015	9.55E-3	0.02	0.043	0.060
Cr (Chromium)	0.0011	7.01E-3	9.04E-3	0.012	0.013
Ti (Titanium)	0.00063	4.01E-3	2.97E-3	2.5E-3	2.8E-3
Ni (Nickel)	0.00055	3.50E-3	4.85E-3	0.043 C	0.048 C
Mn (Manganese)	0.00029	1.85E-3	1.98E-3	5.9E-3	9.3E-3
K (Potassium)	0.000084	5.35E-4	8.95E-4	3.2E-3	4.2E-3
P (Phosphorous)	0.000059	3.76E-4 (376 ppm)	3.68E-4	7.1E-3	9.6E-3
S (Sulfur)				0.36 C? V	0.50 C? V

Notes: C -- possible major element in Earth's core V -- more volatile element during condensation process

A typical proposed condensation sequence for initial planetesimal accretion is:

1. Condensation of solid Os, Re, Zr at  $T >> 1680^\circ\text{K}$ , followed by  $\text{Al}_2\text{O}_3$  and then  $\text{CaTiO}_3$  and  $\text{Ca}_2\text{AlSiO}_7$  down to  $\sim 1500^\circ\text{K}$ .
2.  $\text{CaMgSi}_2\text{O}_6$  appears at  $1387^\circ\text{K}$ . Then Fe-Ni-Co alloys condense. At slightly cooler  $1370^\circ\text{K}$  temperatures,  $\text{Mg}_2\text{SiO}_4$  condenses. This later reacts with vapor to form  $\text{MgSiO}_3$  during which all remaining gaseous Si is consumed. This sequence implies that Fe-Ni alloys have a slightly higher condensation temperature than magnesium silicates, which can be used to argue for initial core condensation, then mantle condensation (heterogeneous accumulation process).
3. Below  $1250^\circ\text{K}$ , Cu, Ge, and Ga condense and form solid-solutions with preexisting condensed metal; Na, K, and Rb condense and enter into solid-solution with  $\text{CaAlSi}_2\text{O}_8$ . All Alkali metals condense by  $\sim 1000^\circ\text{K}$ .
4. Below  $750^\circ\text{K}$ , metallic Fe begins to oxidize. Pb, Bi, In, and Tl condense between  $600^\circ$  and  $400^\circ\text{K}$ , while at  $405^\circ\text{K}$  magnetite forms.

The amount of oxygen condensation into silicate earth is probably limited by how much is incorporated into silicates and iron oxides, while free oxygen remains a volatile phase. This is why steller oxygen concentrations are so large relative to the Earth? A good introductory review to nucleosynthesis is given in *Inorganic Geochemistry* (Henderson, Pergamon Press, 1986)

## 2. Mantle mineralogy

The major elements that make up the mantle are homogeneously distributed on a sub-centimeter scale. They form mineral crystals which are made of particular ratios of these elements incorporated within a regular, crystalline lattice. The fact that minerals are crystals is an important clue to determining the physical properties of the mantle. 99% of the elements that make up the mantle reside in only a handful of minerals. Thus, if we can determine the physical properties of these minerals with temperature and pressure, and how these minerals transform amongst themselves with temperature and pressure, then we have a fair shot at using laboratory measurements, with seismic observations, to constrain the mineralogy (and possible changes in chemistry) of the Earth's mantle.

The main minerals in ophiolites and kimberlite harzburgite and lherzolite assemblages are olivine, two types of pyroxene (on CA-rich), feldspar and garnets. Table 7 lists mineralogical definitions of the predominant surface outcrops of mantle rocks.

Table 7. Mineralogical Definitions of the more common outcrops & xenoliths of mantle rocks

Rock Type	Defining Mineralogy	Examples, Comments
Harzburgite ('oceanic' ophiolite)	Ol+Opx rich rock.	Samail Ophiolite Ol (~74%), Orthopyroxene (~24%), Spinel (2%) Trace cpx.
Dunite (frequent near top of 'oceanic' ophiolite)	Olivine rich rock.	Samail Ophiolite Ol (~97%), spinel (2%) + trace opx, cpx
Lherzolite (in Alpine Peridotites)	Ol (>~60%) + Opx + Cpx rich rock	Rhonda. Lherzolites often contain frequent cpx-rich veins of 'Gabbro' or Pyroxenite (>~60% pyroxene)
Eclogite. Component of Kimberlite xenoliths. (high pressure form of gabbro, plagioclase is replaced with garnet)	Ol + Cpx + Garnet	Usually has higher Fe/Mg ratios in Ol than do Harzburgite or Lherzolite. Suggests that this is a mixed-in prior melt (basalt).
Tholeiitic Basalt/Gabbro. Gabbro is the plutonic (i.e. intrusive, slow cooling) form of basalt.	Ol + Plag. + Cpx.	MORB chemistry. Prototype mid-ocean ridge basalt.
Wehrlite (often in igneous section of 'oceanic' ophiolites, not dredged from ocean floor)	Ol + Diopside (Cpx) + minor Plag. + trace hornblende	Samail Ophiolite. Wehrlite is mainly important because petrologists often use the fact that ocean dredges don't contain wehrlites to argue against the hypothesis that ophiolites are fragments of oceanic crust/lithosphere. I think this is a red herring!

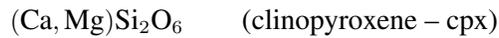
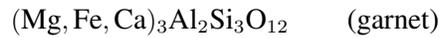
If we compare the pyroxene chemical formula for mantle composition with the compositions of Mg-rich olivine:



(the Mg end member is called forsterite) and orthopyroxene:

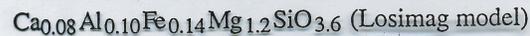
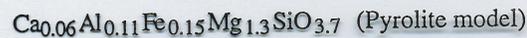


we see that the upper mantle has a predominantly olivine/pyroxene mineralogy. (The parentheses show that Mg and Fe are in solid solution with a roughly 9:1 Mg/Fe ratio.) An olivine/pyroxene mineralogy does not provide a framework into which to put Ca and Al. These elements are thought to reside in garnet, and clinopyroxene:



Omphacite and feldspar are two other (low pressure) minerals which can also accept Ca and Al as well as K and Na. Finally, in the upper mantle, spinels are often a trace non-silicate mineral. For example, magnetite is a spinel and is responsible for the magnetic stripes seen in the ocean floor. All the trace elements in the mantle tend to fit as impurities in garnets, cpx, feldspars, and spinels.

Pyrolite and Losimag compositions expressed as chemical formulae are quite similar:

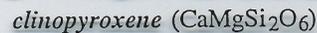
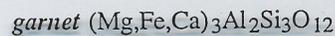


This can be primarily made from:



(Note that this mixture has  $\text{Mg/Si}=1.55$ , i.e. too high)

Ca and Al need to fit in other phases:



While the chemical composition of the mantle may be uniform, mantle minerals are stable only in certain pressure and temperature ranges. Changes in mineralogy may play an important role in mantle dynamics and evolution and are the cause of the property jumps seen in 1-D seismic models. The study of the pressure and temperature conditions under which a particular mineral assemblage is stable is the domain of thermodynamics which we consider in following lectures.