

Earth's mantle composition – present understanding

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The earth is chemically differentiated into the crust, mantle and core, of which the mantle makes up to 82% of the volume of the earth and 65% of its mass. Studies of samples brought from the mantle to the surface of the earth like mid-oceanic-ridge-basalt (MORB), oceanic-island-basalt (OIB), diamonds and xenoliths, by volcanic processes have indicated that the composition of the mantle, up to a depth of 150 km, is dominantly peridotitic, a rock composed of olivine and pyroxene.

A look at the seismic structure of the mantle points to major discontinuities at 410 km, 660 km and D'' layer above the core-mantle boundary which also show pronounced lateral heterogeneities. The shallower discontinuities at 410 km and 660 km can be related to phase transformations of olivine. The study of the Izu Bonin subduction zone show that the displacements of the discontinuities at 410 and 660 km are in good agreement with the predicted values if they are solely due to phase transformations. Study of this subducted zone also shows that subducted lithosphere penetrates the deeper mantle¹.

The heat budget of the Earth also indicates that the mantle is heterogeneous. To account for the entire global heat flux by radioactive decay of uranium, thorium and potassium and by cooling of the planet, it will be reasonable to assume that compared to the MORB source there is a lower layer enriched in such heat-producing elements within the mantle². Also the principal heat-producing elements produce noble gas isotopes through radioactive decay, making an intimate connection between heat and radiogenic ^4He and ^{40}Ar . Only 50% of the ^{40}Ar produced from the decay of ^{40}K over the age of the earth can be accounted for in the atmosphere and MORB source mantle³. The flux of ^4He , produced from decay of uranium and thorium is only 5% of the predicted value from ocean heat flow⁴.

When a mantle region melts, it loses incompatible trace elements such as uranium, thorium and potassium to the melt which is then concentrated in the oceanic or continental crust that forms. OIBs are

much more enriched in incompatible elements than the MORBs with a large variation in concentration, thus characterizing distinct enriched and depleted regions in the mantle for the two. Strontium and neodymium isotopes are often used to characterize the mantle into enriched and depleted regions. Since rubidium is less compatible than strontium, and samarium is more compatible than neodymium, in general $^{143}\text{Nd}/^{144}\text{Nd}$ correlates negatively with $^{87}\text{Sr}/^{86}\text{Sr}$ (^{87}Rb decays to ^{87}Sr and ^{147}Sm to ^{143}Nd). The MORBs which are depleted in incompatible elements have therefore lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios coupled with higher $^{143}\text{Nd}/^{144}\text{Nd}$ ratio. Continental crust which is enriched in the incompatible elements show just the reverse trend. In the array of strontium and neodymium isotopic data, the position of the OIB is between the depleted MORB source and the continental crust, suggesting that the OIBs could simply be the result of recycling of various types of continental material into the mantle. But often the isotopic variability of OIBs has been explained by mixing of different extreme mantle components⁵.

The continental crust and the depleted mantle are believed to be complementary in their chemical compositions of incompatible elements. The primitive composition of the bulk silicate earth (crust + mantle) can be derived from meteorites. Using simple mass balance relations, fraction of depleted mantle calculated using uranium, thorium and potassium turns out to be 50% of the whole mantle¹. However other estimate using $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ varies between 25 and 90% (refs 5, 6).

The three isotopes of noble gases ^3He , ^{36}Ar , ^{20}Ne and ^{22}Ne are all primordial while ^4He is produced from uranium and thorium decay, ^{40}Ar from decay of potassium and ^{21}Ne from (α, n) reactions on ^{18}O , where the α particles are produced by decay of U and Th. From Table 1 it is clear that the MORBs must have undergone a past episode of outgassing, thus reducing the $^3\text{He}/\text{U} + \text{Th}$ ratio and raising the $\text{K}/^{36}\text{Ar}$ and $\text{U} + \text{Th}/^{22}\text{Ne}$ ratios. Subsequent *in situ* production of ^{40}Ar , ^{21}Ne and ^4He modifies the isotopic ratios of

$^3\text{He}/^4\text{He}$, $^{21}\text{Ne}/^{22}\text{Ne}$ and $^{40}\text{Ar}/^{36}\text{Ar}$ to the present measured values. Also since the $^3\text{He}/^4\text{He}$ ratios of the MORBs show less scatter than OIBs, it is generally believed that the MORBs come from a degassed homogeneous upper mantle while the undegassed heterogeneous lower mantle is the source of the OIBs.

Although some geochemists do not believe in the concept of the mantle being layered into degassed and undegassed regions⁷, most of the noble gas experts still believe in a mantle that is partly degassed and partly pristine and undegassed.

The enriched layer should be able to account for the missing ^{40}Ar and ^4He . The D'' layer which is a layer of anomalous seismic velocities can only account for a very small fraction of the global heat flux unless there is extreme internal heat production. Very recently it has been proposed that the seismological heterogeneity observed at a depth of 1600 km is the manifestation of the boundary between the depleted MORB source and enriched dense material in the deeper mantle². A radically opposite view is put forward by some noble gas workers from their nitrogen and argon isotopic and elemental studies of OIBs, who propose that the enriched layer is at a shallower depth right below the lithosphere⁸.

There has been continued debate on whole mantle convection versus layered convection in the mantle. Support for whole mantle convection comes from geophysical evidence that indicates presence of deep high seismic velocity structures interpreted as subducted lithosphere in the lower mantle. But seismological evidences in favour of whole mantle convection and subduction of slabs into deep mantle all speak against a layered mantle where pristine mantle signatures

Table 1. Noble gas isotopic ratios for air, upper and lower mantle.

	MORBs	OIBs	AIR
$^3\text{He}/^4\text{He}$ (10^{-5})	1.2	4.6	0.12
$^{40}\text{Ar}/^{36}\text{Ar}$	> 40000	≈ 3000	295.5
$^{21}\text{Ne}/^{22}\text{Ne}$	0.06	0.04	0.029

are preserved in the deeper layer. Although the radioactive elements at present produce only 45% of the total heat flow, which lead to the common belief that there is an enriched layer deep within rich in radioactive elements, this does not take into account the fact that the mantle has a long thermal time constant, and advective removal of heat is related to the vigour of plate motions that vary over time. So the heat loss now must have been generated in the distant past. Taking this into account the present heat flux may be accounted for¹. Recently a laterally heterogeneous model of the earth has been proposed where the mantle is basically unstratified¹ (Figure 1). Subducting slabs introduce heterogeneities into the mantle, but they are disaggregated by the convective stirring of the mantle, resulting in isolated heterogeneities that scatter seismic energy but are too small to be observed tomographically. Also a portion of the subducted oceanic crust (excess oceanic crust) is stored at the base of the convecting system and thus sequesters incompatible elements temporarily.

Considering the present rate of subduction and assuming that this rate has been maintained throughout the geological past, about 45% 'sterile' mantle, that has undergone the process of partial melting and MORB extraction is likely to be present in the mantle⁹. However it is possible that a higher rate of heat production in the past, about 3.6 times the present value at 4 Ga, was compensated by higher heat loss through formation of

oceanic lithosphere, in which case the whole mantle must have been processed through MORB formation and presence of any undepleted mantle is unlikely¹. The lateral heterogeneous model of the mantle was put to test with lanthanum, zirconium and titanium. In the mass balance calculations, best overall agreement was achieved with 0.3% recycled continental material and 5% excess recycled oceanic crust. The depleted mantle constituted by sterile mantle and recycled material distributed as small km sized blobs would be 90% of the total mantle¹. MORBs come from the depleted mantle but the melting process of OIBs must involve more the recycled oceanic and continental crust to generate its highly incompatible trace element content.

Both continental crust and depleted mantle have subchondritic Nb/Ta, Nb/La and Ti/Zr ratios. Recently Archean eclogites have shown to have the complementary superchondritic ratios¹⁰, and since the lithosphere have lower eclogites than that required to explain the mass imbalance, it is possible that much of it resides in the deeper mantle to form the excess recycled oceanic crust as predicted by the lateral heterogeneous model of the earth.

The above model can successfully explain the Sm-Nd and Rb-Sr signatures of the MORBs and OIBs. Also some ratios like Nb/U \approx 47 and Ta/U \approx 2.7 are similar both in MORBs and OIBs, but the values are higher than that of continental crust and that estimated for pristine mantle¹¹. This is a strong indication

that OIBs do not sample exclusive primitive mantle reservoirs and the origins of MORBs and OIBs may be similar, a strong point in favour of lateral heterogeneous model of the mantle. But this model fails to account for the enriched $^3\text{He}/^4\text{He}$ ratios of the OIBs.

Enrichment of ^3He in OIBs is possible if helium is more compatible in mantle crystals than uranium and thorium. Very old subducted mantle, like the Archean mafic crust that was recycled back into the mantle¹² (excess oceanic crust) and has undergone partial melting in the distant past may have high $^3\text{He}/^4\text{He}$ ratio.

Another possibility is if helium is present in the core of the earth¹³. To quantify precisely the helium that may be present in the core, it is important to consider whether silicates were solid or molten during core segregation and the relevant partition coefficient during this separation. Also the quantity of helium available during partitioning becomes vital. There is evidence for substantial rare gas losses up to 100 Ma after the start of the solar system at 4.56 Ga, implying higher initial concentration of noble gases¹⁴. Such quantities of solar gases may be acquired from the atmosphere by the mantle by dissolution from a hot dense atmosphere of solar composition following simple Henry's law. Indeed recently, neon isotopes clearly demonstrate the presence of solar rare gases in the mantle¹⁵. Even with low partition coefficients between silicates and liquid Fe, there might have been substantial transfer of helium and neon to the core¹³. Subsequent transfer of helium and neon from the core to the mantle may be simply by entrainment of core material¹³. Different contributions of core material, and hence core helium to the source regions may explain the range of values of $^3\text{He}/^4\text{He}$ obtained for different OIBs and presently explained by variable degassing.

Although it cannot at this point be said with certainty, core might be a source of helium, in which case the lateral heterogeneous model of the mantle will replace the conventional stratified mantle model. However it will be difficult to ascertain core contribution to mantle by using a single tracer. A better approach would be to use multiple tracers to confirm core contribution.

If the earth's mantle was in equilibrium with the core, the mantle content of platinum-group-elements (PGE) would

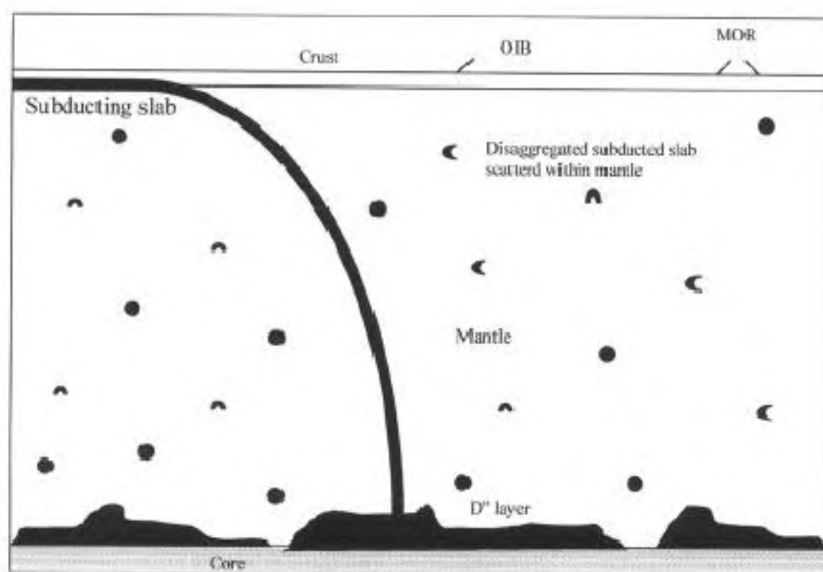


Figure 1. Laterally heterogeneous, unstratified mantle.

be less by several orders than what is observed. Recently it has been shown that differentiated outer core material back into the mantle could well account for the observed noble metal ratios and abundance in the mantle without any late accretionary veneer¹⁶. Studies of iron meteorites have revealed that crystallization of inner core would enhance the Re/Os and Pt/Os ratios of the liquid outer core. So contribution of outer core material to plumes should show a coupled enrichment in ¹⁸⁷Os/¹⁸⁸Os and ¹⁸⁶Os/¹⁸⁸Os as ¹⁸⁷Re and ¹⁹⁰Pt decay respectively to ¹⁸⁷Os and ¹⁸⁶Os (ref. 17). In fact such simultaneous enhancement has been reported from Hawaiian plume derived lavas¹⁷. Addition of recycled crustal material can enhance Re/Os ratio in the mantle leading to ¹⁸⁷Os enrichment, but the coupled enrichment cannot result from such recycling.

A correlation between ³He/⁴He and ¹⁸⁶Os/¹⁸⁸Os, as already reported from Hawaiian volcanoes¹⁸ will be a sure indication of core contribution in the samples but may be difficult to assess owing to the high mobility of He.

Future experimental works on partition coefficients between silicate-metal of

³He and ⁴He may prove useful if they attain a distinct ratio in the core and thus can be useful to identify any core contribution to the mantle.

1. Helffrich, G. R. and Wood, B. J., *Nature*, 2001, **412**, 501–507.
2. Kellogg, L. H., Hager, B. H. and Vander Hilst, R. D., *Science*, 1999, **283**, 1881–1884.
3. Allegre, C. J., Hofman, A. W. and O’Nions, R. K., *Geophys. Res. Lett.*, 1996, **23**, 3555–3557.
4. O’Nions, R. K. and Oxburgh, R., *Nature*, 1983, **306**, 429–431.
5. Hofmann, A. W., *Nature*, 1997, **385**, 219–229.
6. Jacobsen, S. B. and Wasserburg, G. J., *J. Geophys. Res.*, 1979, **84**, 7411–7427.
7. Anderson, D. L., *Earth Planet. Sci. Lett.*, 2001, **193**, 77–82.
8. Mohapatra, R. K. and Murty, S. V. S., *Geophys. Res. Lett.*, 2000, **27**, 1583–1586.
9. Plank, T. and Langmuir, C. H., *Chem. Geol.*, 1998, **145**, 325–394.
10. Rudnick, R. L., Barth, M., Horn, I. and McDonough, W. F., *Science*, 2000, **287**, 278–281.
11. Hofmann, A. W., Iochum, K. P., Seufert, M. and White, W. M., *Earth Planet. Sci. Lett.*, 1986, **79**, 33–45.

12. Vervoort, J. D., Patchett, P. J., Cehrels, G. E. and Nutman, A. P., *Nature*, 1996, **379**, 624–627.
13. Porcelli, D. and Halliday, A. N., *Earth Planet. Sci. Lett.*, 2001, **192**, 45–56.
14. Porcelli, D., Woolum, D. and Cassen, P., *ibid*, 2001, **193**, 237–251.
15. Honda, M., McDougall, I., Patterson, D. B., Dougeris, A. and Clague, D. A., *Nature*, 1991, **349**, 149–151.
16. Snow, J. E. and Schmidt, G., *ibid*, 1998, **391**, 166–169.
17. Brandon, A. D., Walker, R. J., Morgan, J. W., Norman, M. D. and Prichard, H. M., *Science*, 1998, **280**, 1570–1573.
18. Brandon, A. D., Norman, M. D., Walker, R. J. and Morgan, J. W., *Earth Planet. Sci. Lett.*, 1999, **174**, 25–42.

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COMMENTARY

Molecular markers and QTL analysis in crop plants

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Majority of quantitative traits in crop plants are controlled by polygenes; most of them have minor effects and only occasionally some of the genes have major effects¹. These gene loci are described as quantitative trait loci (QTL) and can be detected with the help of molecular markers, which should also segregate in a Mendelian manner. A QTL actually describes a region of the chromosome defined either by linkage to an individual molecular marker or by two flanking markers which may or may not be linked with the QTL². Molecular markers that are used for QTL analysis largely refer to DNA-based markers, which can be used in plant breeding for

various purposes. For instance, marker-trait association has been successfully used for indirect marker-assisted selection (MAS) in cases like soybean cyst nematode (SCN) resistance and rice bacterial blight resistance^{3,4}. A variety of molecular marker systems are now available for use in QTL analysis, and several laboratories in India are involved in work on molecular markers for crop improvement⁵. Elsewhere, we have discussed in some detail the principles, methodology and relative merits of individual marker systems^{6–9}.

A variety of methods are available for QTL analysis in crop plants^{10,11}. The earliest and simplest methods for QTL

analysis included single-marker regression or tests for independence in segregation of molecular marker and the trait of interest. A serious limitation of this approach is confounding of the effect of one QTL by many others that influence the trait. Another serious limitation is that a QTL with major effect and loose linkage cannot be distinguished from a QTL with minor effect and tight linkage. A milestone in QTL analysis, therefore, was the QTL interval mapping¹², which had several advantages over the above traditional methods involving regression analysis or tests of independence. This approach needs a molecular genetic map and utilizes, in a single analysis, infor-