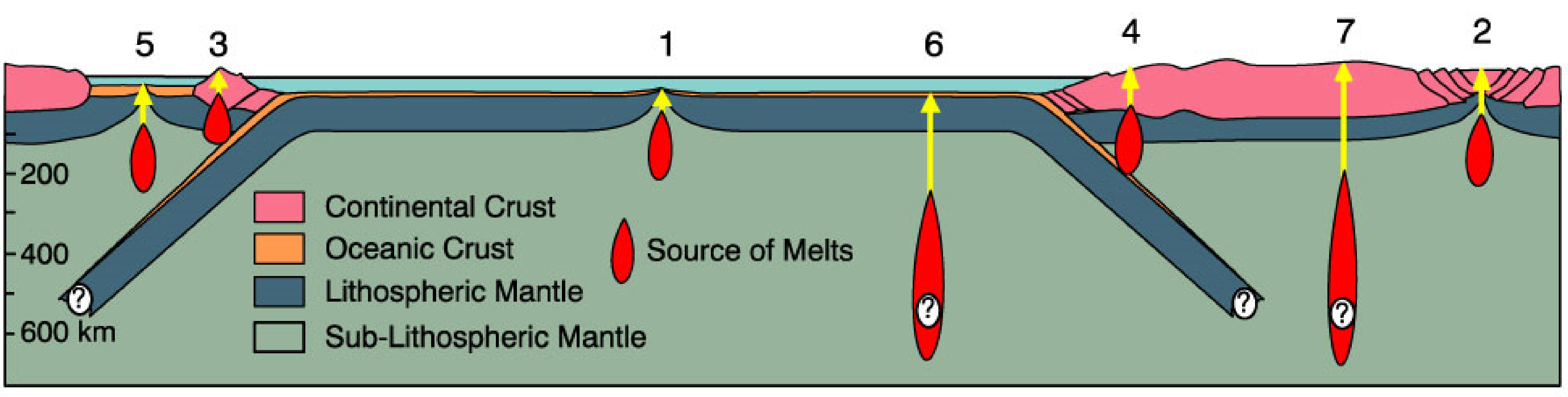
# Tectonic-Igneous Associations

- Associations on a larger scale than the petrogenetic provinces
- An attempt to address global patterns of igneous activity by grouping provinces based upon similarities in occurrence and genesis



# Tectonic-Igneous Associations

- Mid-Ocean Ridge Volcanism
- Ocean Intra-plate (Island) volcanism
- Continental Plateau Basalts
- Subduction-related volcanism and plutonism
  - Island Arcs
  - Continental Arcs
- Granites (not a true T-I Association)
- Mostly alkaline igneous processes of stable craton interiors
- Anorthosite Massifs

# Chapter 13: Mid-Ocean Rifts

## The Mid-Ocean Ridge System

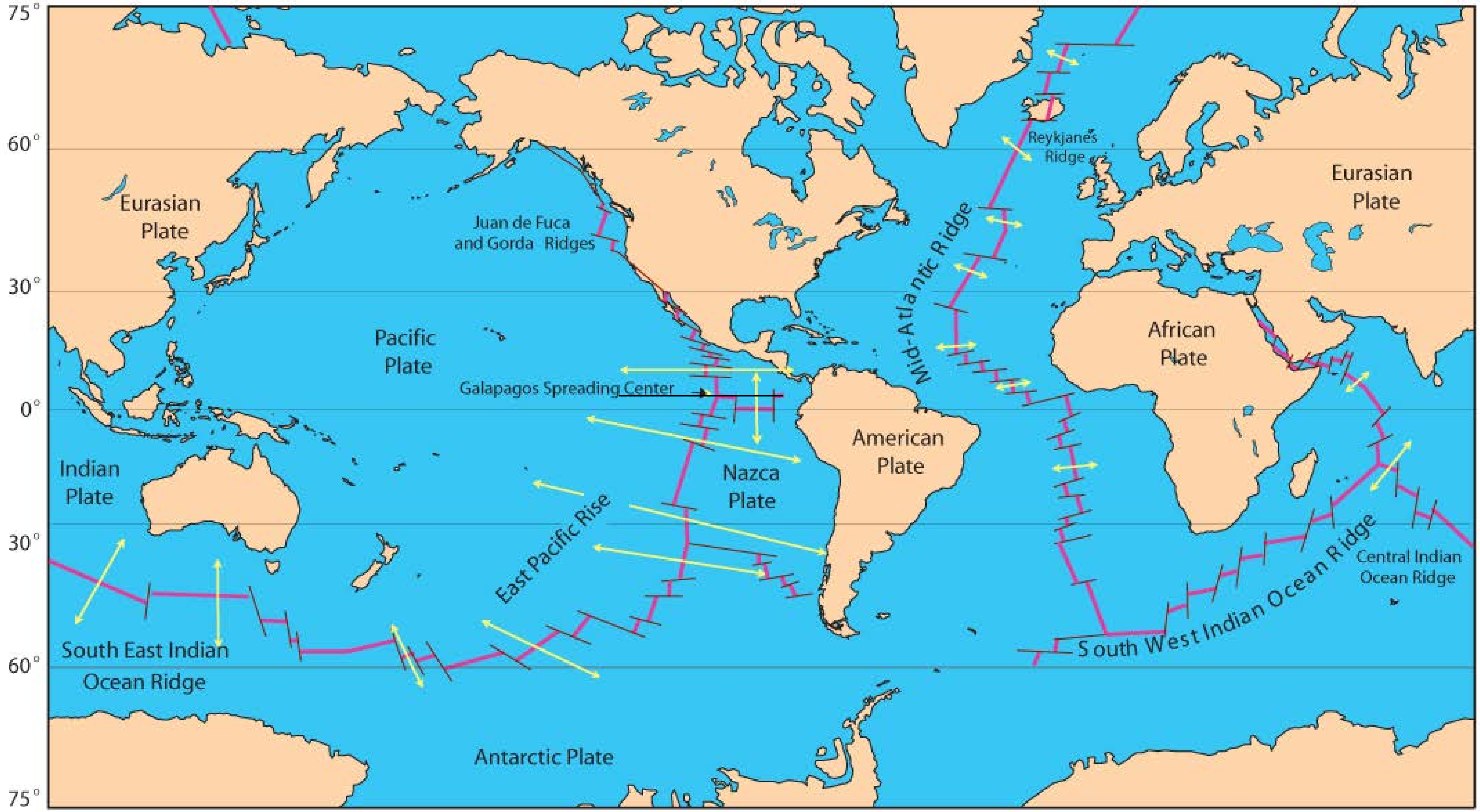



**Figure 13.1. After Minster et al. (1974) Geophys. J. Roy. Astr. Soc., 36, 541-576.**

# Ridge Segments and Spreading Rates

- Slow-spreading ridges:
  < 3 cm/a
- Fast-spreading ridges:
  > 4 cm/a are considered
- Temporal variations are also known

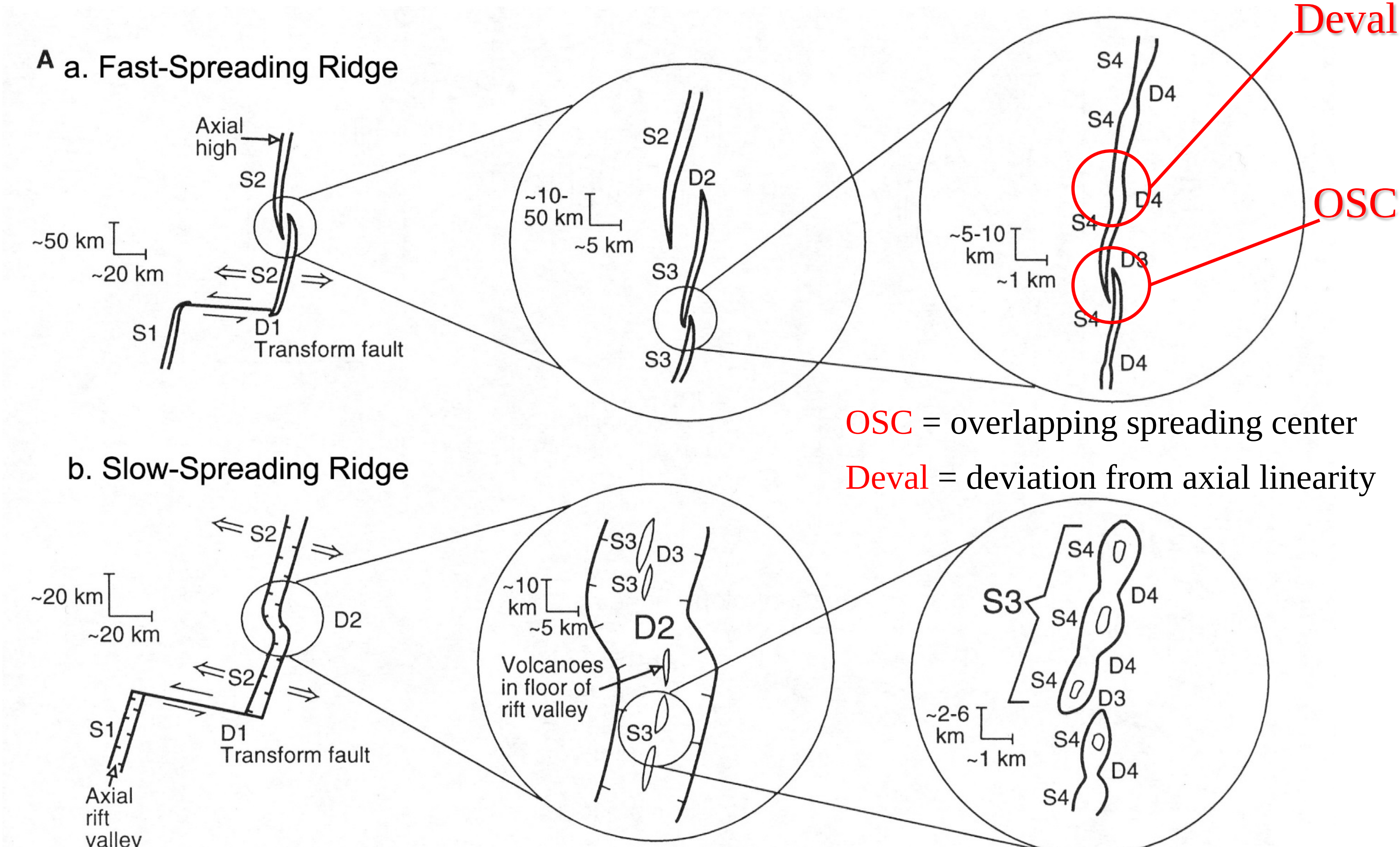
**Table 13-1.** Spreading Rates of Some Mid-Ocean Ridge Segments

| Category | Ridge | Latitude | Rate (cm/a)* |
|---|---|---|---|
| Fast | East Pacific Rise | 21-23°N | 3 |
| | | 13°N | 5.3 |
| | | 11°N | 5.6 |
| | | 8-9°N | 6 |
| | | 2°N | 6.3 |
| | | 20-21°S | 8 |
| | | 33°S | 5.5 |
| | | 54°S | 4 |
| | | 56°S | 4.6 |
| Slow | Indian Ocean | SW | 1 |
| | | SE | 3-3.7 |
| | | Central | 0.9 |
| | Mid-Atlantic Ridge | 85°N | 0.6 |
| | | 45°N | 1-3 |
| | | 36°N | 2.2 |
| | | 23°N | 1.3 |
| | | 48°S | 1.8 |

From Wilson (1989). Data from Hekinian (1982), Sclater *et al*. (1976), Jackson and Reid (1983). *half spreading

# Ridge Segments and Spreading Rates

## Hierarchy of ridge segmentation



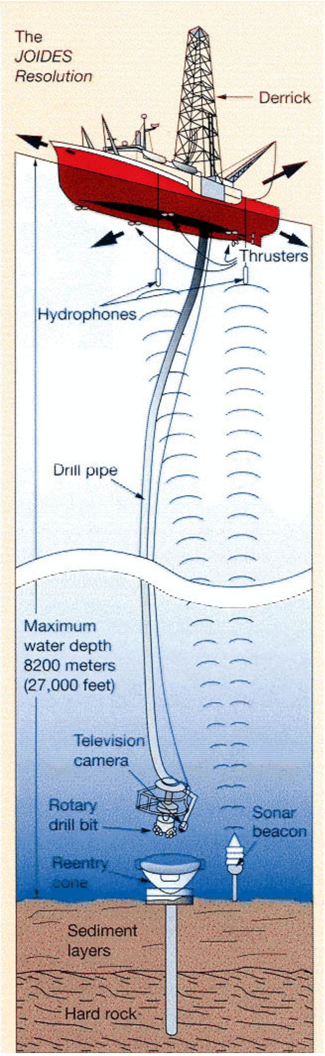

OSC = overlapping spreading center

Deval = deviation from axial linearity

**Figure 13.3.** S1-S4 refer to ridge segments of first- to fourth-order and D1-D4 refer to discontinuities between corresponding segments. After Macdonald (1998).

# Oceanic Crust and Upper Mantle Structure

- 4 layers distinguished via seismic vel
- Deep S                    n
- Dredg                    scarps
- Ophio



# *Oceanic Crust and Upper Mantle Structure*

## Typical Ophiolite

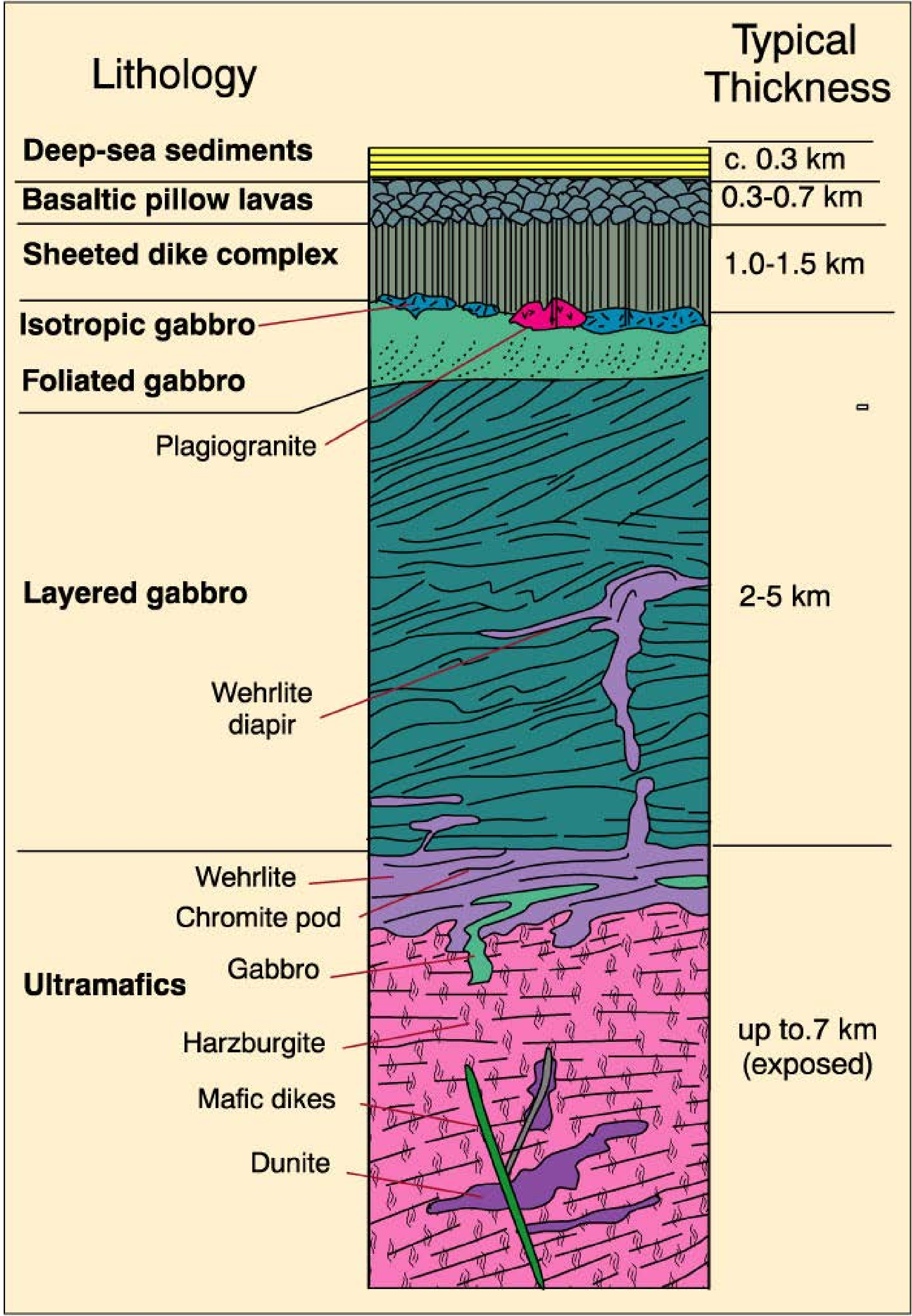



**Figure 13.4.** **Lithology and thickness of a typical ophiolite sequence, based on the Samial Ophiolite in Oman. After Boudier and Nicolas (1985) Earth Planet. Sci. Lett., 76, 84-92.**

# Oceanic Crust and Upper Mantle Structure

## *Layer 1*

A thin layer of pelagic sediment

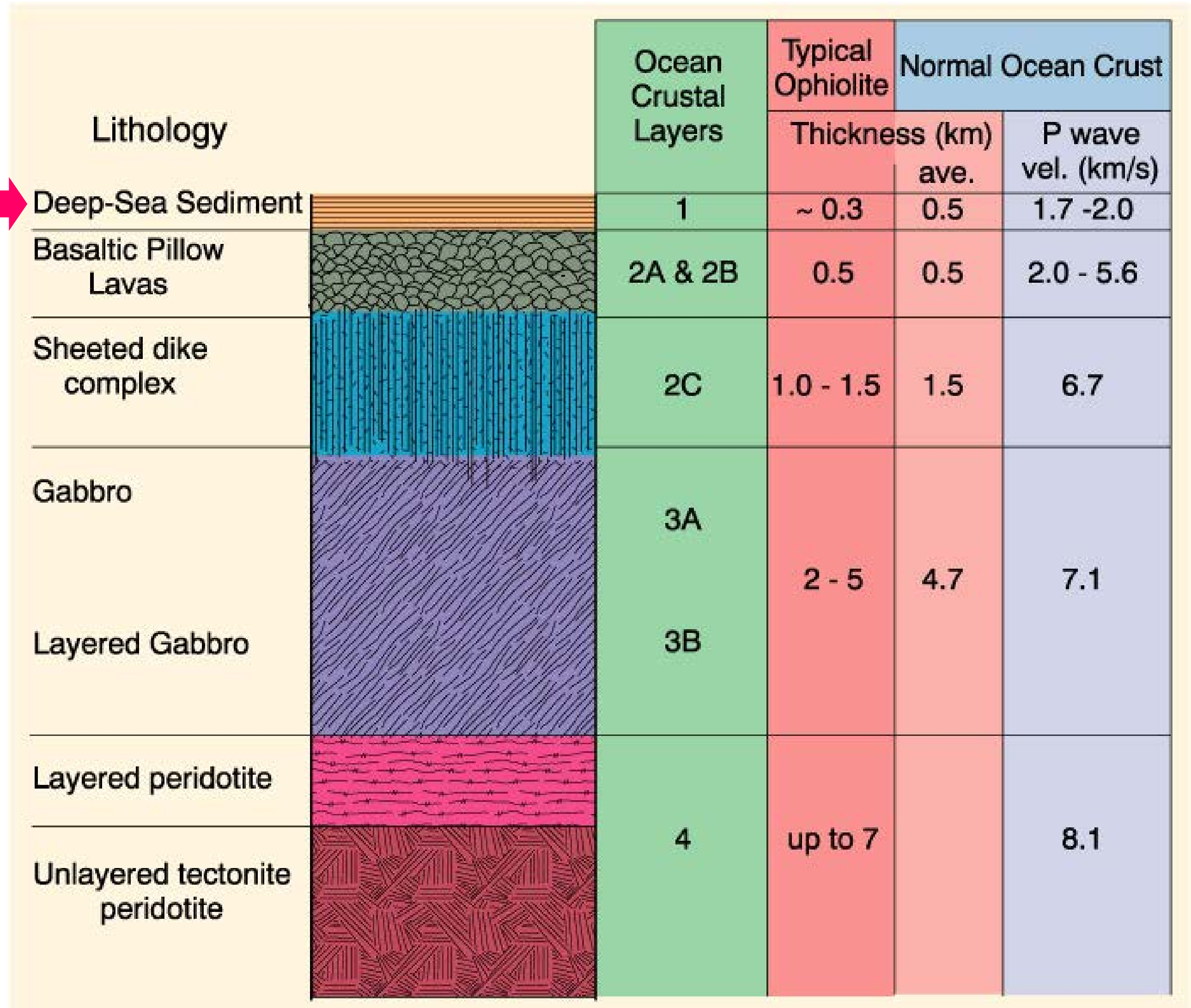



**Figure 13.5.** **Modified after Brown and Mussett (1993) The Inaccessible Earth: An Integrated View of Its Structure and Composition. Chapman & Hall. London.**

# Oceanic Crust and Upper Mantle Structure

## *Layer 2* is basaltic

Subdivided into two sub-layers

Layer 2A & B = pillow basalts

Layer 2C = vertical sheeted dikes

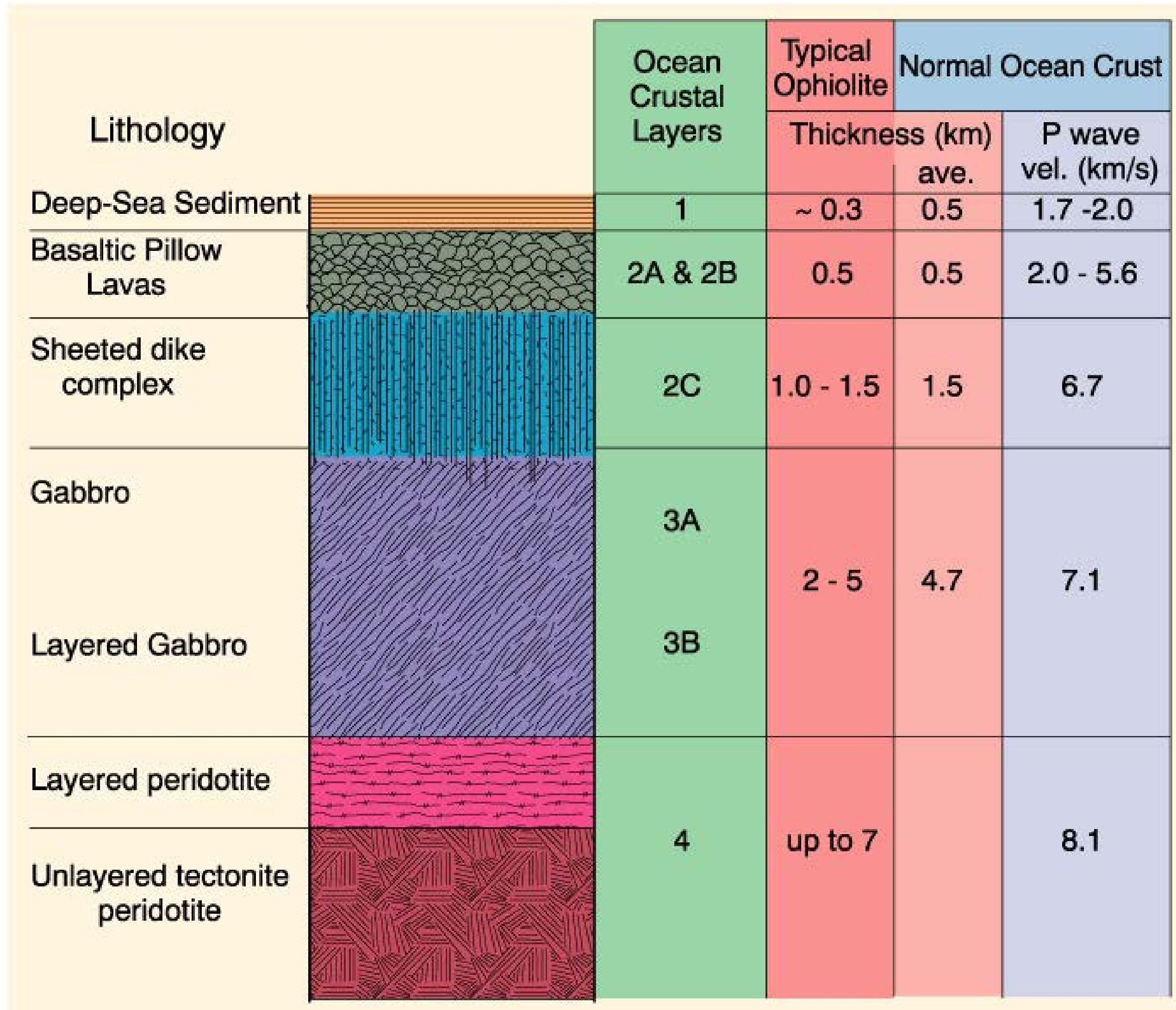


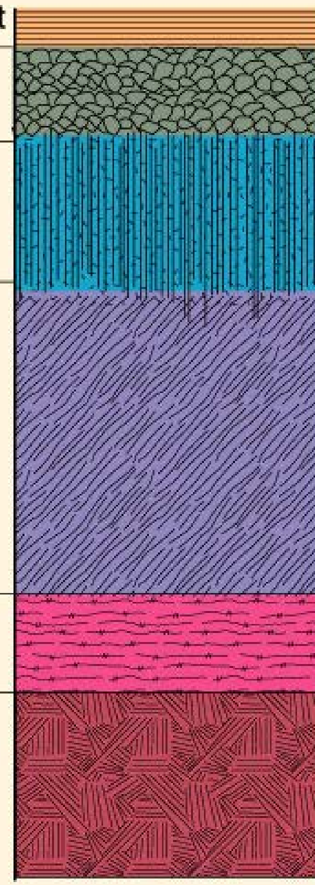

**Figure 13.5.** **Modified after Brown and Mussett (1993) The Inaccessible Earth: An Integrated View of Its Structure and Composition. Chapman & Hall. London.**

***Layer 3*** more complex and controversial

Believed to be mostly gabbros, crystallized from a shallow axial magma chamber (feeds the dikes and basalts)

Layer 3A = upper isotropic and lower, somewhat foliated ("transitional") gabbros

Layer 3B is more layered, & may exhibit cumulate textures



| Lithology | Ocean Crustal Layers | Typical Ophiolite Thickness (km) | Normal Ocean Crust Thickness (km) ave. | Normal Ocean Crust P wave vel. (km/s) |
|---|---|---|---|---|
| Deep-Sea Sediment | 1 | ~ 0.3 | 0.5 | 1.7 -2.0 |
| Basaltic Pillow Lavas | 2A & 2B | 0.5 | 0.5 | 2.0 - 5.6 |
| Sheeted dike complex | 2C | 1.0 - 1.5 | 1.5 | 6.7 |
| Gabbro | 3A | 2 - 5 | 4.7 | 7.1 |
| Layered Gabbro | 3B | | | |
| Layered peridotite | 4 | up to 7 | | 8.1 |
| Unlayered tectonite peridotite | | | | |

# Oceanic Crust and Upper Mantle Structure

Discontinuous diorite and tonalite ("plagiogranite") bodies = late differentiated liquids

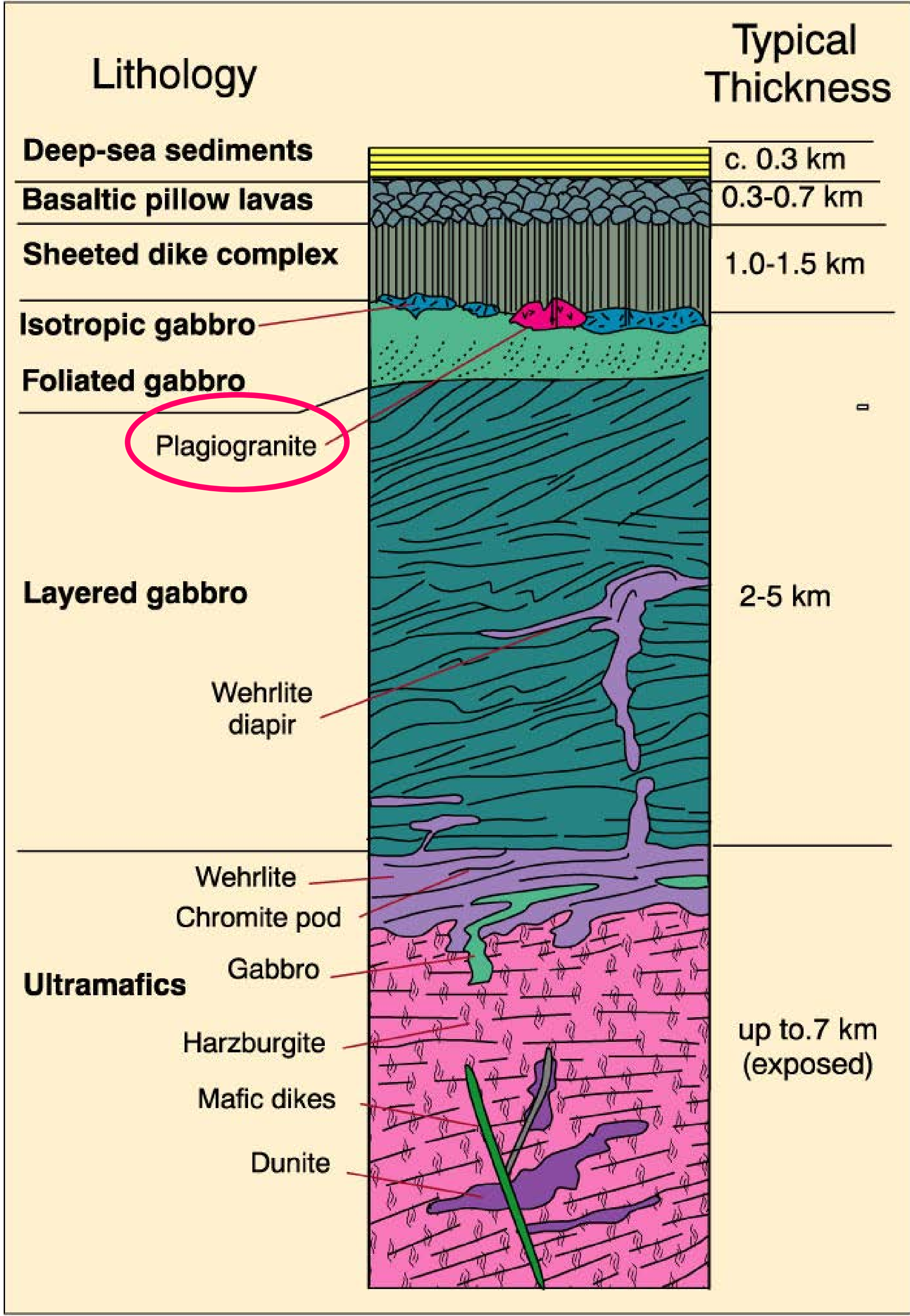
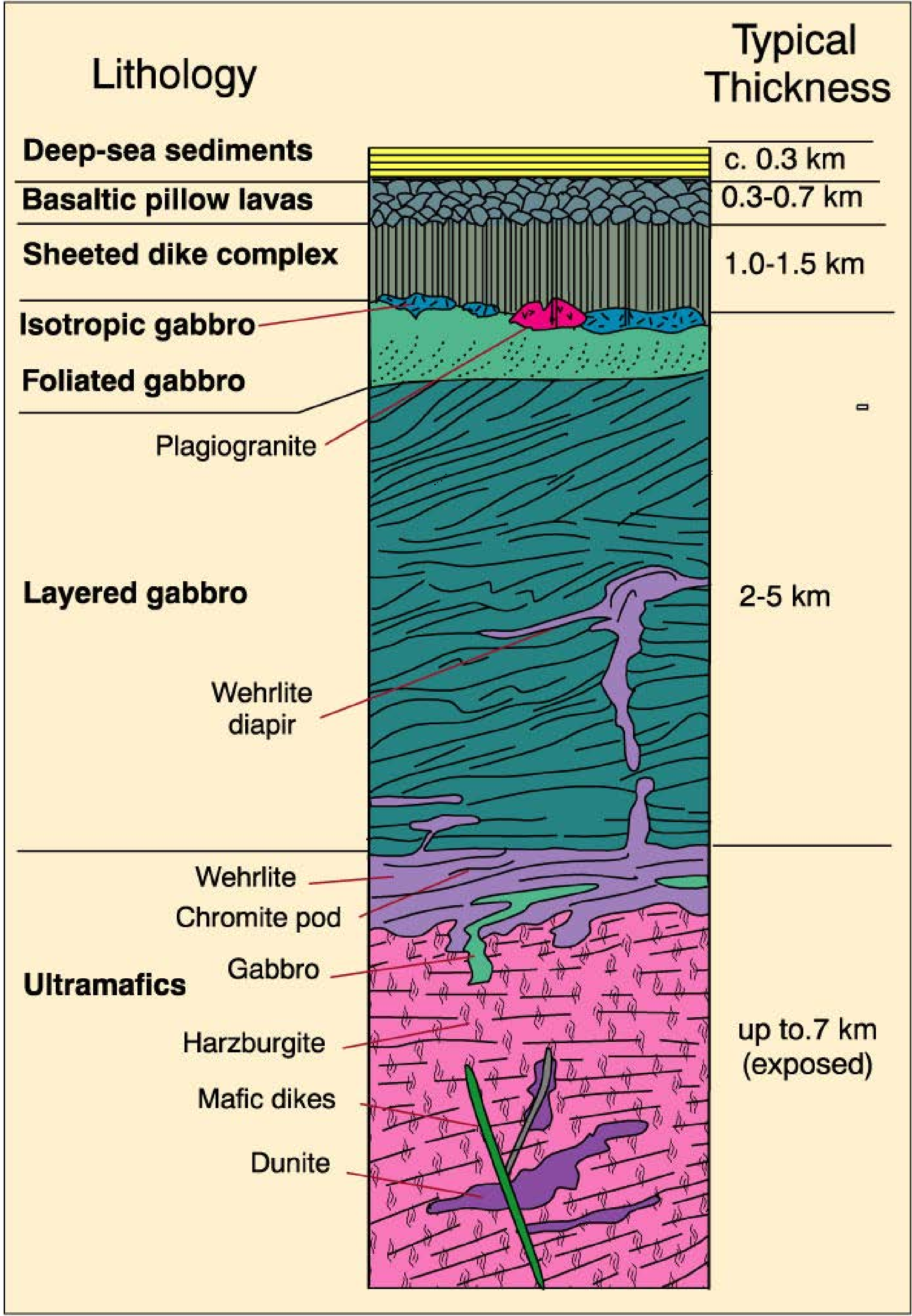


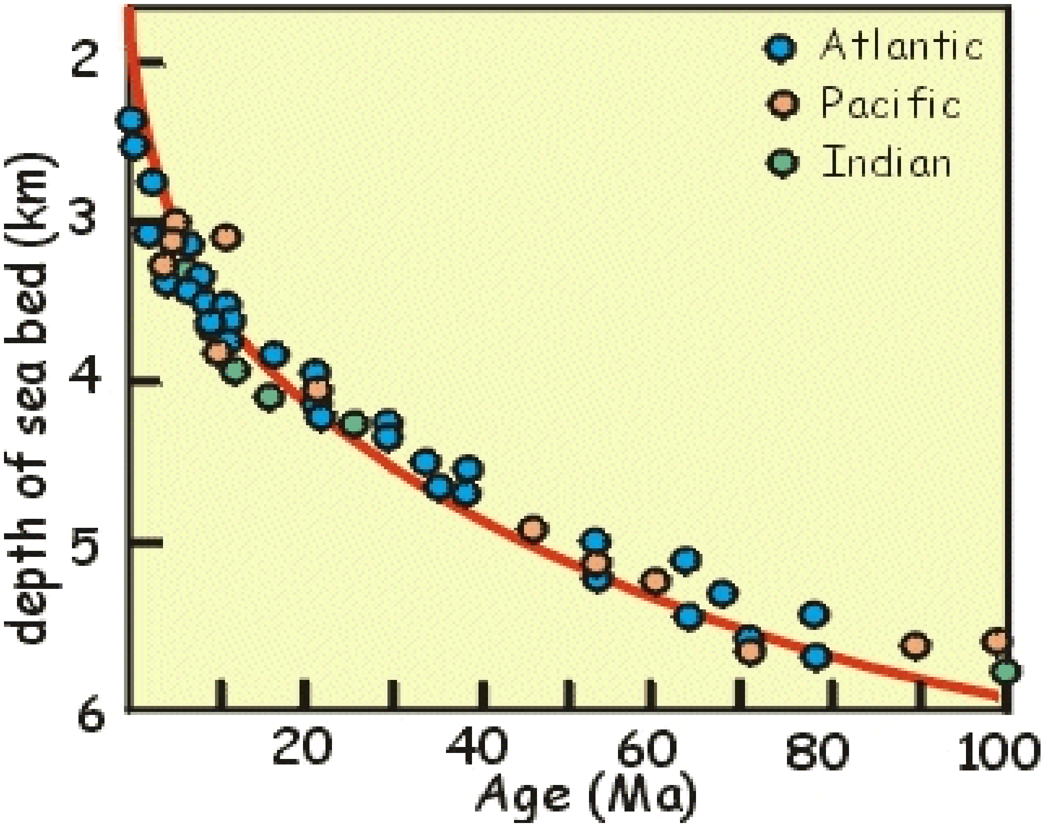

**Figure 13.4.** **Lithology and thickness of a typical ophiolite sequence, based on the Samial Ophiolite in Oman. After Boudier and Nicolas (1985) Earth Planet. Sci. Lett., 76, 84-92.**

# Layer 4 = ultramafic rocks

- Ophiolites: base of 3B grades into layered cumulate wehrlite & gabbro
- Wehrlite intruded into layered gabbros
- Below → cumulate dunite with harzburgite xenoliths
- Below this is a tectonite harzburgite and dunite (unmelted residuum of the original mantle)



# Elevation of ridge reduces with time as plate cools



# Petrography and Major Element Chemistry

A "typical" MORB is an olivine tholeiite with low $K_2O$ (< 0.2%) and low $TiO_2$ (< 2.0%)

Only glass is certain to represent *liquid* compositions

The common crystallization sequence is: olivine (± Mg-Cr spinel), olivine + plagioclase (± Mg-Cr spinel), olivine + plagioclase + clinopyroxene

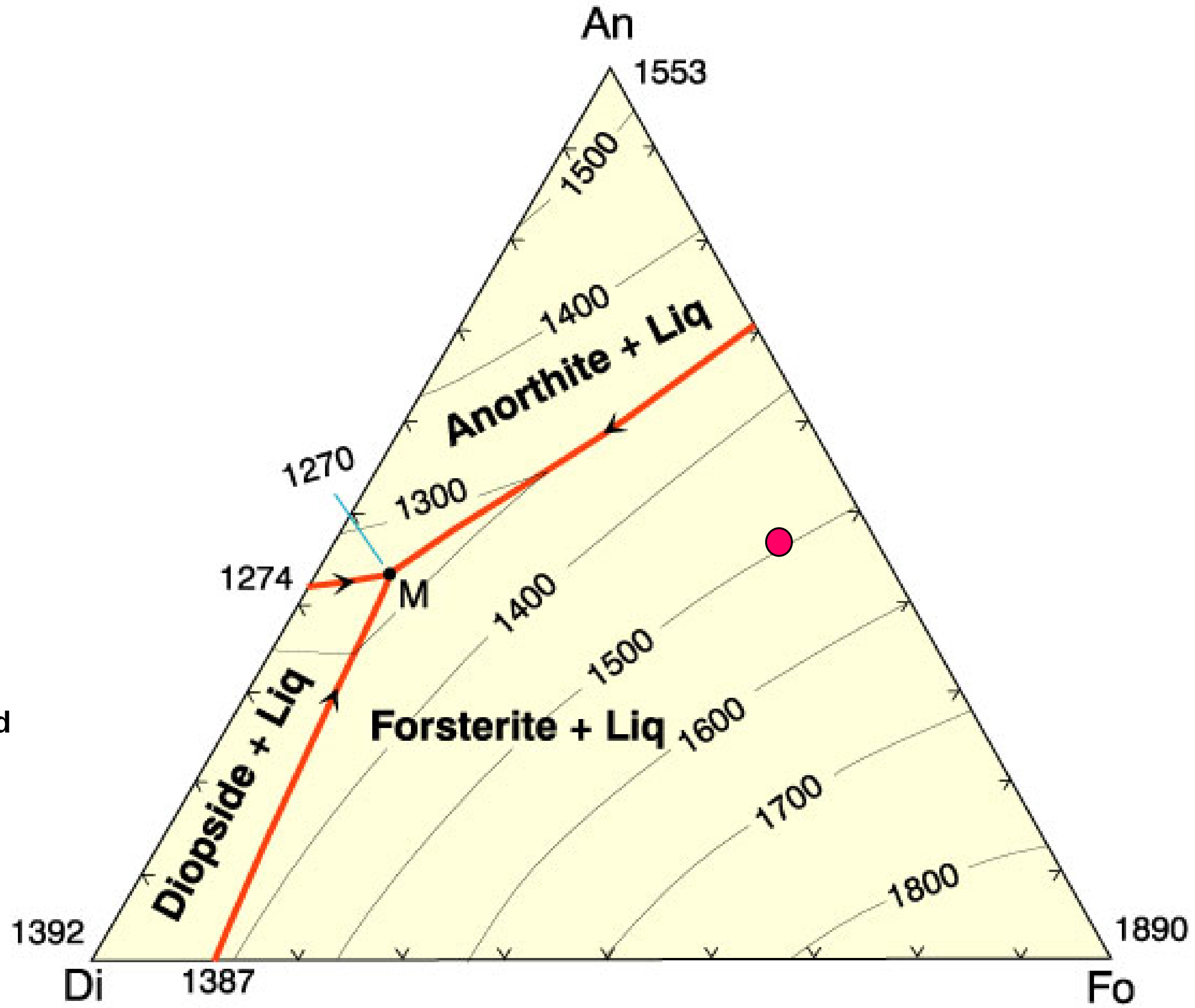



**Figure 7.2.** **After Bowen (1915), A. J. Sci., and Morse (1994), Basalts and Phase Diagrams. Krieger Publishers.**

- Fe-Ti oxides are restricted to the groundmass, and thus form late in the MORB sequence

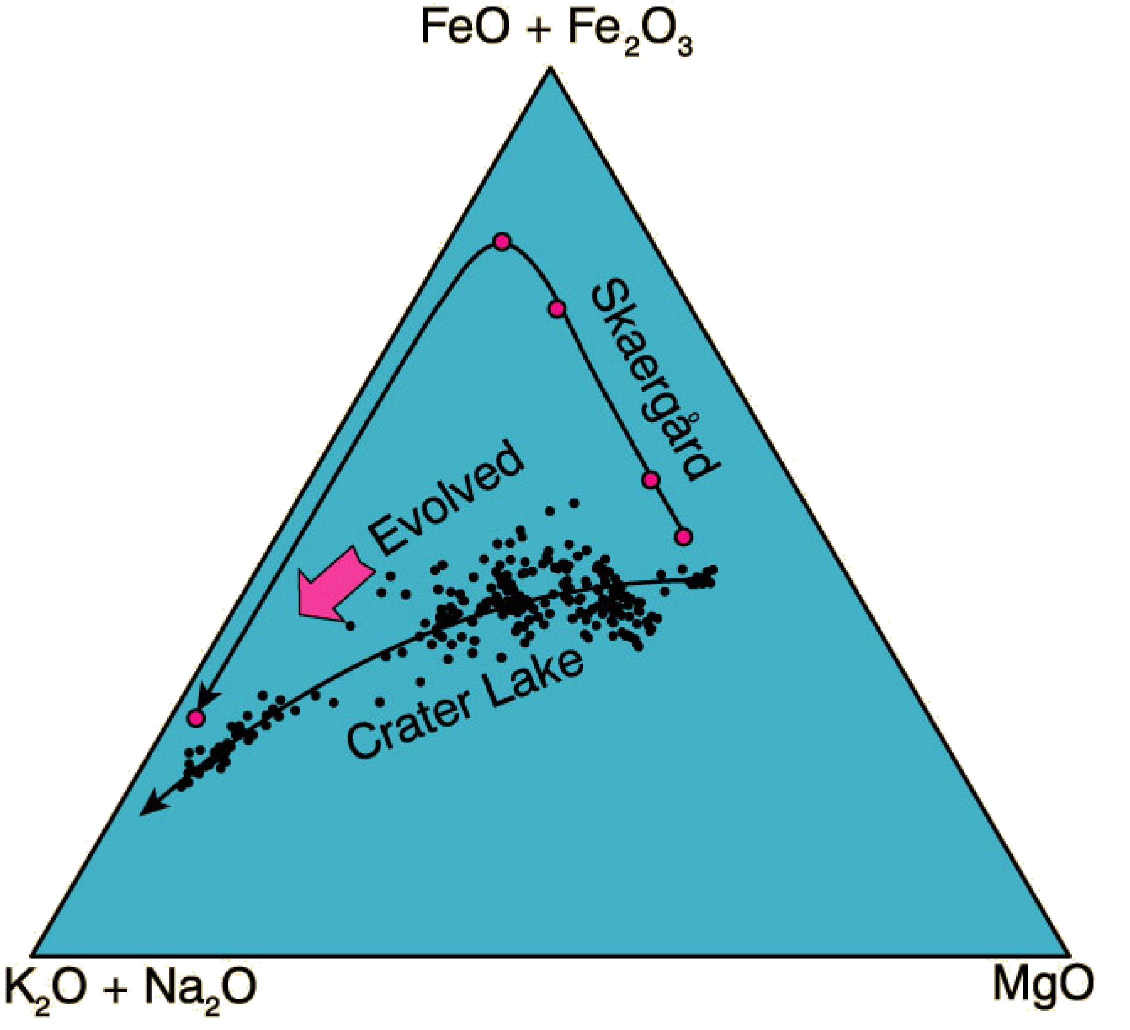



**Figure 8.2.** **AFM diagram for Crater Lake volcanics, Oregon Cascades. Data compiled by Rick Conrey (personal communication).**

# The major element chemistry of MORBs

Originally considered to be extremely uniform, interpreted as a simple petrogenesis

More extensive sampling has shown that they display a (restricted) range of compositions

# The major element chemistry of MORBs

**Table 13-2.** Average Analyses and CIPW Norms of MORBs (BVTP Table 1.2.5.2)

| Oxide (wt%) | All | MAR | EPR | IOR |
|---|---|---|---|---|
| $SiO_2$ | 50.5 | 50.7 | 50.2 | 50.9 |
| $TiO_2$ | 1.56 | 1.49 | 1.77 | 1.19 |
| $Al_2O_3$ | 15.3 | 15.6 | 14.9 | 15.2 |
| FeO* | 10.5 | 9.85 | 11.3 | 10.3 |
| MgO | 7.47 | 7.69 | 7.10 | 7.69 |
| CaO | 11.5 | 11.4 | 11.4 | 11.8 |
| $Na_2O$ | 2.62 | 2.66 | 2.66 | 2.32 |
| $K_2O$ | 0.16 | 0.17 | 0.16 | 0.14 |
| $P_2O_5$ | 0.13 | 0.12 | 0.14 | 0.10 |
| Total | 99.74 | 99.68 | 99.63 | 99.64 |
| | | | | |
| Norm | | | | |
| q | 0.94 | 0.76 | 0.93 | 1.60 |
| or | 0.95 | 1.0 | 0.95 | 0.83 |
| ab | 22.17 | 22.51 | 22.51 | 19.64 |
| an | 29.44 | 30.13 | 28.14 | 30.53 |
| di | 21.62 | 20.84 | 22.5 | 22.38 |
| hy | 17.19 | 17.32 | 16.53 | 18.62 |
| ol | 0.0 | 0.0 | 0.0 | 0.0 |
| mt | 4.44 | 4.34 | 4.74 | 3.90 |
| il | 2.96 | 2.83 | 3.36 | 2.26 |
| ap | 0.30 | 0.28 | 0.32 | 0.23 |

All: Ave of glasses from Atlantic, Pacific and Indian Ocean ridges.

MAR: Ave. of MAR glasses. EPR: Ave. of EPR glasses.

IOR: Ave. of Indian Ocean ridge glasses.

# The major element chemistry of MORBs

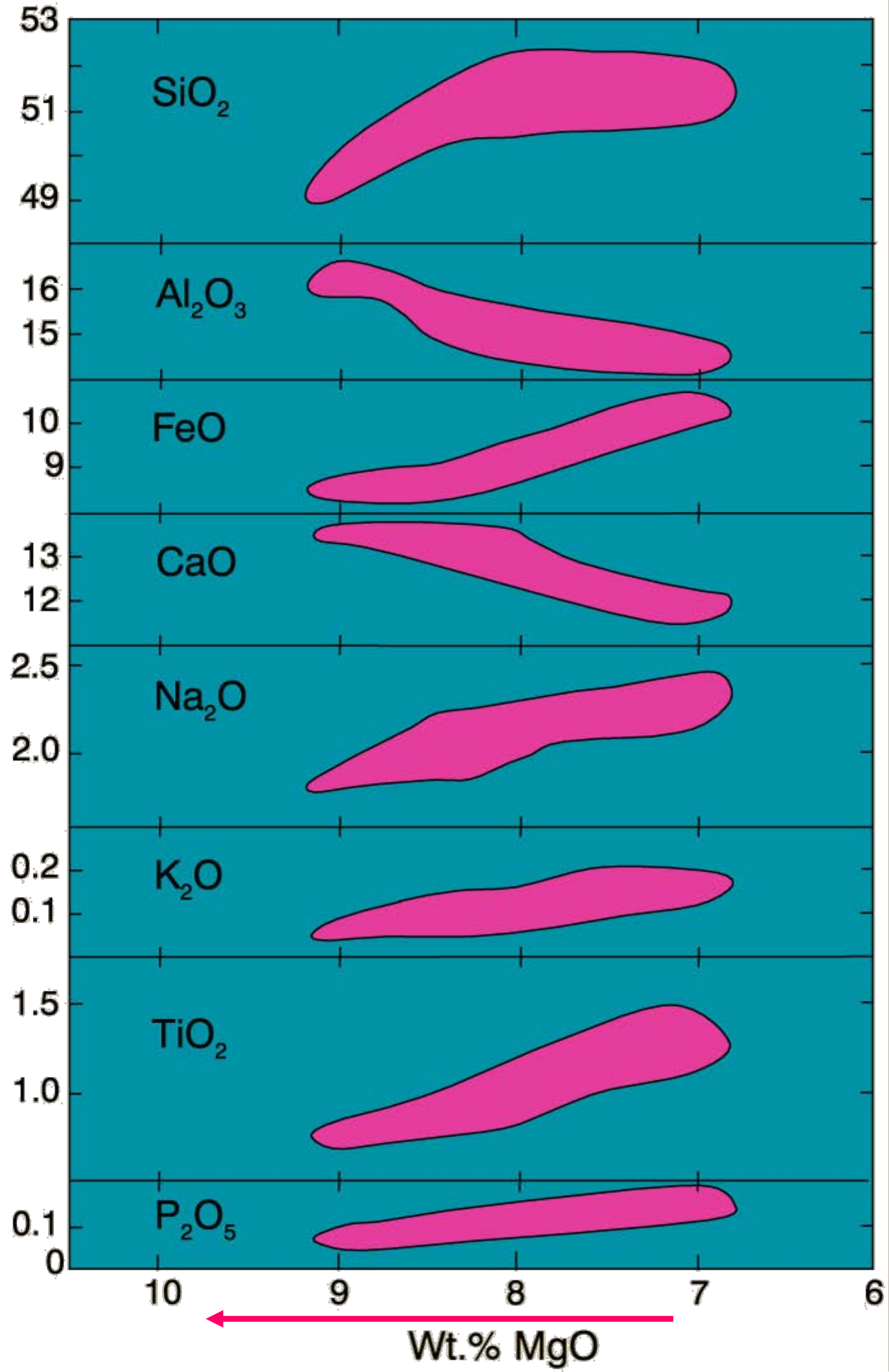



**Figure 13.6.** **"Fenner-type" variation diagrams for basaltic glasses from the Afar region of the MAR. Note different ordinate scales. From Stakes et al. (1984) J. Geophys. Res., 89, 6995-7028.**

Conclusions about MORBs, and the processes beneath mid-ocean ridges

- MORBs are not such completely uniform magmas
  - Chemical trends consistent with fractional crystallization of olivine, plagioclase, and perhaps clinopyroxene
- MORBs cannot be primary magmas, but are derivative magmas resulting from fractional crystallization (up to ~ 60%)

- Fast ridge segments (EPR) → a broader range of compositions and a larger proportion of evolved liquids
- (magmas erupted slightly off the axis of ridges are more evolved than those at the axis itself)

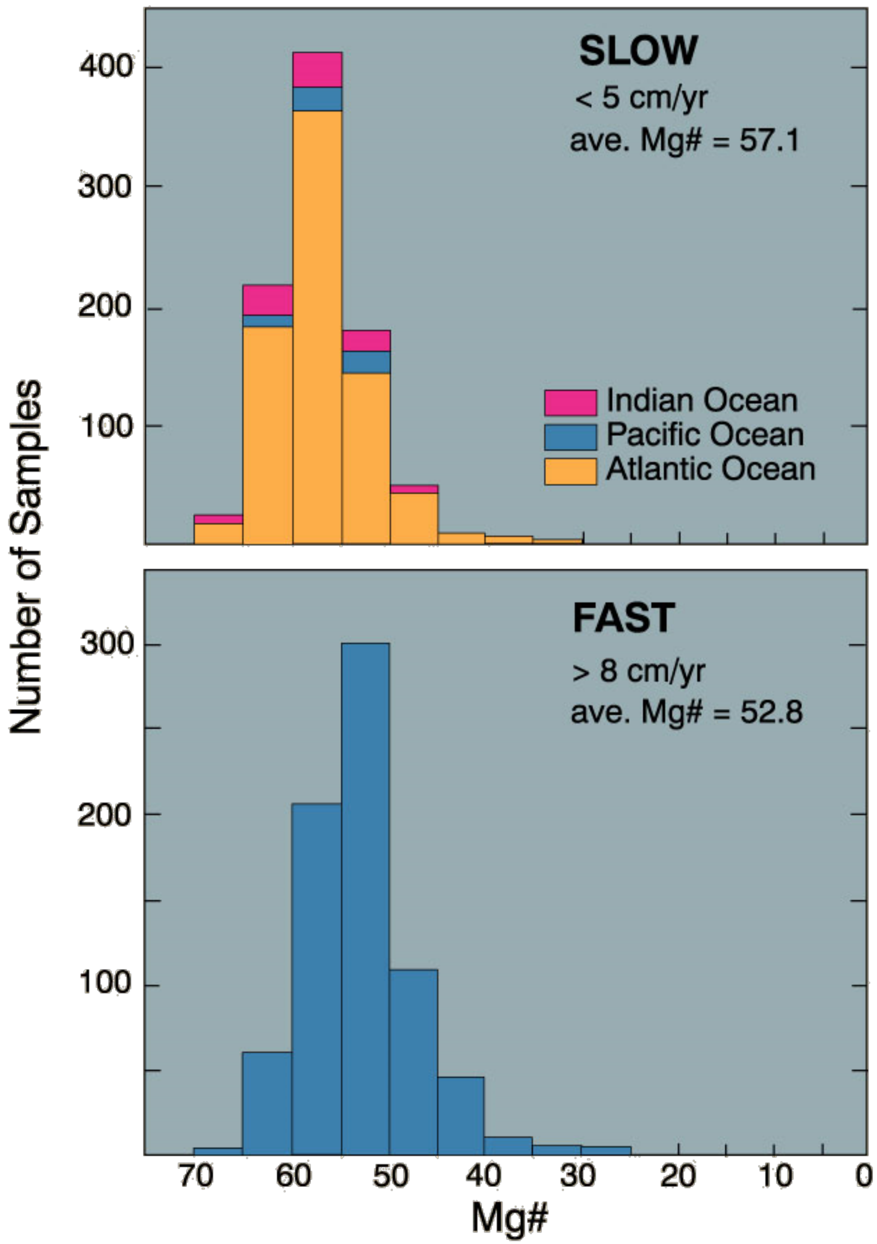



**Figure 13.9.** Histograms of over 1600 glass compositions from slow and fast mid-ocean ridges. After Sinton and Detrick (1992) J. Geophys. Res., 97, 197-216.

- For constant Mg# considerable variation is still apparent.

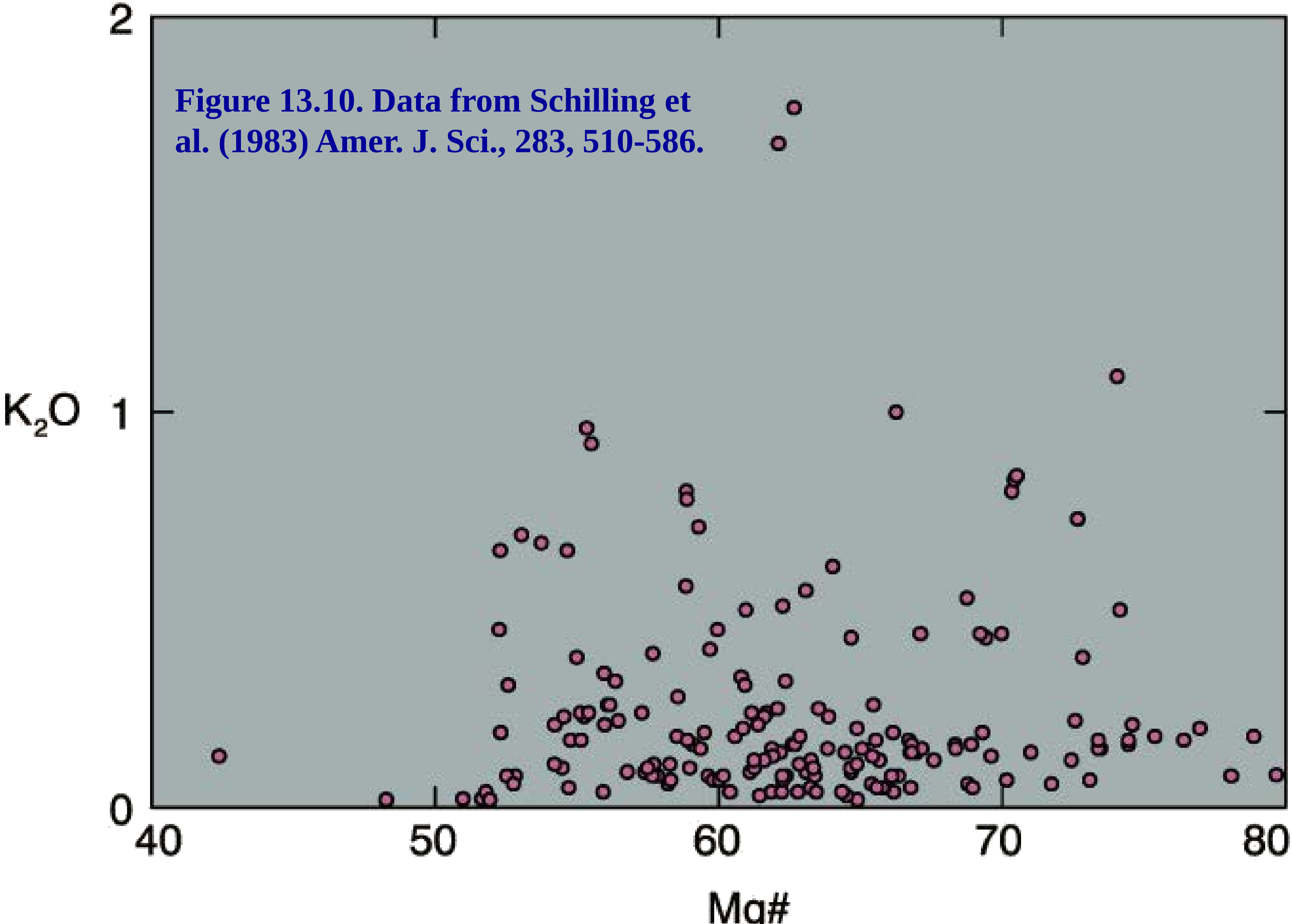



**Figure 13.10. Data from Schilling et al. (1983) Amer. J. Sci., 283, 510-586.**

Incompatible-rich and incompatible-poor mantle source regions for MORB magmas

- N-MORB (normal MORB) taps the depleted upper mantle source
  - Mg# > 65: $K_2O < 0.10$   $TiO_2 < 1.0$
- E-MORB (enriched MORB, also called P-MORB for plume) taps the (deeper) fertile mantle
  - Mg# > 65: $K_2O > 0.10$   $TiO_2 > 1.0$

# Trace Element and Isotope Chemistry

## REE diagram for MORBs

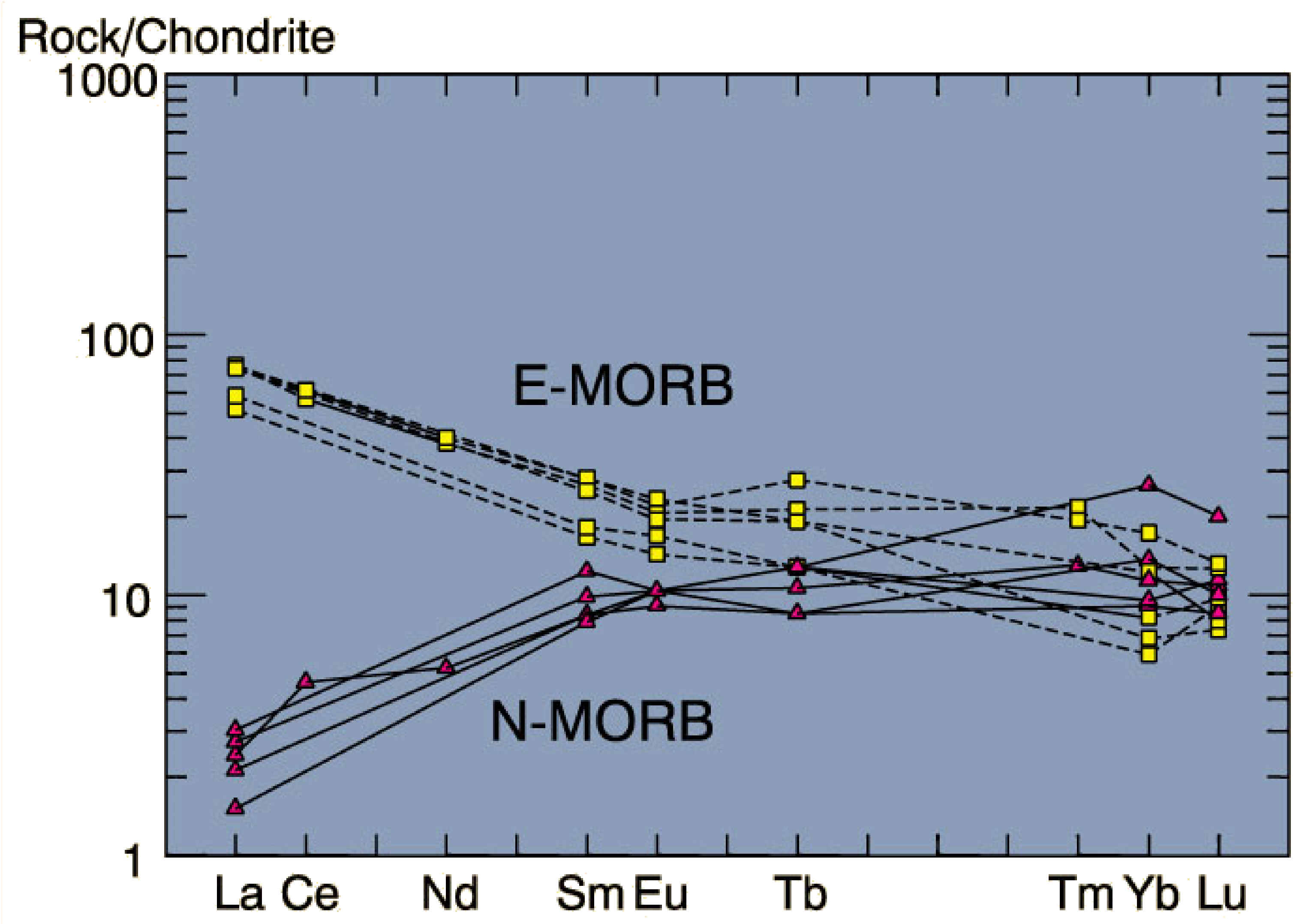


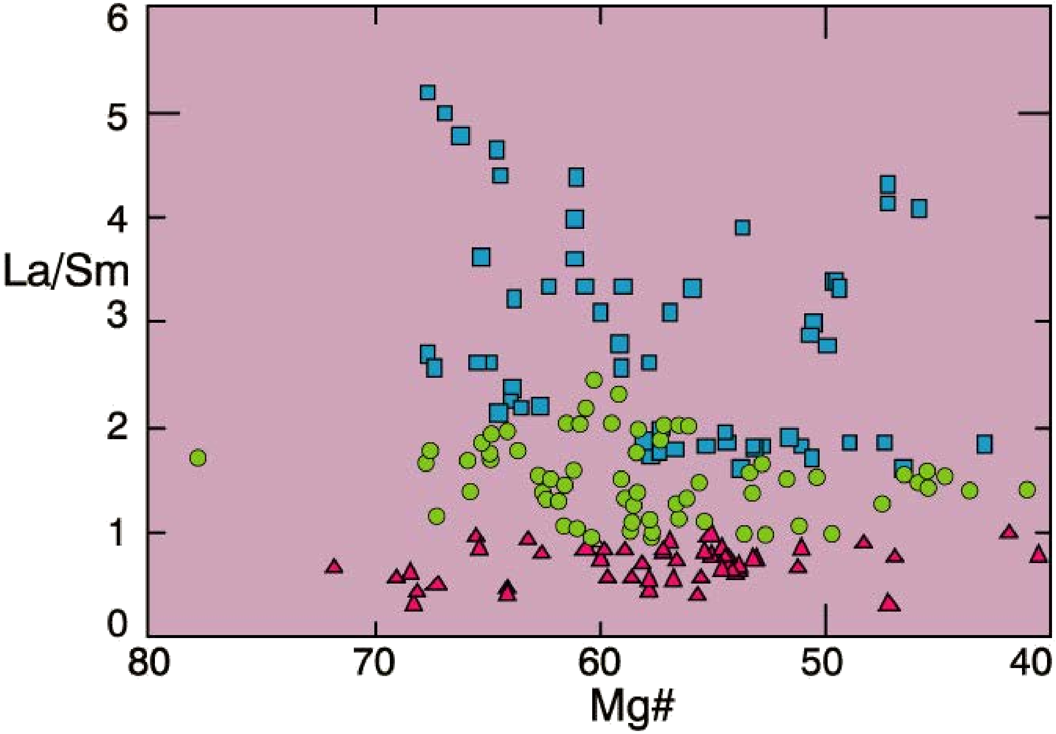

**Figure 13.11.** Data from Schilling et al. (1983) Amer. J. Sci., 283, 510-586.

E-MORBs are enriched over N-MORBs: regardless of Mg#

Lack of a distinct break suggests three MORB types

- E-MORBs ■ La/Sm > 1.8
- N-MORBs ▲ La/Sm < 0.7
- T-MORBs (transitional) ● intermediate values

**Figure 13.12.** **Data from Schilling et al. (1983) Amer. J. Sci., 283, 510-586.**



- N-MORBs: $^{87}Sr/^{86}Sr < 0.7035$ and $^{143}Nd/^{144}Nd > 0.5030$, → *depleted mantle source*
- E-MORBs extend to more enriched values → stronger support distinct mantle reservoirs for N-type and E-type MORBs

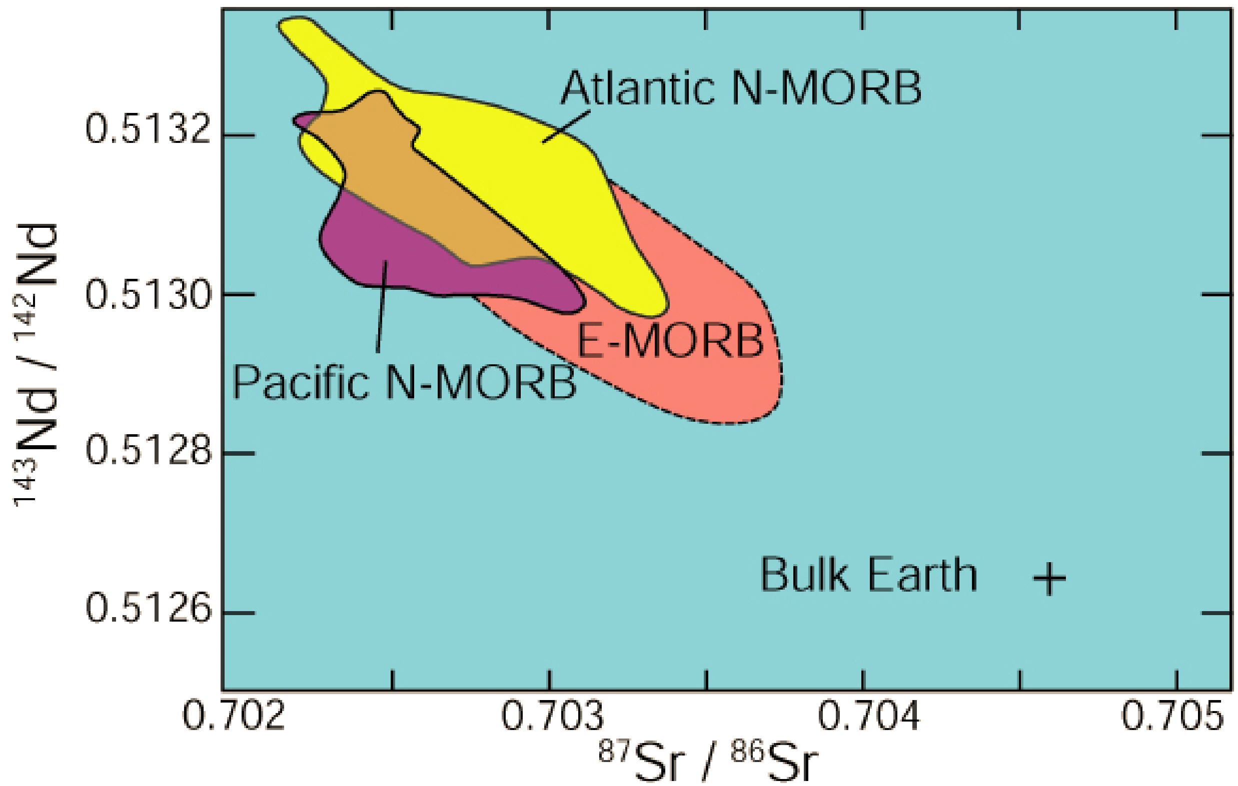


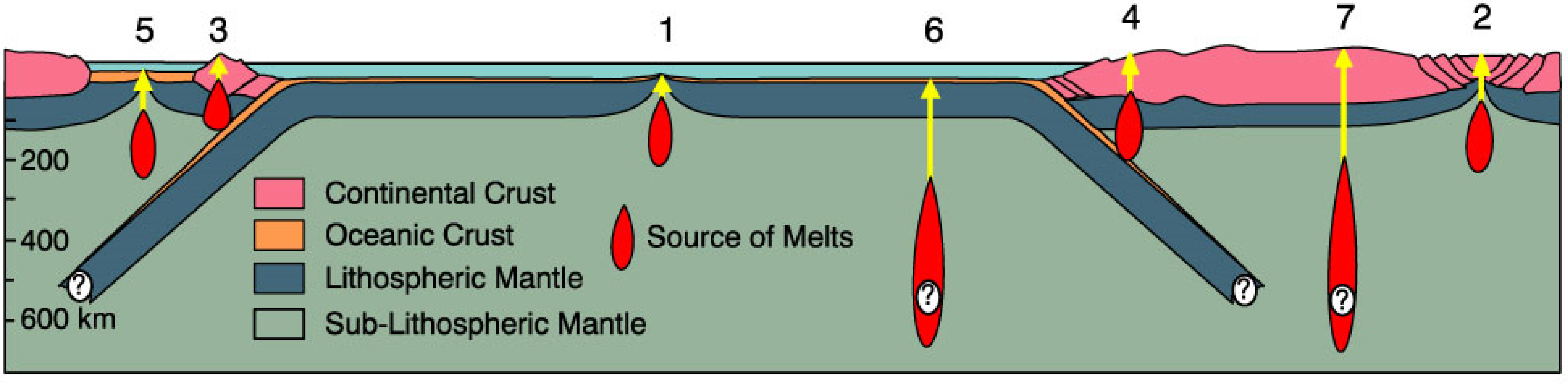

**Figure 13.13.** Data from Ito et al. (1987) Chemical Geology, 62, 157-176; and LeRoex et al. (1983) J. Petrol., 24, 267-318.

## Conclusions:

- MORBs have > 1 source region
- The mantle beneath the ocean basins is not homogeneous
  - N-MORBs tap an upper, depleted mantle
  - E-MORBs tap a deeper enriched source
  - T-MORBs = mixing of N- and E- magmas during ascent and/or in shallow chambers

# Tectonic-Igneous Associations

- Associations on a larger scale than the petrogenetic provinces
- An attempt to address global patterns of igneous activity by grouping provinces based upon similarities in occurrence and genesis



# Tectonic-Igneous Associations

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- **Ocean Intra-plate (Island) volcanism**
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- Mostly alkaline igneous processes of stable craton interiors
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# Chapter 14: Ocean Intraplate Volcanism

## Ocean islands and seamounts

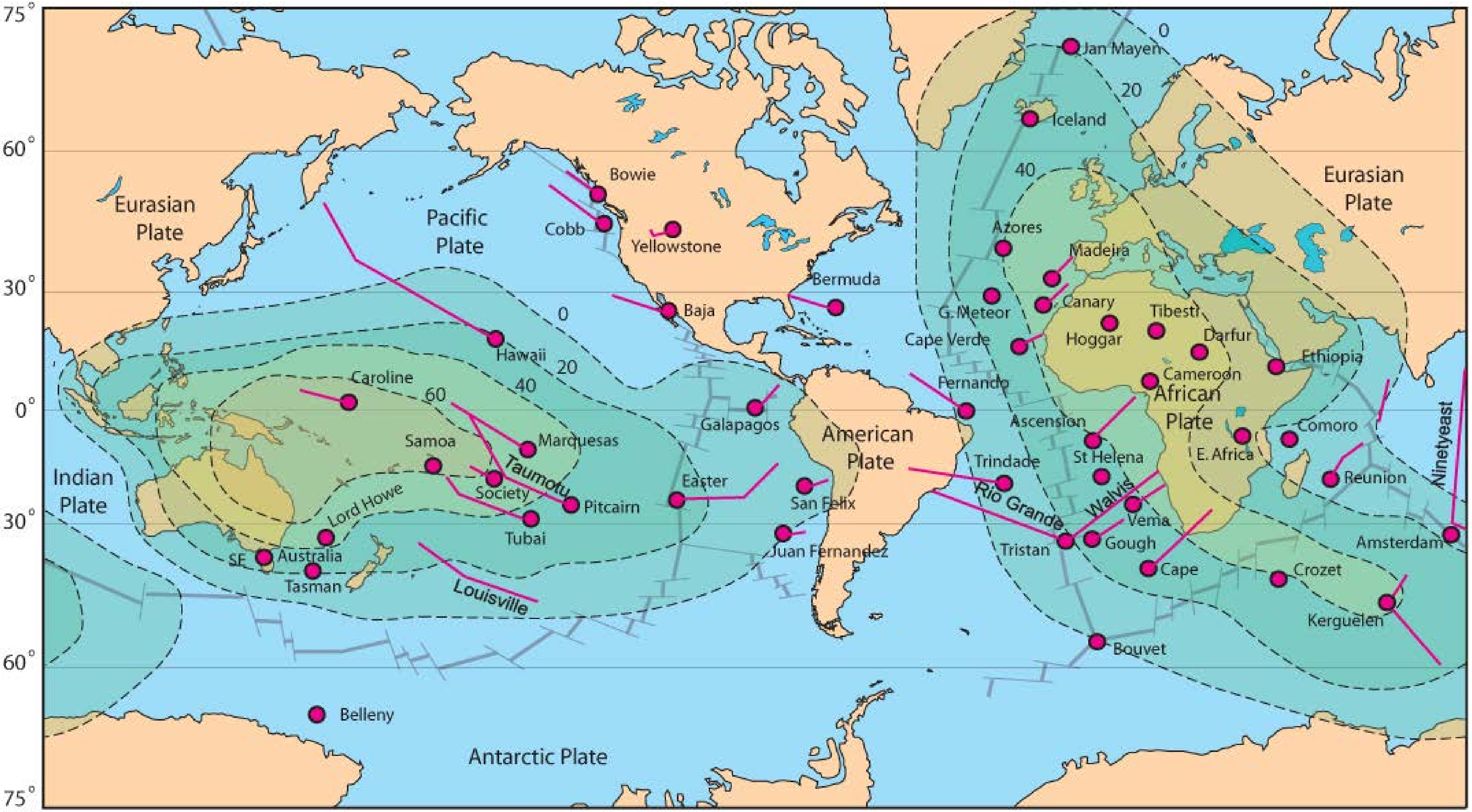
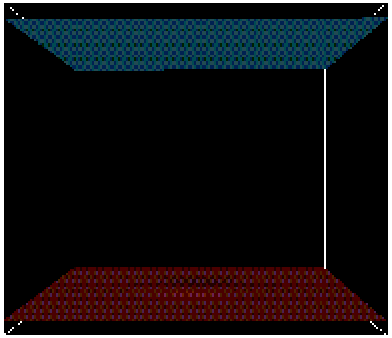
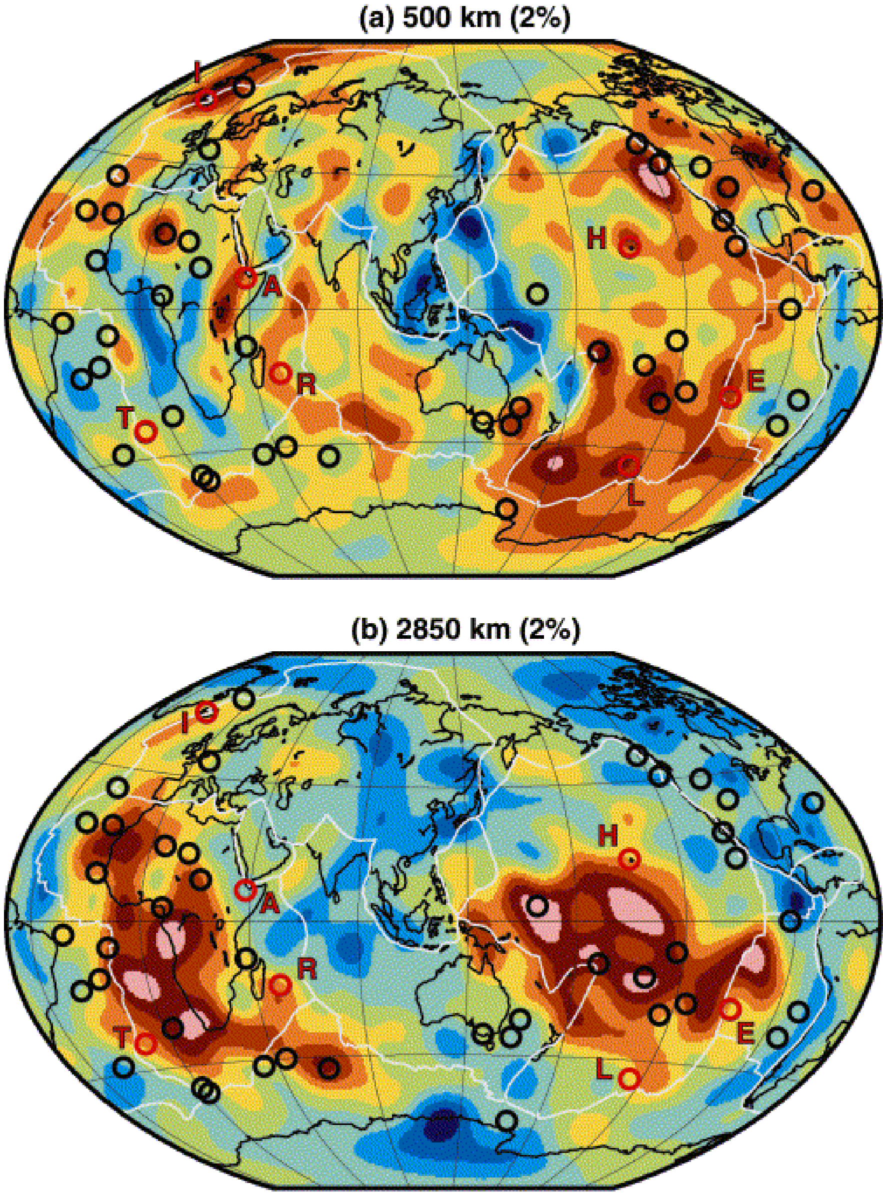



**Figure 14.1.** Map of relatively well-established hotspots and selected hotspot trails (island chains or aseismic ridges). Hotspots and trails from Crough (1983) with selected more recent hotspots from Anderson and Schramm (2005). Also shown are the geoid anomaly contours of Crough and Jurdy (1980, in meters). Note the preponderance of hotspots in the two major geoid highs (superswells).

# Ocean islands and seamounts

## Commonly associated with ***hotspots***





# Plume

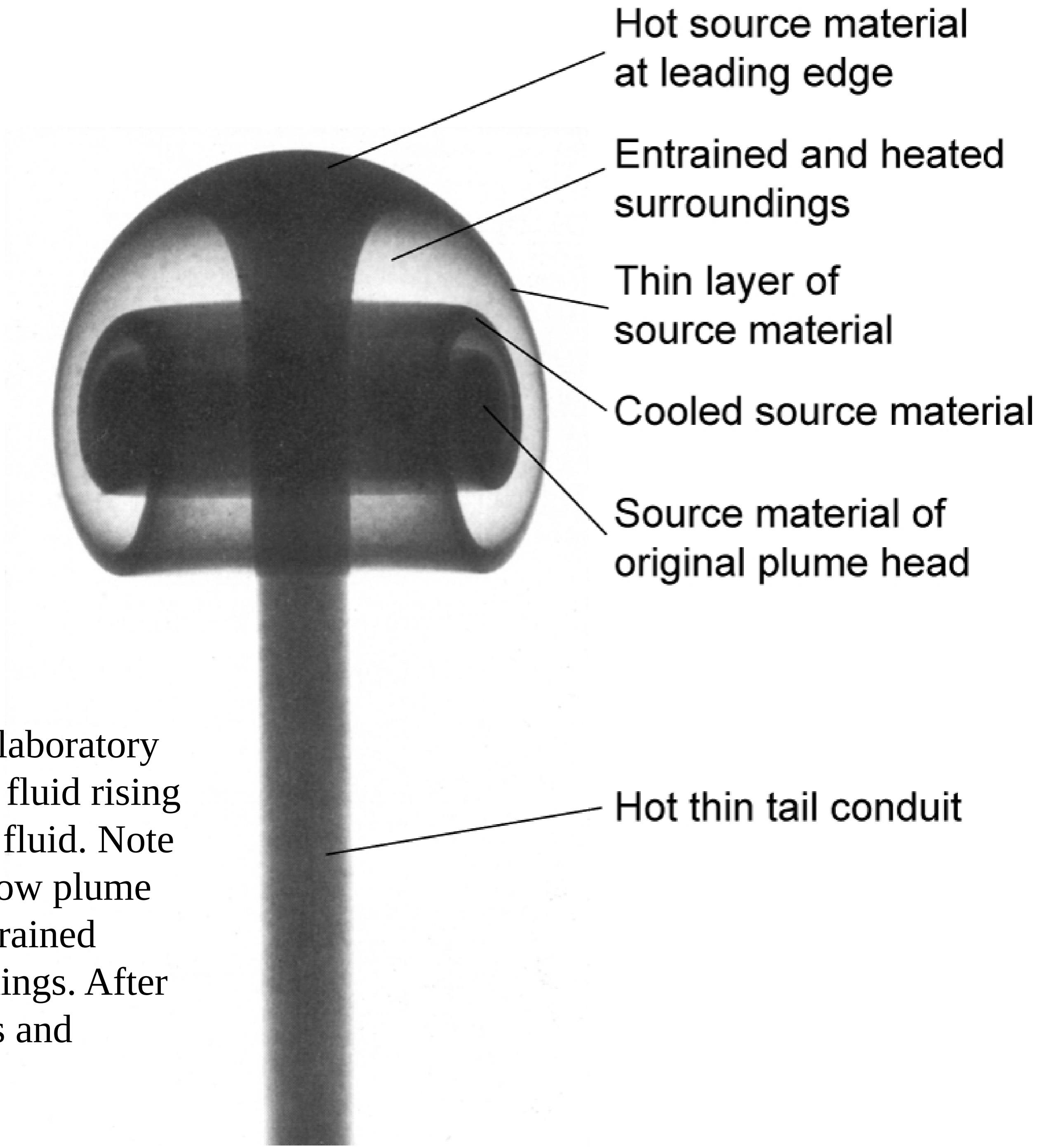



**Figure 14.2** Photograph of a laboratory thermal plume of heated dyed fluid rising buoyantly through a colorless fluid. Note the enlarged plume head, narrow plume tail, and vortex containing entrained colorless fluid of the surroundings. After Campbell (1998) and Griffiths and Campbell (1990).

# Types of OIB Magmas

Two principal magma series

***Tholeiitic*** (dominant type)

- Parent: ocean island tholeiitic basalt (OIT)
- Similar to MORB, but some distinct chemical and mineralogical differences

***Alkaline*** series (subordinate)

- Parent: ocean island alkaline basalt (OIA)
- Two principal alkaline sub-series
  - Silica undersaturated
  - Slightly silica oversaturated (less common)

# Hawaiian Scenario

Cyclic, pattern to the eruptive history

1. ***Pre-shield-building stage*** (variable)
2. ***Shield-building stage*** begins with tremendous outpourings of tholeiitic basalts
3. Waning activity more alkaline, episodic, diverse, and violent (Mauna Kea, Hualalai, and Kohala).
4. A long period of dormancy, followed by a late, ***post-erosional stage***. Characterized by highly alkaline and silica-undersaturated magmas, including alkali basalts, nephelinites, melilite basalts, and basanites

# Evolution in the Series

Tholeiitic, alkaline, and highly alkaline

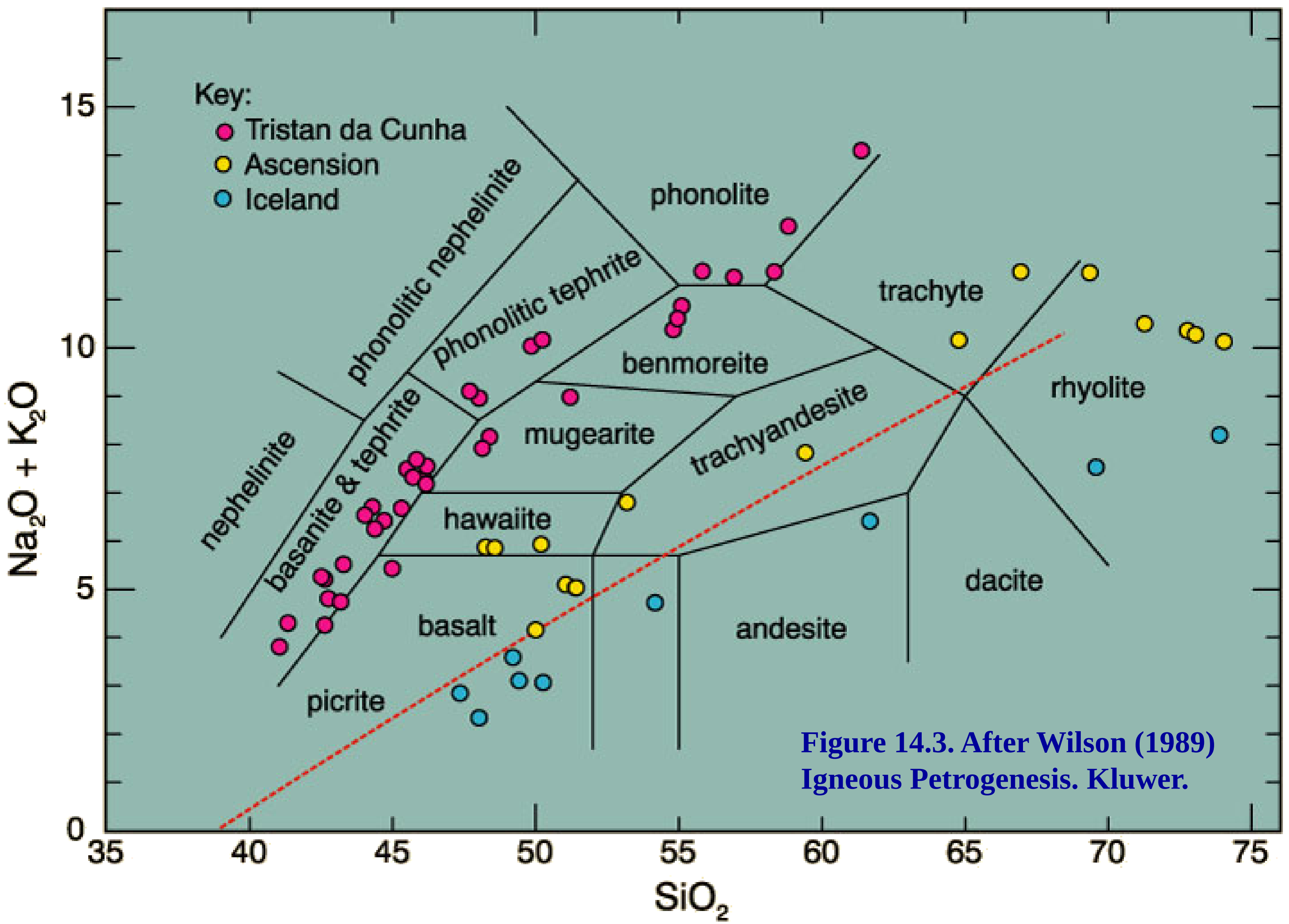



Figure 14.3. After Wilson (1989) Igneous Petrogenesis. Kluwer.

Alkalinity is highly variable

Alkalis are incompatible elements, unaffected by less than 50% shallow fractional crystallization, this again argues for distinct mantle sources or generating mechanisms

**Table 14-4**. Alkali/silica ratios (regression) for selected ocean island lava suites.

| Island | Alk/Silica | $Na_2O/SiO_2$ | $K_2O/SiO_2$ |
|---|---|---|---|
| Tahiti | 0.86 | 0.54 | 0.32 |
| Principe | 0.86 | 0.52 | 0.34 |
| Trinidade | 0.83 | 0.47 | 0.35 |
| Fernando de Noronha | 0.74 | 0.42 | 0.33 |
| Gough | 0.74 | 0.30 | 0.44 |
| St. Helena | 0.56 | 0.34 | 0.22 |
| Tristan da Cunha | 0.46 | 0.24 | 0.22 |
| Azores | 0.45 | 0.24 | 0.21 |
| Ascension | 0.42 | 0.18 | 0.24 |
| Canary Is | 0.41 | 0.22 | 0.19 |
| Tenerife | 0.41 | 0.20 | 0.21 |
| Galapagos | 0.25 | 0.12 | 0.13 |
| Iceland | 0.20 | 0.08 | 0.12 |

# Trace Elements

The ***LIL*** trace elements (K, Rb, Cs, Ba, $Pb^{2+}$ and Sr) are incompatible and are all enriched in OIB magmas with respect to MORBs

The ratios of incompatible elements have been employed to distinguish between source reservoirs

- N-MORB: the K/Ba ratio is high (usually > 100)
- E-MORB: the K/Ba ratio is in the mid 30's
- OITs range from 25-40, and OIAs in the upper 20's

Thus all appear to have distinctive sources

# Trace Elements

***HFS*** elements (Th, U, Ce, Zr, Hf, Nb, Ta, and Ti) are also incompatible, and are enriched in OIBs > MORBs

Ratios of these elements are also used to distinguish mantle sources

- The Zr/Nb ratio
  - N-MORBs are generally quite high (>30)
  - OIBs are low (<10)

# Trace Elements: REEs

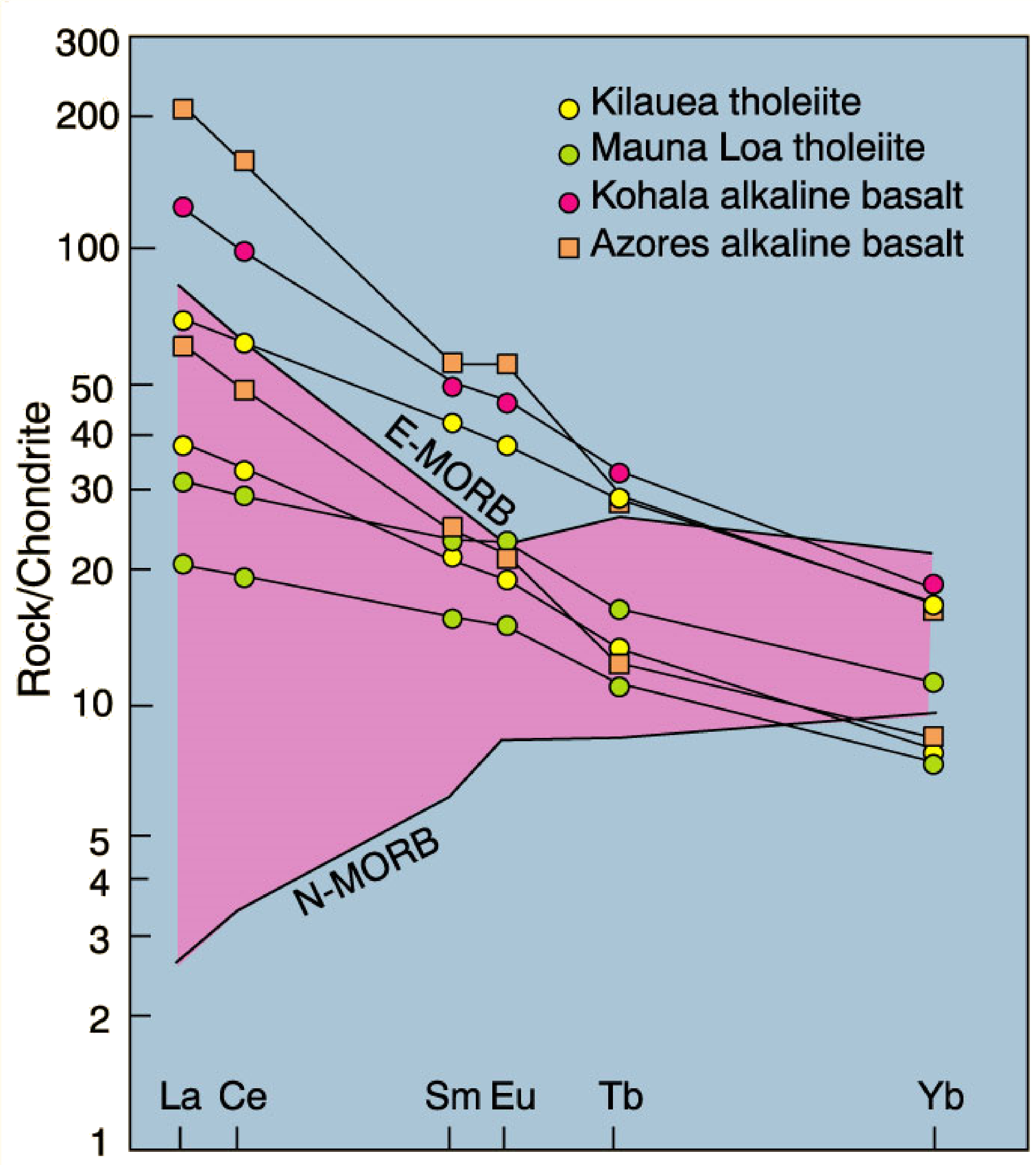



**Figure 14.4.** After Wilson (1989) Igneous Petrogenesis. Kluwer.

# Trace Elements: REEs

La/Yb (REE slope) correlates with the degree of silica undersaturation in OIBs

- Highly undersaturated magmas: La/Yb > 30
- OIA: closer to 12
- OIT: ~ 4
- (-) slopes → E-MORB and all OIBs ≠ N-MORB
- (-) slope and appear to originate in the lower enriched mantle

# MORB-normalized Spider Diagrams

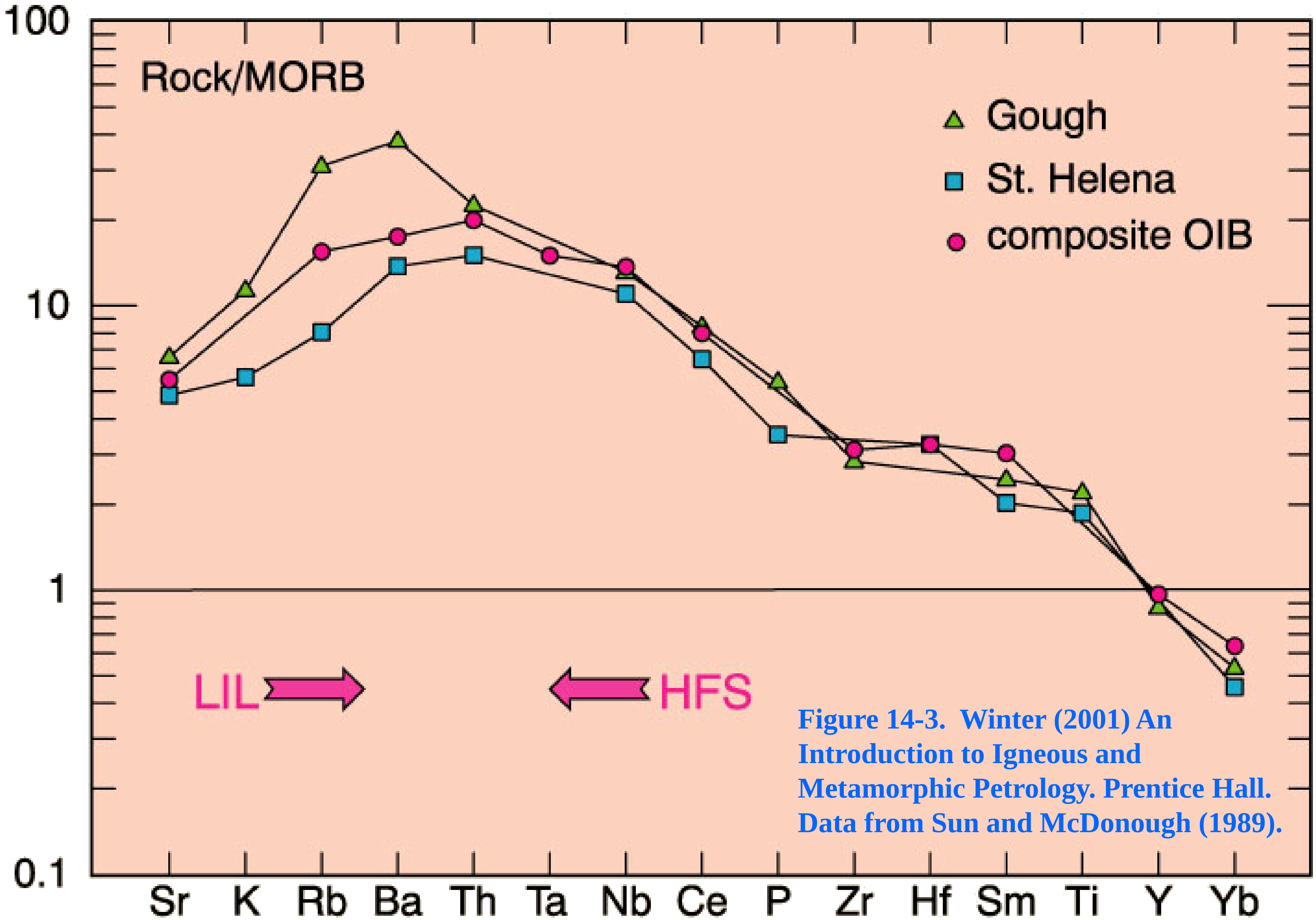



**Figure 14-3. Winter (2001) An Introduction to Igneous and Metamorphic Petrology. Prentice Hall. Data from Sun and McDonough (1989).**

# Isotope Geochemistry

Isotopes do not fractionate during partial melting of fractional melting processes, so will reflect the characteristics of the source

OIBs, which sample a great expanse of oceanic mantle in places where crustal contamination is minimal, provide incomparable evidence as to the nature of the mantle

# Simple Mixing Models

## Binary

All analyses fall between two reservoirs as magmas mix

## Ternary

All analyses fall within triangle determined by three reservoirs

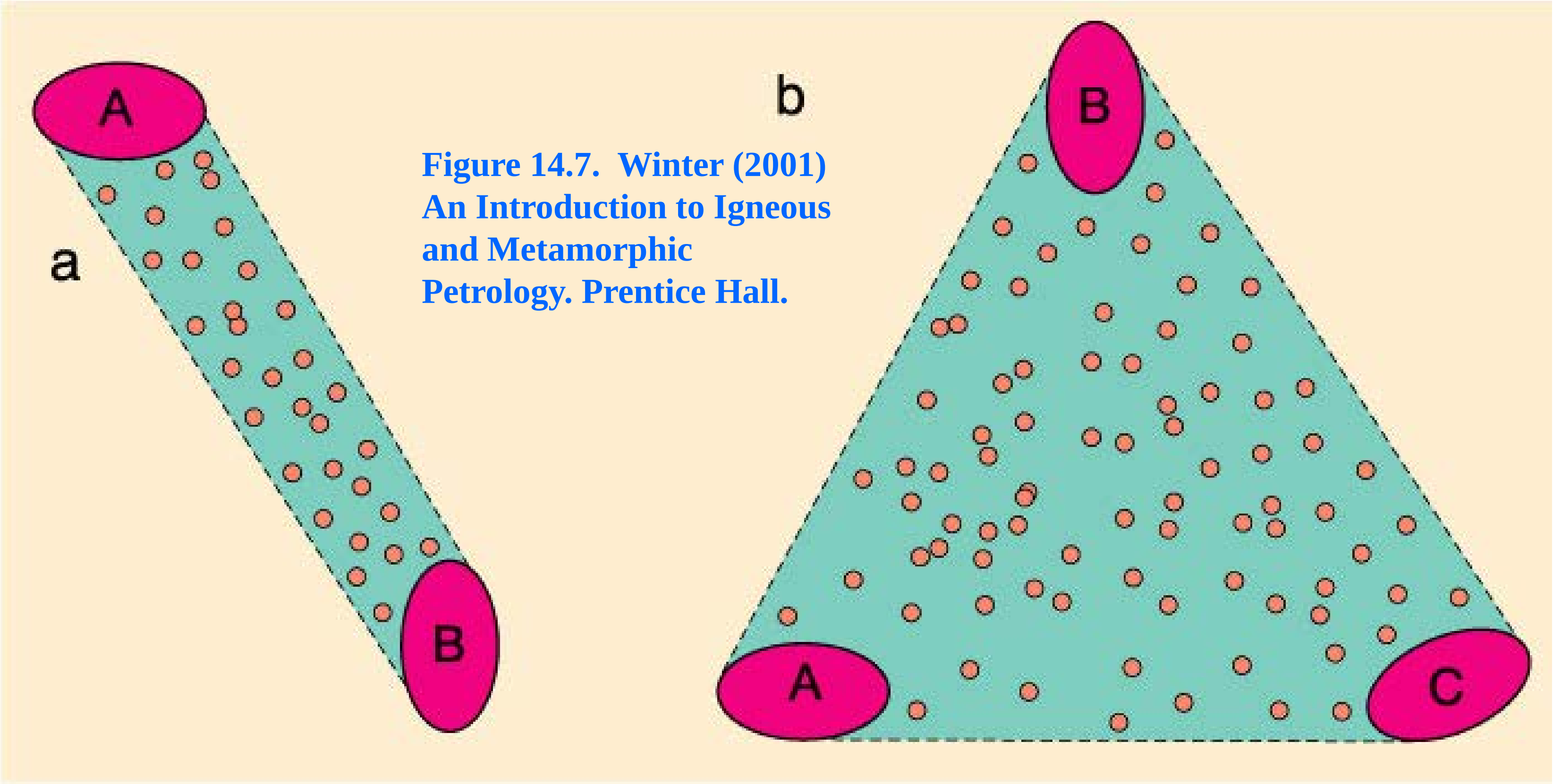



Figure 14.7. Winter (2001) An Introduction to Igneous and Metamorphic Petrology. Prentice Hall.

# Sr - Nd Isotopes

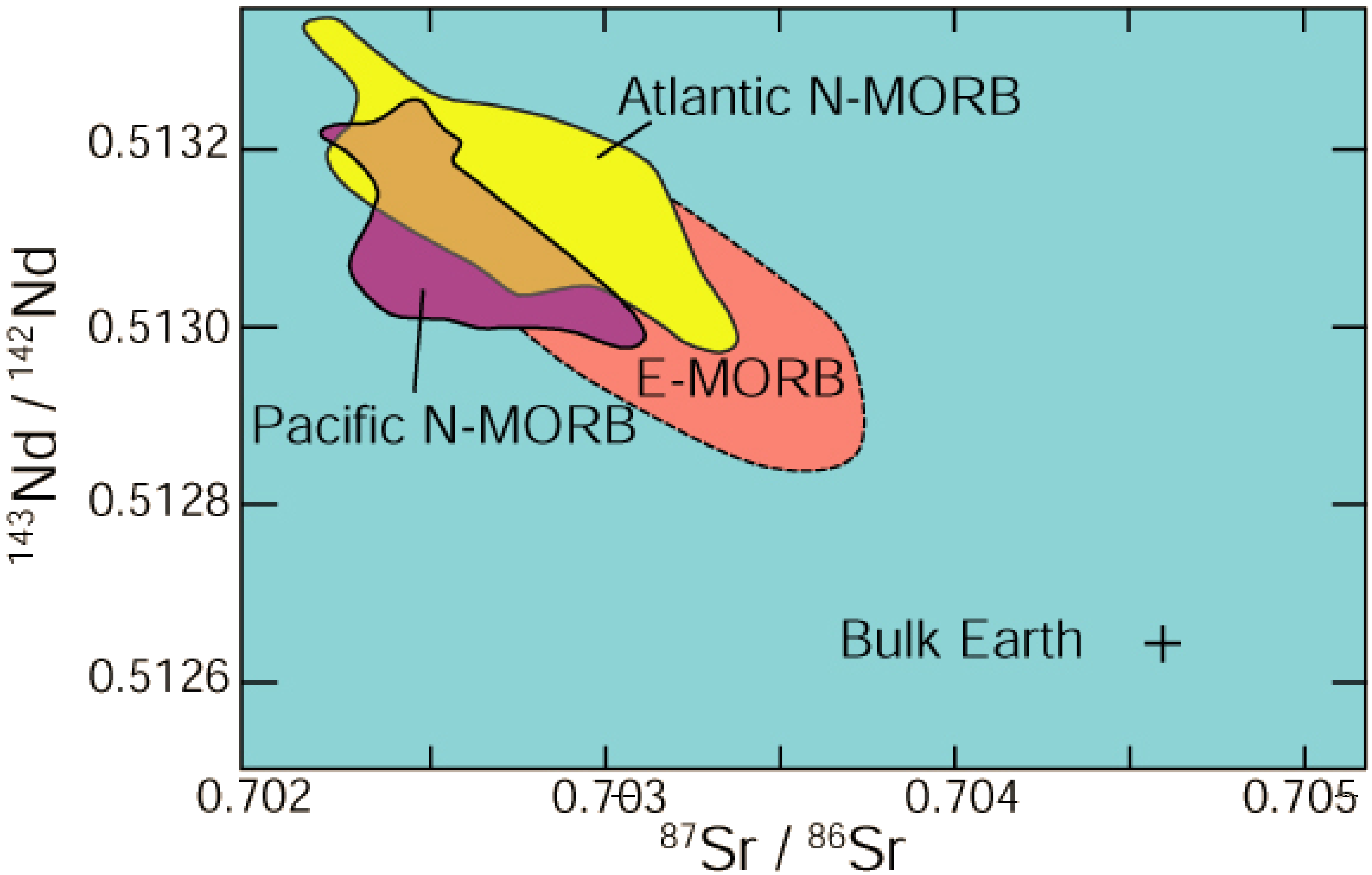
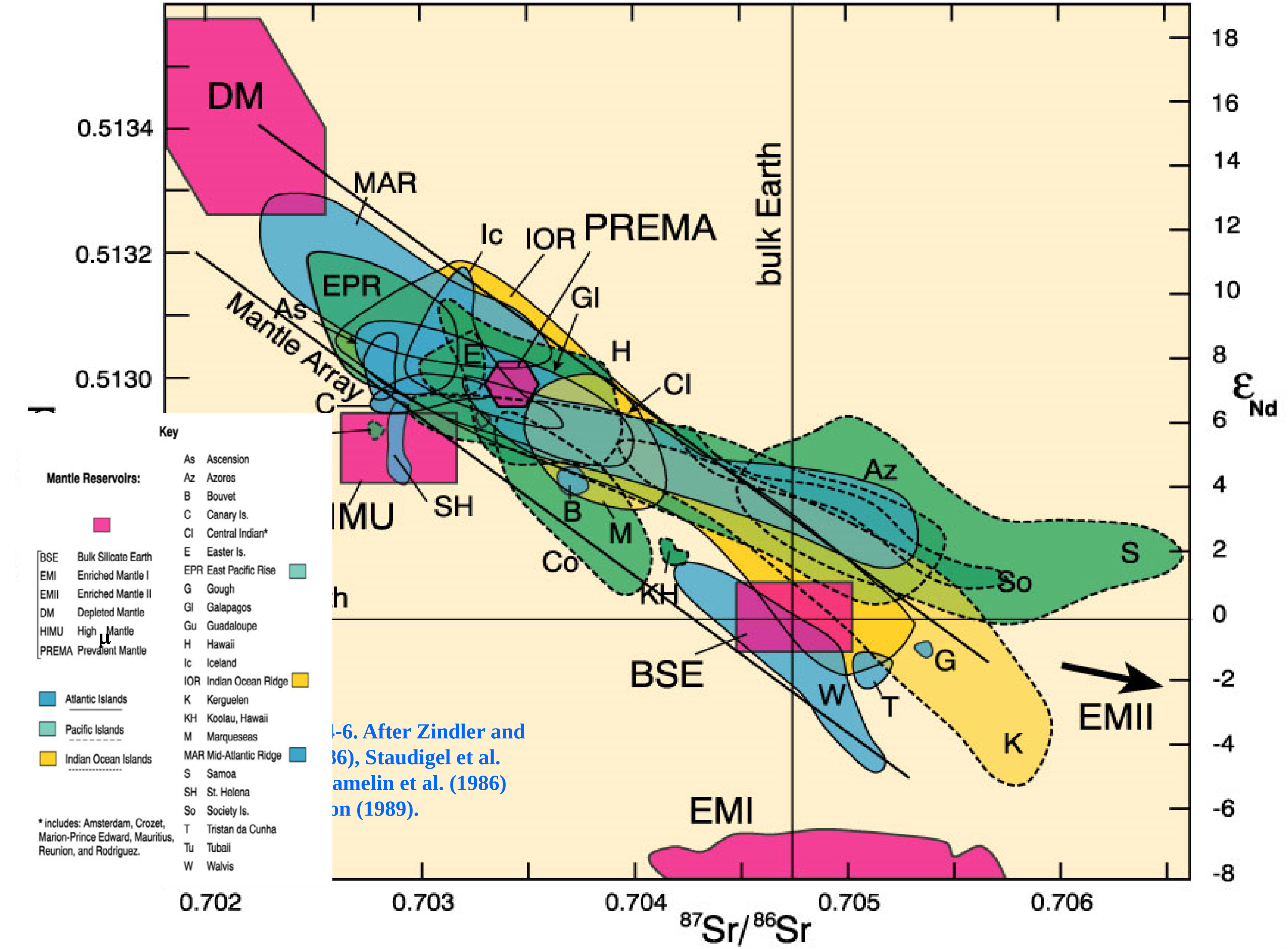



**Figure 13.13.** **Data from Ito et al. (1987) Chemical Geology, 62, 157-176; and LeRoex et al. (1983) J. Petrol., 24, 267-318.**




**-6. After Zindler and**
**86), Staudigel et al.**
**amelin et al. (1986)**
**on (1989).**

# Mantle Reservoirs

1. ***DM*** (Depleted Mantle) = N-MORB source

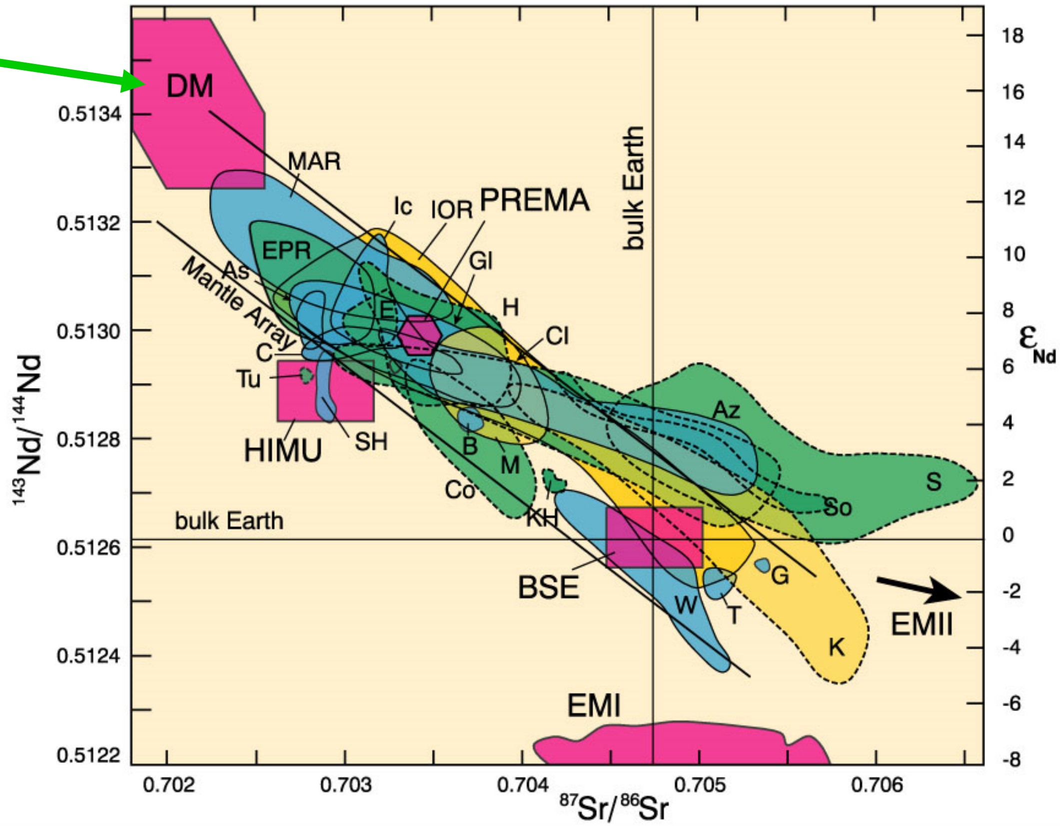



**Figure 14.8.** After Zindler and Hart (1986), Staudigel et al. (1984), Hamelin et al. (1986) and Wilson (1989).

# 2. ***BSE*** (Bulk Silicate Earth) or the Primary Uniform Reservoir

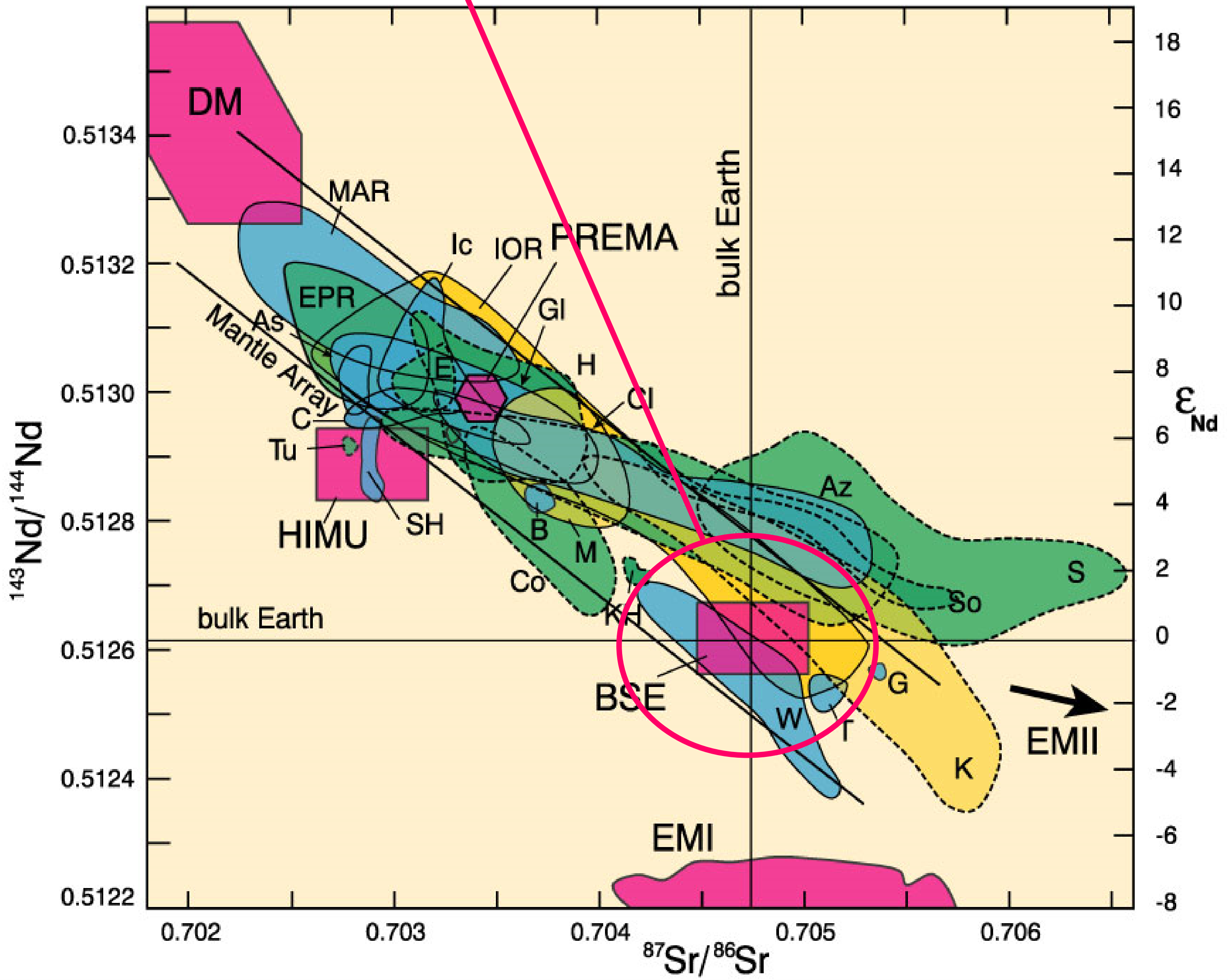



**Figure 14.8.** After Zindler and Hart (1986), Staudigel et al. (1984), Hamelin et al. (1986) and Wilson (1989).

3. ***EMI*** = enriched mantle type I has lower $^{87}Sr/^{86}Sr$ (near primordial)

4. ***EMII*** = enriched mantle type II has higher $^{87}Sr/^{86}Sr$ (> 0.720), well above any reasonable mantle sources

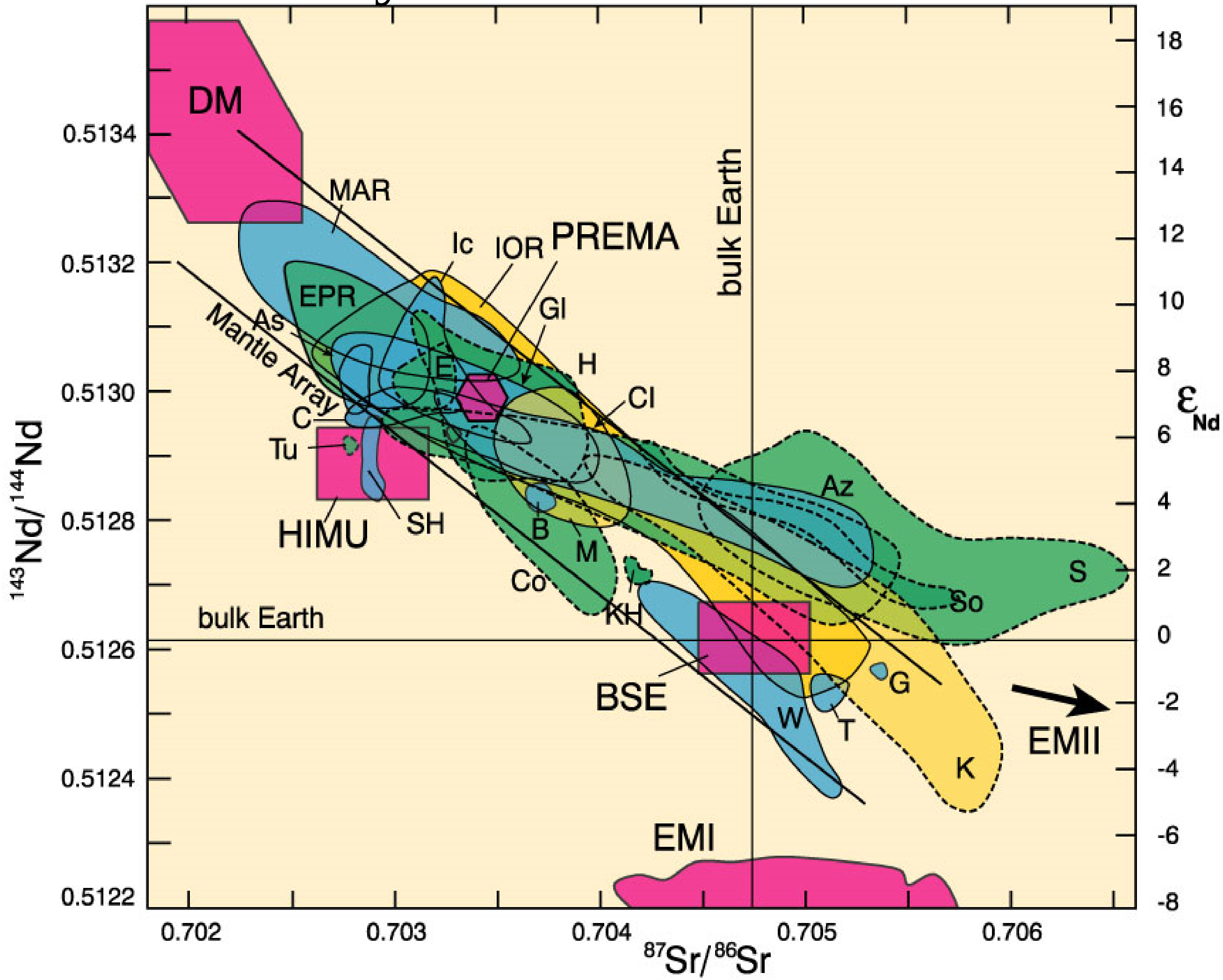



**Figure 14.8.** After Zindler and Hart (1986), Staudigel et al. (1984), Hamelin et al. (1986) and Wilson (1989).

# 5. ***PREMA*** (PREvalent MAntle)

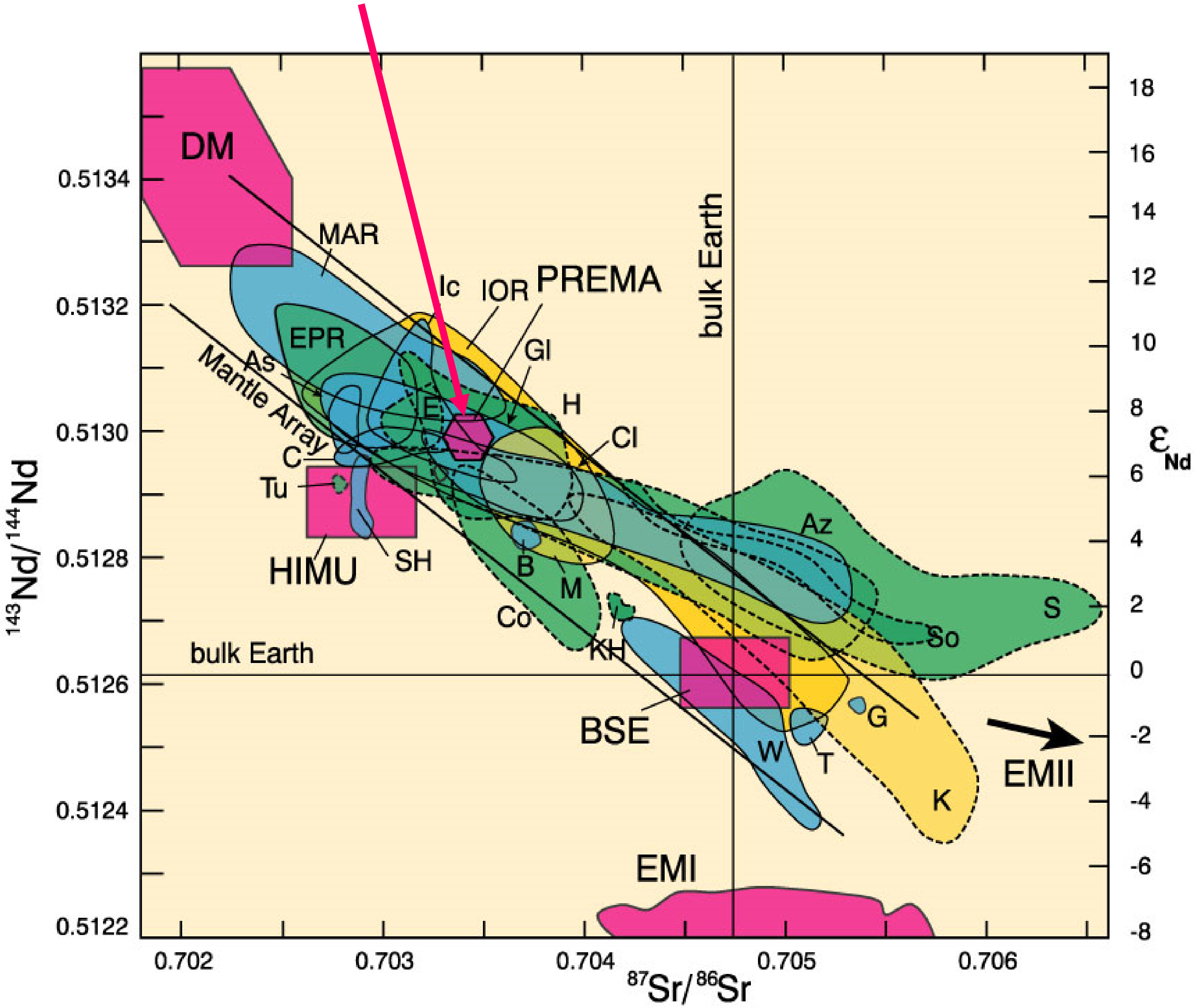



**Figure 14.8.** After Zindler and Hart (1986), Staudigel et al. (1984), Hamelin et al. (1986) and Wilson (1989).

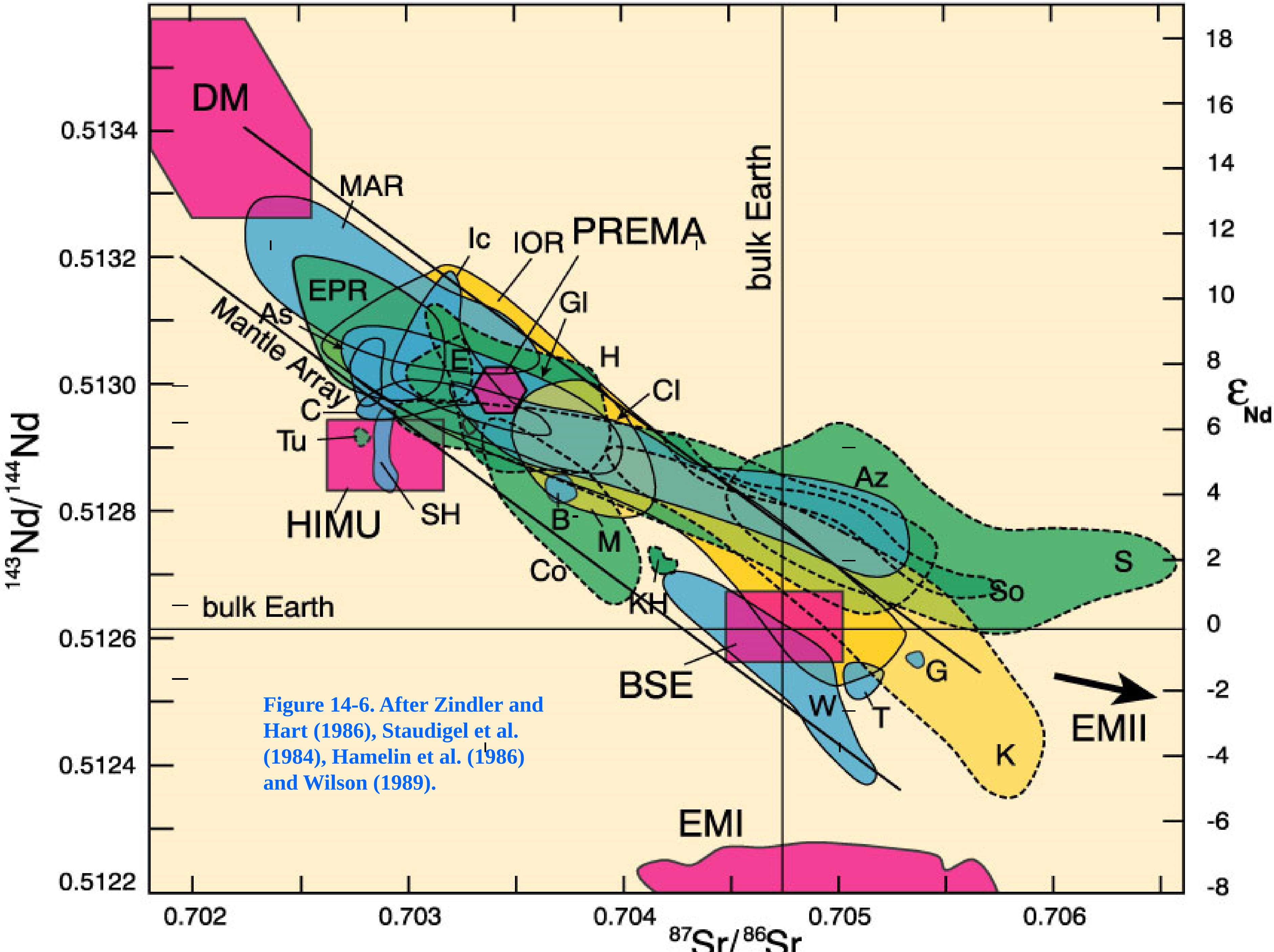



**Figure 14-6. After Zindler and Hart (1986), Staudigel et al. (1984), Hamelin et al. (1986) and Wilson (1989).**

# Pb Isotopes

Pb produced by radioactive decay of U & Th

Eq. 9.20 $^{238}U \rightarrow {}^{234}U \rightarrow {}^{206}Pb$

Eq. 9.21 $^{235}U \rightarrow {}^{207}Pb$

Eq. 9.22 $^{232}Th \rightarrow {}^{208}Pb$

# Pb is quite scarce in the mantle

- Low-Pb mantle-derived melts susceptible to Pb contamination
- U, Pb, and Th are concentrated in continental crust (high radiogenic daughter Pb isotopes)
- $^{204}Pb$ non-radiogenic: $^{208}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{206}Pb/^{204}Pb$ increase as U and Th decay
- Oceanic crust also has elevated U and Th content (compared to the mantle)
- Sediments derived from oceanic and continental crust
- Pb is a sensitive measure of crustal (including sediment) components in mantle isotopic systems
- 93.7% of natural U is $^{238}U$, so $^{206}Pb/^{204}Pb$ will be most sensitive to a crustal-enriched component

| | |
|---|---|
| **9-20** | $^{238}U \rightarrow {}^{234}U \rightarrow {}^{206}Pb$ |
| **9-21** | $^{235}U \rightarrow {}^{207}Pb$ |
| **9-22** | $^{232}Th \rightarrow {}^{208}Pb$ |

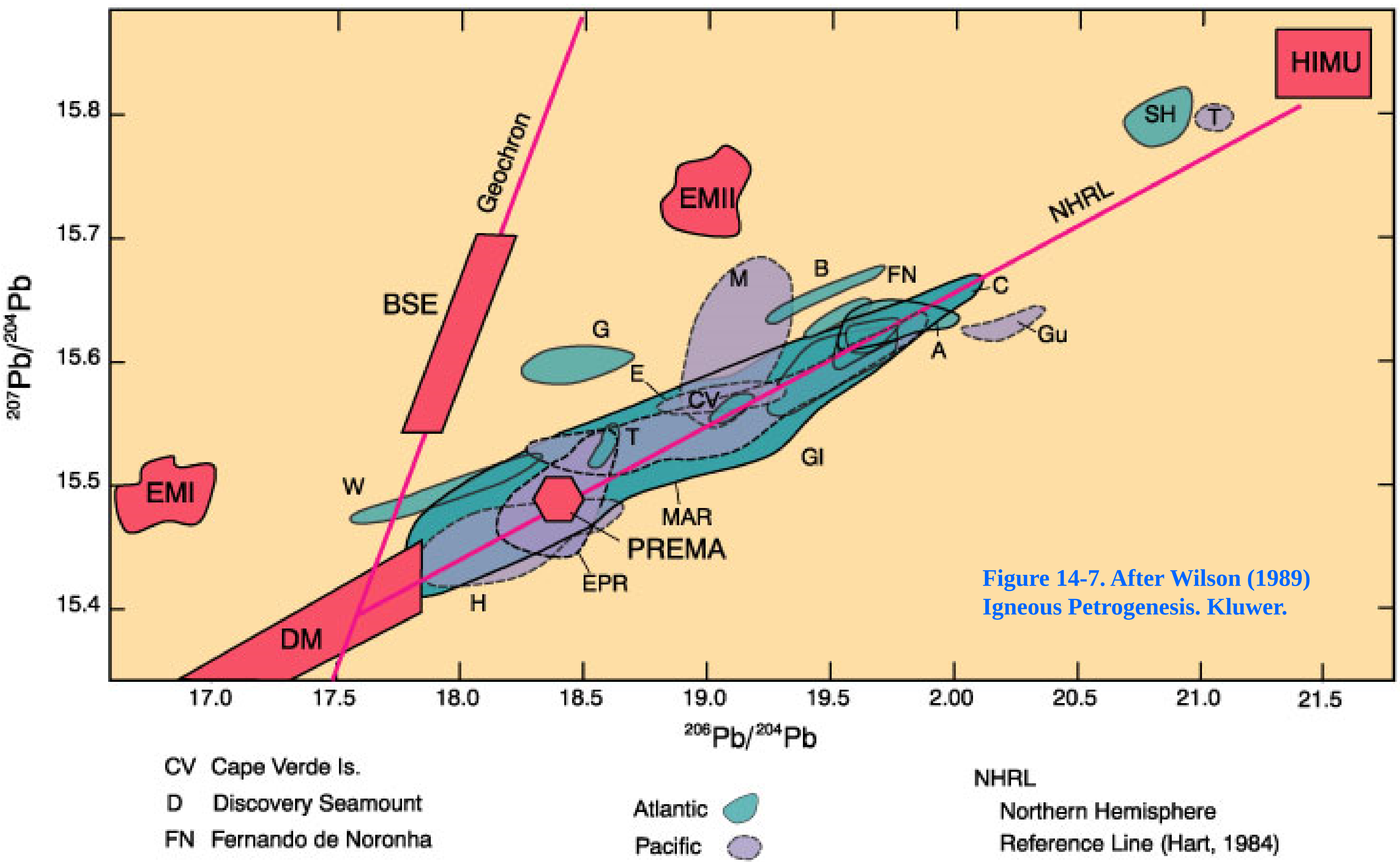



**Figure 14-7. After Wilson (1989) Igneous Petrogenesis. Kluwer.**

$\mu = {}^{238}U/{}^{204}Pb$ (evaluate uranium enrichment)

***HIMU*** reservoir: very high $^{206}Pb/^{204}Pb$ ratio

Source with high U, yet not enriched in Rb (modest $^{87}Sr/^{86}Sr$)

Old enough (> 1 Ga) to → observed isotopic ratios

HIMU models:

- ☞ Subducted and recycled oceanic crust (± seawater)
- ☞ Localized mantle lead loss to the core
- ☞ Pb-Rb removal by those dependable (but difficult to document) metasomatic fluids

# EMI and EMII

- High $^{87}Sr/^{86}Sr$ require high Rb & long time to → $^{87}Sr$
  - ☞ Correlates with continental crust (or sediments derived from it)
  - ☞ Oceanic crust and sediment are other likely candidates

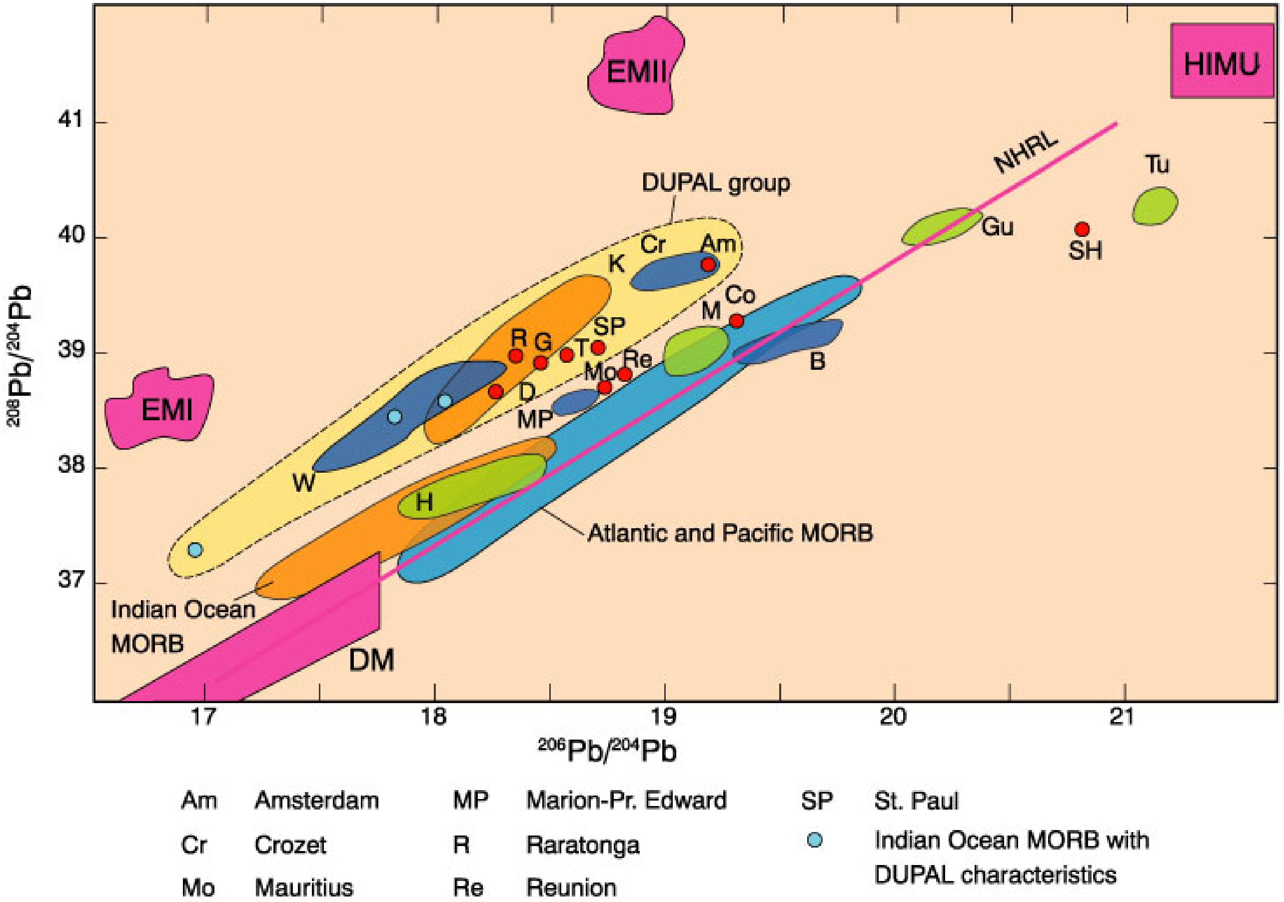



**Figure 14.10** **After Wilson (1989) Igneous Petrogenesis. Kluwer. Data from Hamelin and Allègre (1985), Hart (1984), Vidal et al. (1984).**

$^{207}Pb/^{204}Pb$ data (especially from the N hemisphere) → ~linear mixing line between DM and HIMU, a line called the Northern Hemisphere Reference Line (NHRL)

Data from the southern hemisphere (particularly Indian Ocean) departs from this line, and appears to include a larger EM component (probably EMII)

# Other isotopic systems that contribute to our understanding of mantle reservoirs and dynamics

## He Isotopes

Noble gases are inert and volatile

$^4$He is an alpha particle, produced principally by $\alpha$-decay of U and Th, enriching primordial $^4$He

$^3$He is largely primordial (constant)

The mantle is continually degassing and He lost (cannot recycle back)

$^4$He enrichment expressed as $R = (^3He/^4He)$

unusual among isotopes in that radiogenic is the denominator

Common reference is $R_A$ (air) = $1.39 \times 10^{-6}$

# He Isotopes

N-MORB is fairly uniform at 8±1 $R_A$ suggesting an extensive depleted (degassed) DM-type N-MORB source

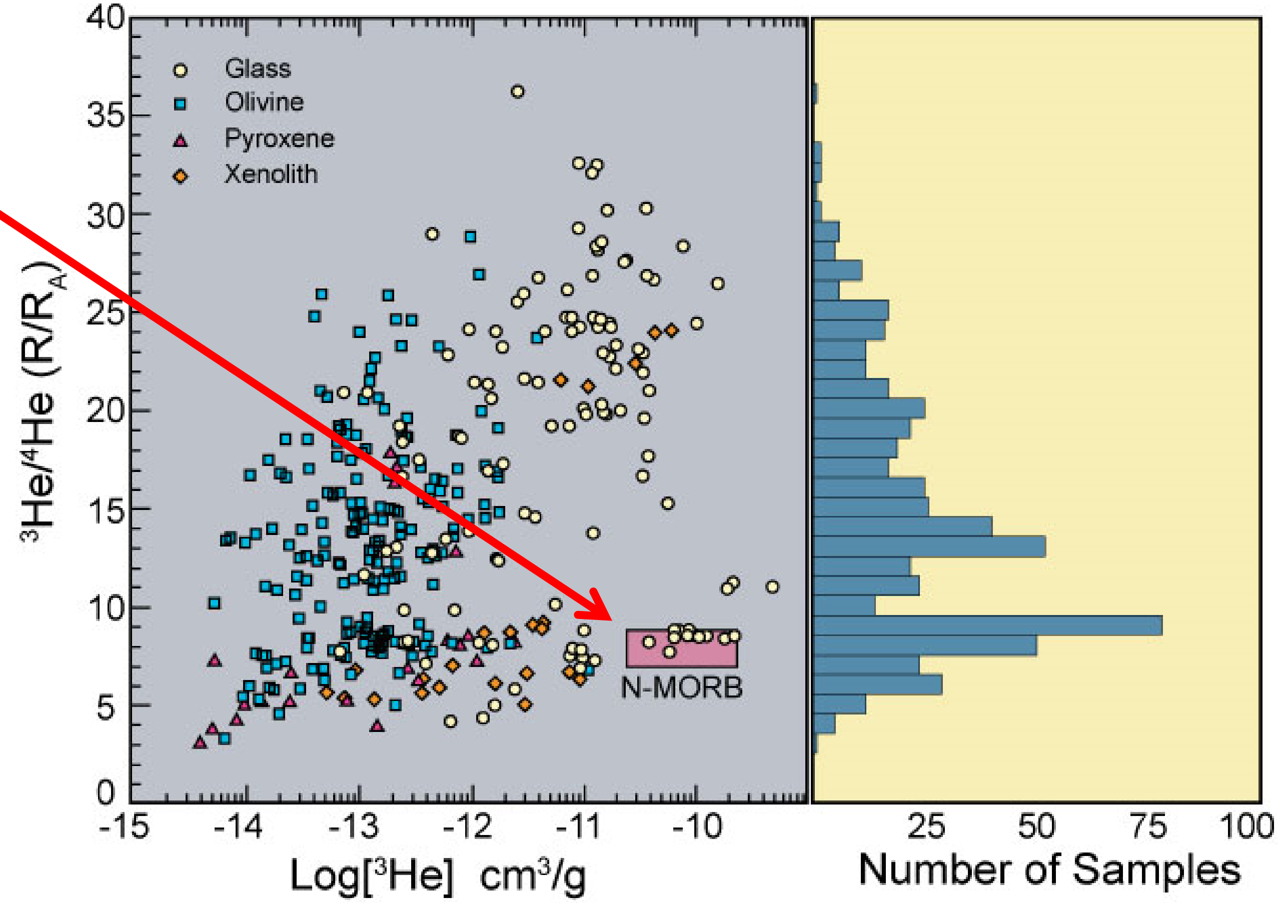



**Figure 14.12** $^3He/^4He$ isotope ratios in ocean island basalts and their relation to He concentration. Concentrations of $^3He$ are in $cm^3$ at 1 atm and 298K.After Sarda and Graham (1990) and Farley and Neroda (1998).

# He Isotopes

OIB $^3He/^4He$ values extend to both higher and lower values than N-MORBs, but are typically higher (low $^4He$).

Simplest explanations:

High $R/R_A$ is deeper mantle with more primordial signature

Low $R/R_A$ has higher $^4He$ due to recycled (EM-type?) U and Th.

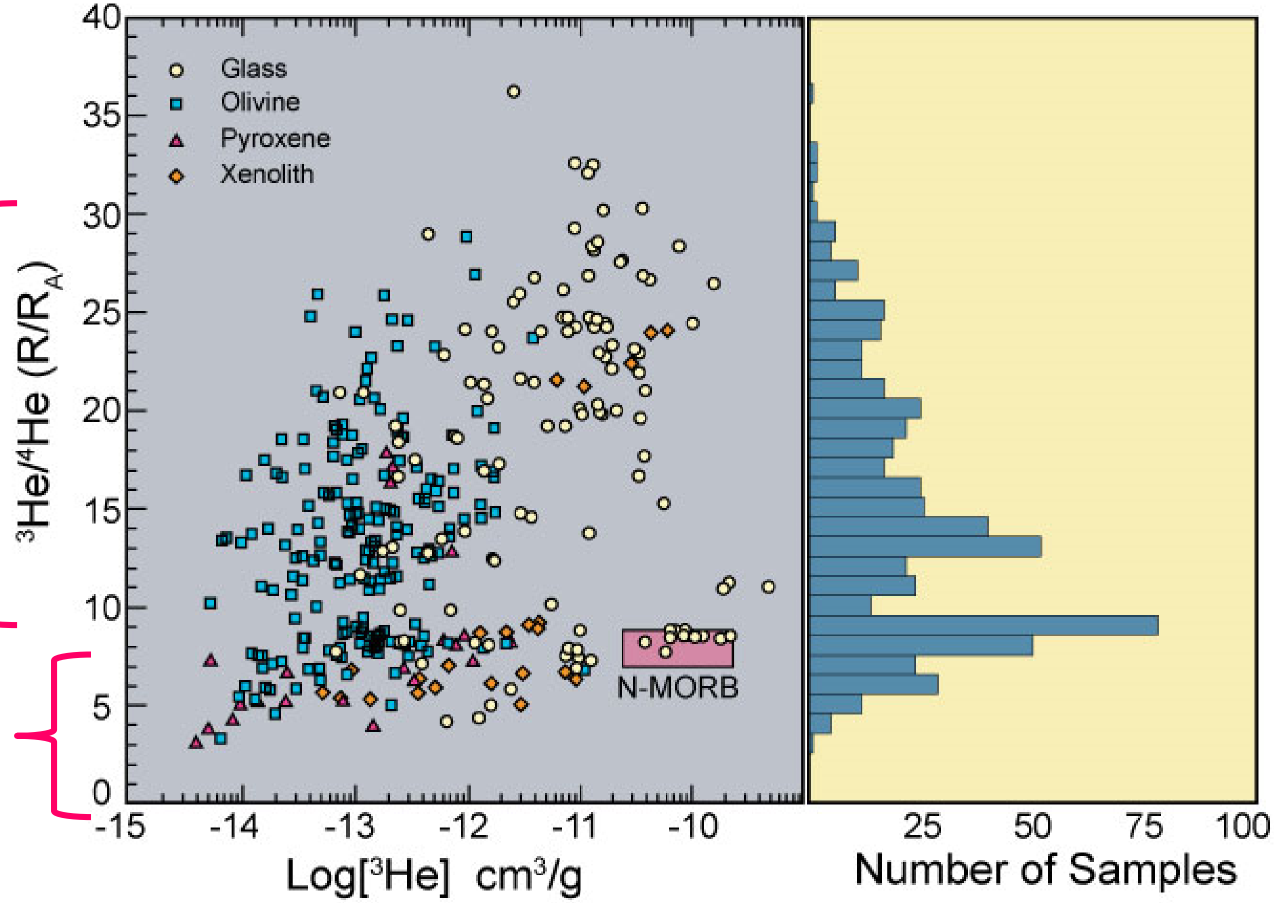



**Figure 14.12** $^3He/^4He$ isotope ratios in ocean island basalts and their relation to He concentration. Concentrations of $^3He$ are in $cm^3$ at 1 atm and 298K.After Sarda and Graham (1990) and Farley and Neroda (1998).

# He Isotopes

PHEM (**p**rimitive **he**lium **m**antle) is a hi-$^3He/^4He$ mantle end-member reservoir with near-primitive Sr-Nd-Pb characteristics.

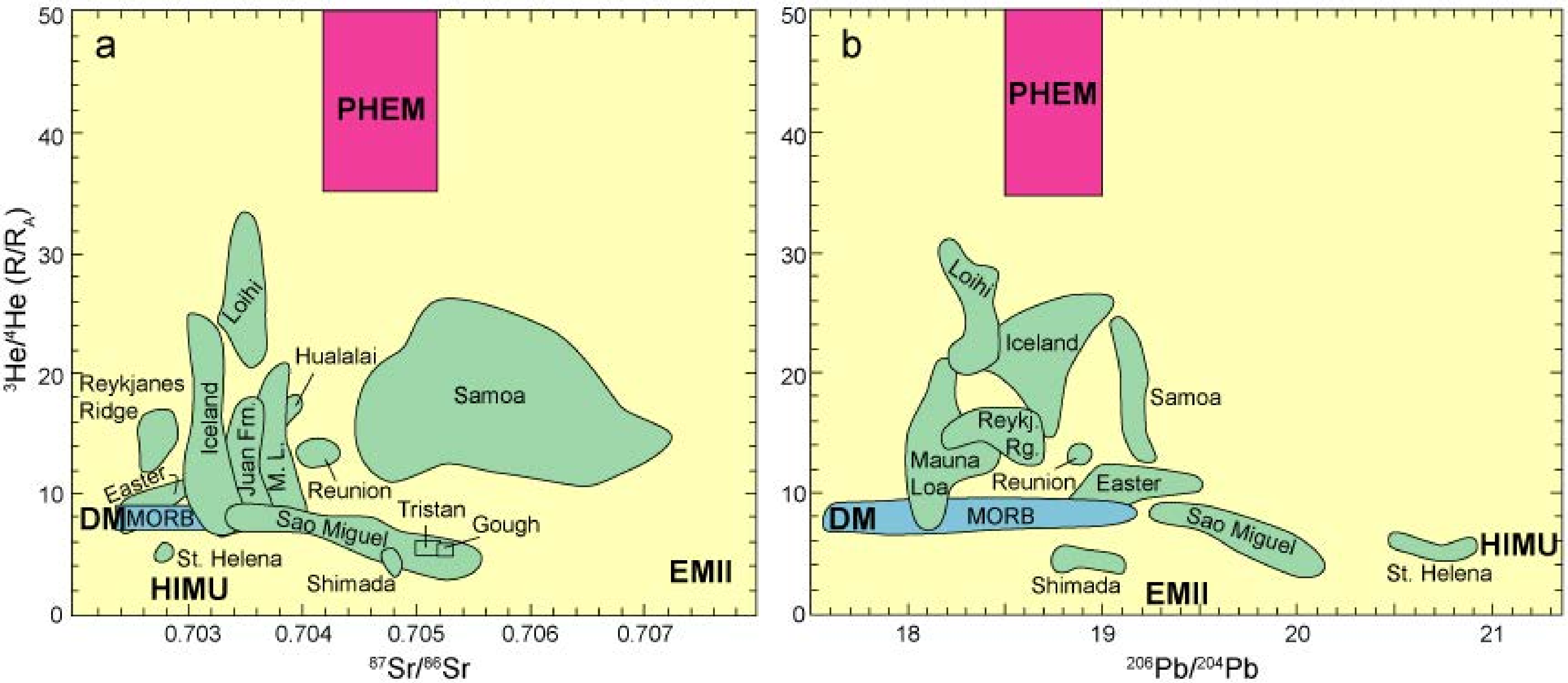



**Figure 14.13** $^3He/^4He$ vs. **a.** $^{87}Sr/^{86}Sr$ and **b.** $^{206}Pb/^{204}Pb$ for several OIB localities and MORB. The spread in the diagrams are most simply explained by mixing between four mantle components: DM, EMII, HIMU, and PHEM. After Farley et al. (1992).

# He Isotopes

Summary:

Shallow mantle MORB source is relatively homogeneous and depleted in He

Deeper mantle has more primordial (high) $^3He/^4He$, but still degassed and less than primordial (100-200$R_A$) values

PHEM may be that more primitive reservoir

Low $^3He/^4He$ may be due to recycled crustal U and Th

# Re/Os system and Os Isotopes

$^{187}Re \rightarrow {}^{187}Os$

Both are platinum group elements (PGEs) and highly siderophile (→ core or sulfides)

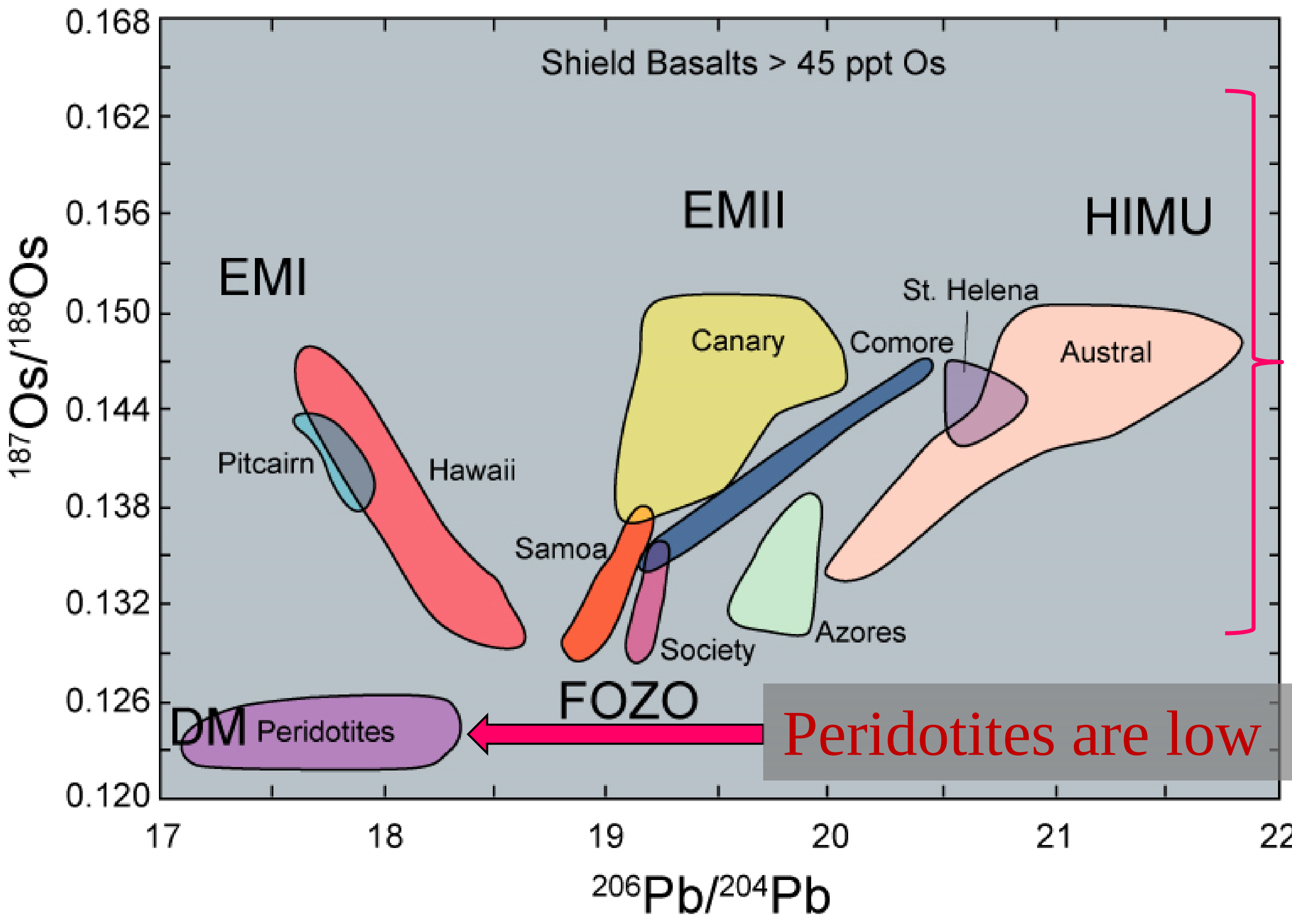
Mantle values of ($^{187}Os/^{188}Os$) are near chondritic (~0.13)

Os is compatible during mantle partial melting (→ trace sulfides), but Re is moderately incompatible (→ melts and silicates)

The mantle is thus enriched in Os relative to crustal rocks and crustal rocks (higher Re and lower Os) develop a high ($^{187}Os/^{188}Os$) which should show up if crustal rocks are recycled back into the mantle.

# Re/Os system and Os Isotopes

All of the basalt provinces are enriched in $^{187}Os$ over the values in mantle peridotites and require more than one $^{187}Os$-enriched reservoir to explain the distribution.




Volcanics are high (with little overlap to peridotites). High Re indicates a **crustal** component.

Peridotites are low (low Re)

**Figure 14.13** $^{187}Os/^{188}Os$ vs. $^{206}Pb/^{204}Pb$ for mantle peridotites and several oceanic basalt provinces. Os values for the various mantle isotopic reservoirs are estimates. After Hauri (2002) and van Keken et al. (2002b).

# Other Mantle Reservoirs

FOZO (focal zone): another "convergence" reservoir toward which many trends approach. Thus perhaps a common mixing end-member

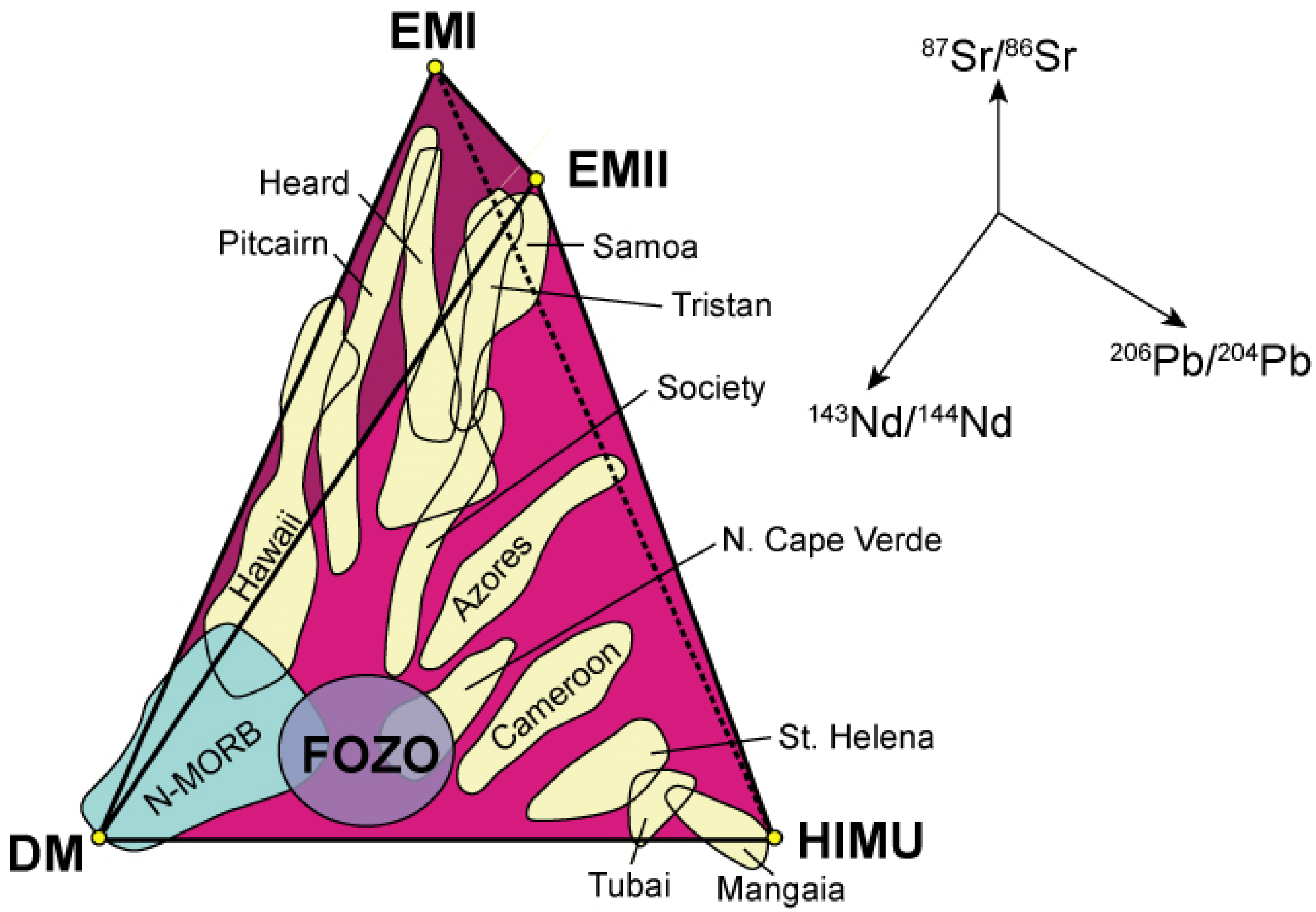



**Figure 14.15.** After Hart et al., 1992).

# Mantle Reservoirs

**Table 14-5**. Approximate Isotopic Ratios of Various Reservoirs.

| Reservoir | $^{87}Sr/^{86}Sr$ | $^{144}Nd/^{143}Nd$ | $^{206}Pb/^{204}Pb$ | $^{207}Pb/^{204}Pb$ | $^{3}He/^{4}He$ | $^{187}Os/^{186}Os$ | $^{18}O/^{16}O$ |
|---|---|---|---|---|---|---|---|
| End-Member Mantle | | | | | | | |
| **DM** | 0.7015 - 0.7025 | 0.5133 - 0.5136 | 15.5 - 17.7 | < 15.45 | 7 - 9 $R_A$ | 0.123 - 0.126 | low |
| **HIMU** | 0.7025 - 0.7035 | 0.511 - 0.5121 | 21.2 - 21.7 | 15.8 - 15.9 | 2 - 6 $R_A$? | 0.15 | low |
| **EMI** | c 0.705 | < 0.5112 | 17.6 - 17.7 | 15.46 - 15.49 | 2 - 6 $R_A$? | 0.152 | low |
| **EMII** | > 0.722 | 0.511 - 0.512 | 16.3 - 17.3 | 15.4 - 15.5 | 2 - 6 $R_A$? | 0.156? | high |
| Other Mantle | | | | | | | |
| **BSE** | 0.7052 | 0.51264 | 18.4 | 15.58 | 40 - 80 $R_A$? | 0.129 | low |
| **PHEM** | 0.7042 - 0.7052 | 0.51265 - 0.51280 | 18.4 - 19.0 | 15.5 - 15.6 | > 35 $R_A$ | low | low |
| **PREMA** | 0.7033 | < 0.5128 | 18.2 - 18.5 | 15.4 - 15.5 | | | |
| **FOZO** | 0.7028 - 0.7034 | 0.51287 - 0.51301 | 19.5 - 20.5 | 15.55 - 15.71 | 8 - 32 $R_A$ | low | low |
| **C** | 0.703 - 0.704 | 0.5128 - 0.5129 | 19.2 - 19.8 | 15.55 - 15.65 | 20 - 25+? | low | low |
| **Continental Crust** | 0.72 - 0.74 | 0.507 - 0.513 | up to 28 | up to 20 | ~ 0.1 $R_A$ | high | high |

DM, HIMU, EMI, EMII, BSE from Rollinson (1993) p. 233-236, PHEM from Farley et al. (1992), FOZO from Hauri et al. (1994) and Stracke et al. (2005), C from Hanan and Graham (1996), Os values from Shirey and Walker (1998) and van Keken et al. (2002), O estimates based on Eiler (2001).

- EMI, EMII, and HIMU: too enriched for any known *mantle* process...must correspond to crustal rocks and/or sediments
- EMI
  - Slightly enriched
  - Deeper continental crust or oceanic crust
- EMII
  - More enriched
  - Specially in $^{87}Sr$ (Rb parent) and Pb (U/Th parents)
  - Upper continental crust or ocean-island crust
- If the EM and HIMU = continental crust (or older oceanic crust and sediments), only → deeper mantle by subduction and recycling
- To remain isotopically distinct: could not have rehomogenized or re-equilibrated with rest of mantle

# The Nature of the Mantle

- N-MORBs involve shallow melting of passively rising upper mantle

  → a significant volume of depleted upper mantle (lost lithophile elements and considerable He and other noble gases).

- OIBs typically originate from deeper levels.

  Major- and trace-element data → the deep source of OIB magmas (both tholeiitic and alkaline) is distinct from that of N-MORB.

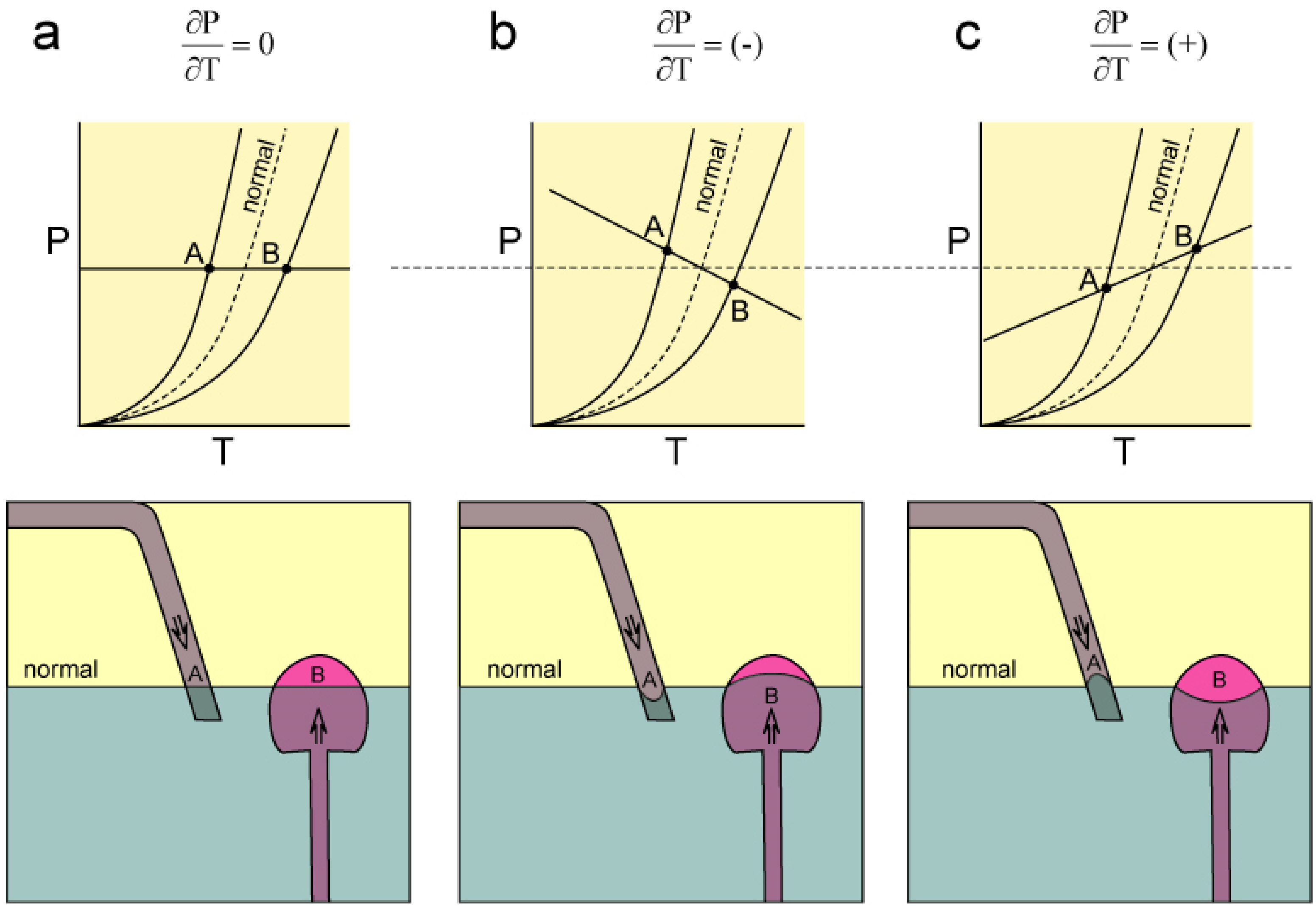
  Trace element and isotopic data reinforce this notion and further indicate that the **deeper mantle is relatively heterogeneous and complex**, consisting of several domains of contrasting composition and origin. In addition to the depleted MORB mantle, there are at least four enriched components, including one or **more containing recycled crustal and/or sedimentary material reintroduced into the mantle by subduction**, and at least one (FOZO, PHEM, or C) that retains much of its primordial noble gases.

- MORBs are not as homogenous as originally thought, and exhibit most of the compositional variability of OIBs, although the variation is expressed in far more subordinate proportions. **This implies that the shallow depleted mantle also contains some enriched components.**

# The Nature of the Mantle

- So is the mantle layered (shallow depleted and deeper non-depleted and even enriched)?
- Or are the enriched components stirred into the entire mantle (like fudge ripple ice cream)?
- How effective is the 660-km transition at impeding convective stirring??

- This depends on the Clapeyron slope of the phase transformation at the boundary!




No Effect

Retards Penetration
→ 2-Layer Mantle Model more likely

Enhances Penetration
→ Whole-Mantle mixing more likely

**Figure 14.16.** Effectiveness of the 660-km transition in preventing penetration of a subducting slab or a rising plume

# Mantle dynamics

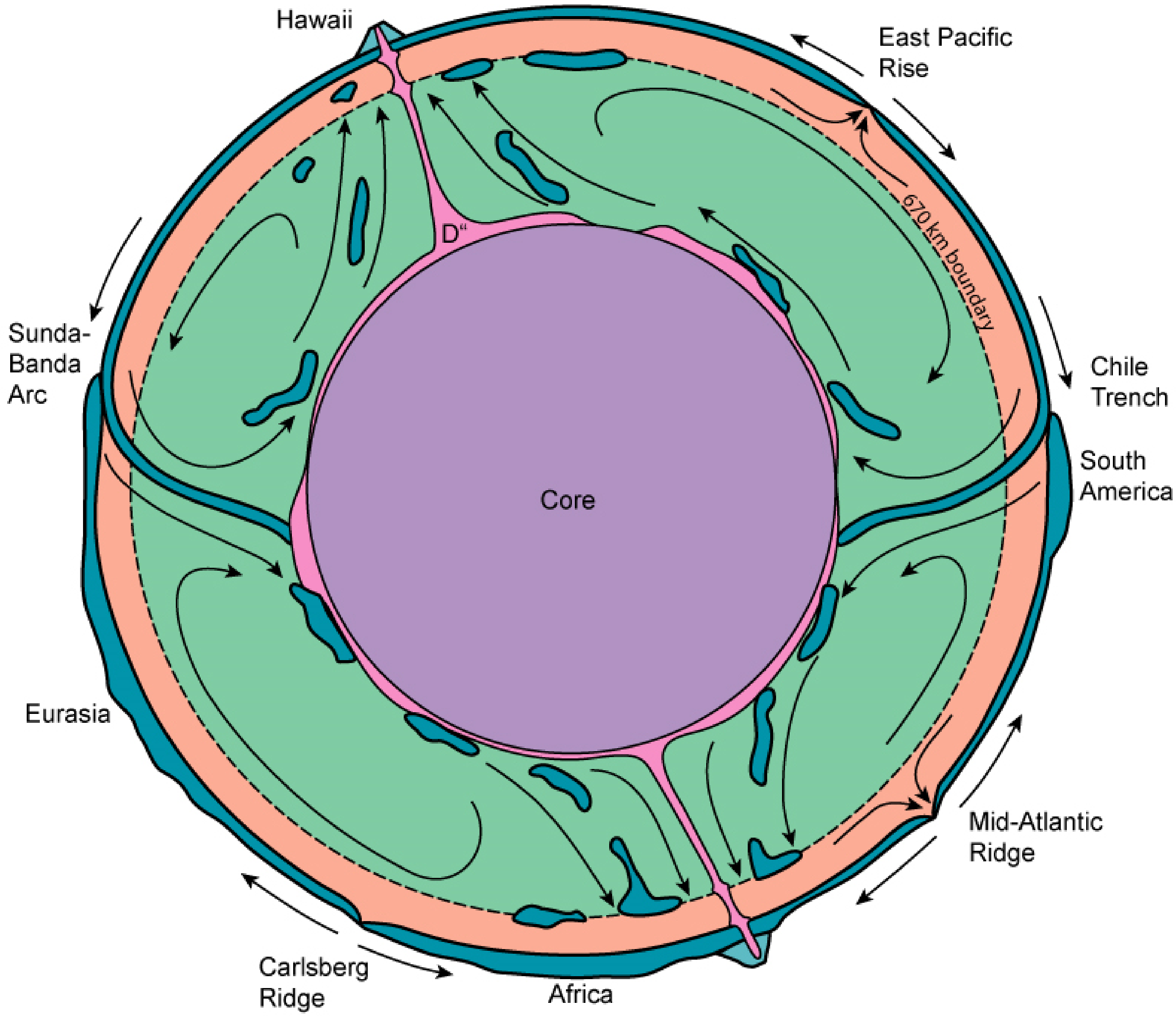



**Figure 1.14. Schematic diagram of a 2-layer dynamic mantle model in which the 660 km transition is a sufficient density barrier to separate lower mantle convection (arrows represent flow patterns) from upper mantle flow, largely a response to plate separation.** The only significant things that can penetrate this barrier are vigorous rising hotspot plumes and subducted lithosphere (which sink to become incorporated in the D'' layer where they may be heated by the core and return as plumes). After Silver et al. (1988).

# Mantle dynamics

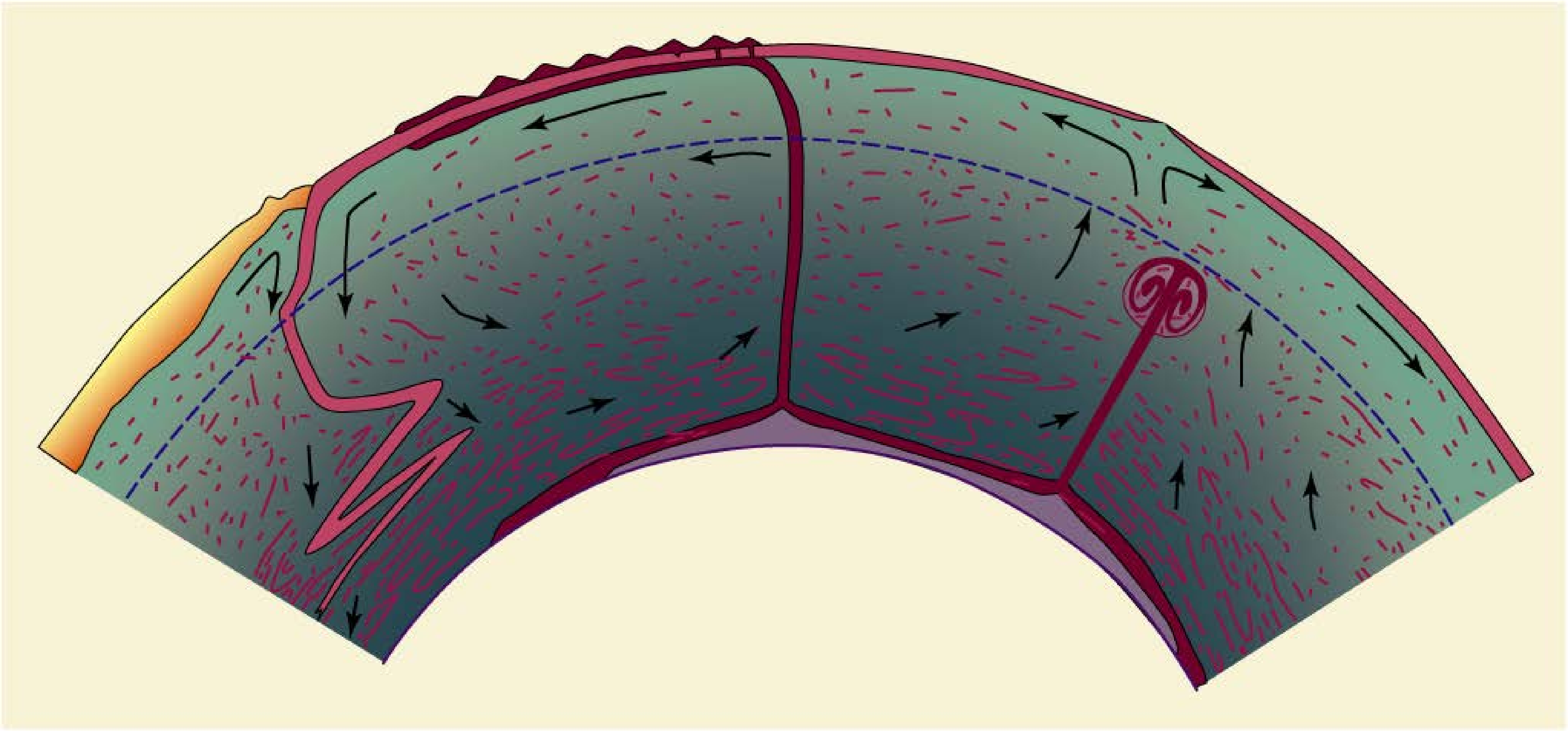


Figure14.17. **Whole-mantle convection model with geochemical heterogeneity preserved as blobs of fertile mantle in a host of depleted mantle**. Higher density of the blobs results in their concentration in the lower mantle where they may be tapped by deep-seated plumes, probably rising from a discontinuous D" layer of dense “dregs” at the base of the mantle. After Davies (1984) .

# Mantle dynamics

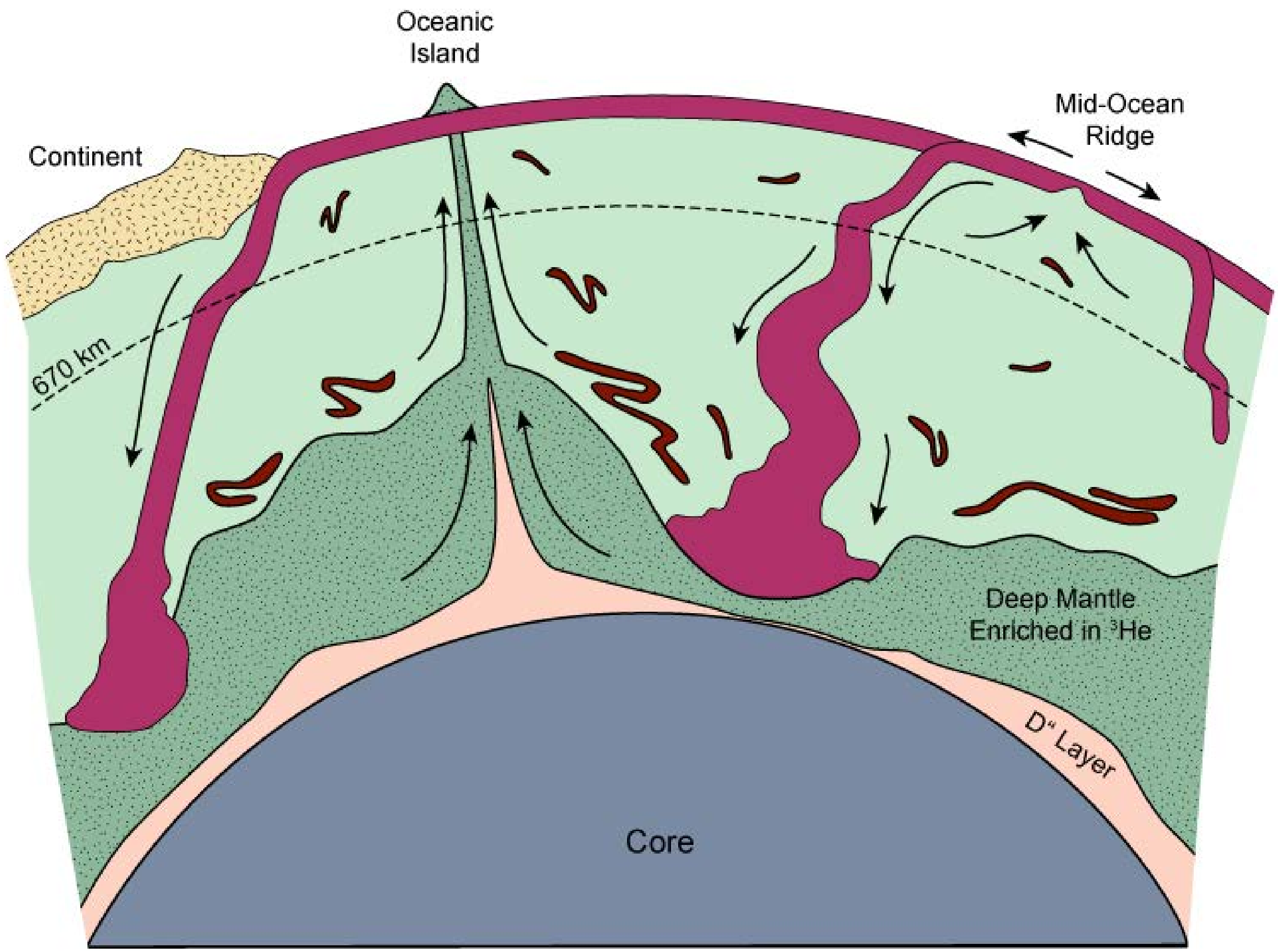
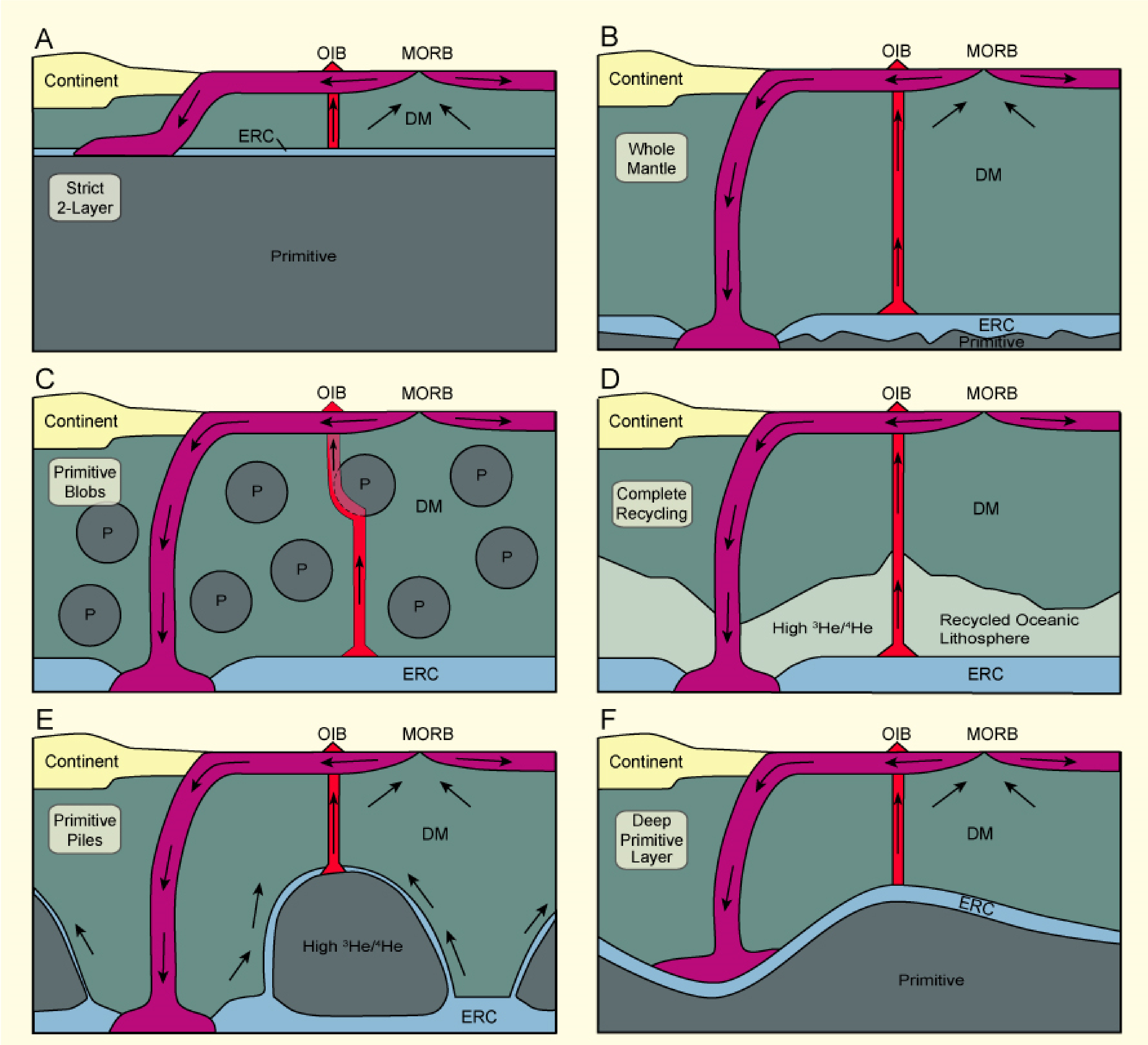



Figure14.18. **2-layer mantle model with a dense layer in the lower mantle with less depletion in lithophile elements and noble gases. The top of the layer varies in depth from ~ 1600 km to near the core-mantle boundary.** After Kellogg et al. (1999).




**Various mantle convection models.**

After Tackley (2000). *Mantle Convection and Plate Tectonics: Toward an Integrated Physical and Chemical Theory*. Science, **288**, 2002-2007.

# A Model for Oceanic Magmatism

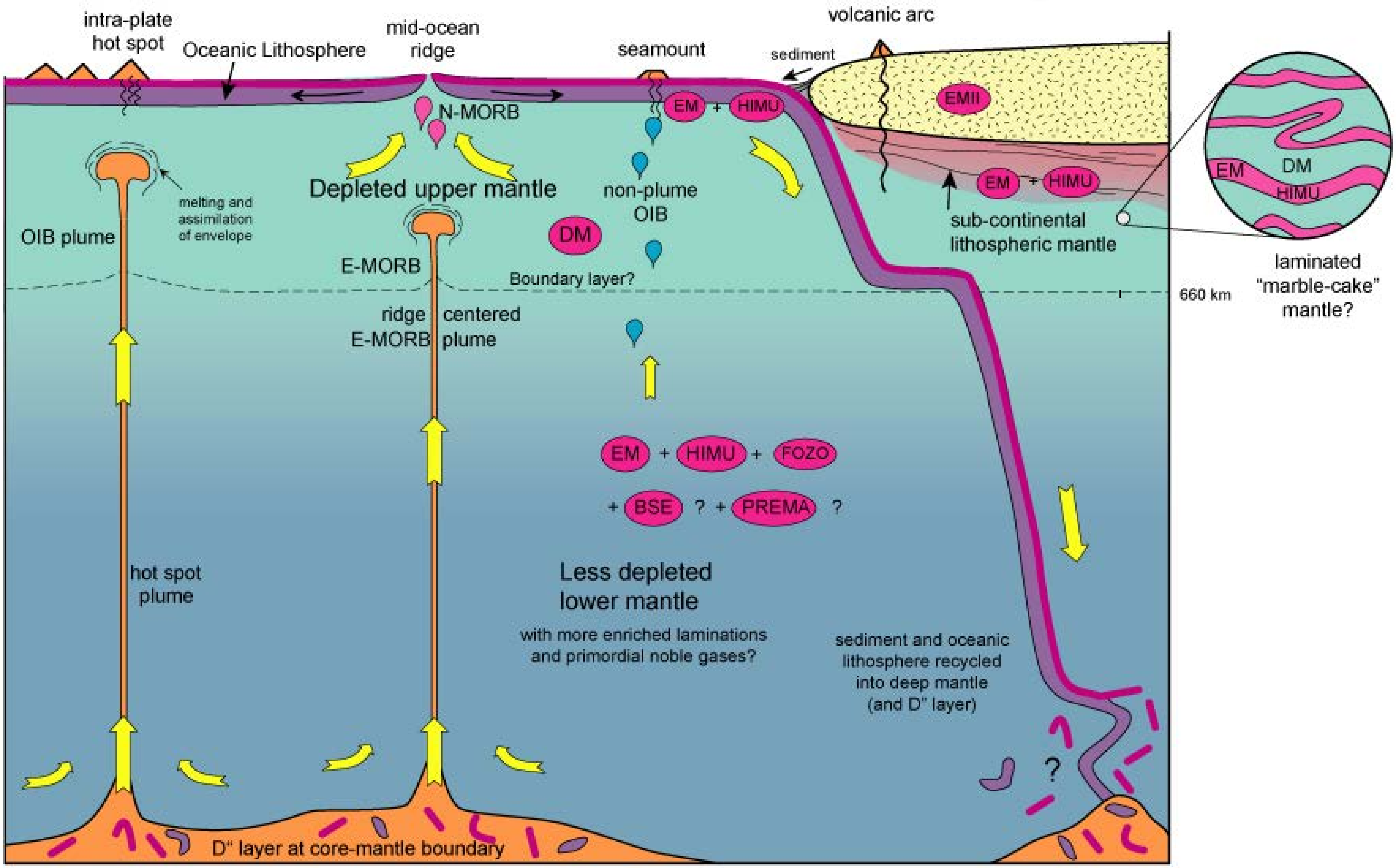



**Figure 14.19.** Schematic model for oceanic volcanism. Nomenclature from Zindler and Hart (1986) and Hart and Zindler (1989).

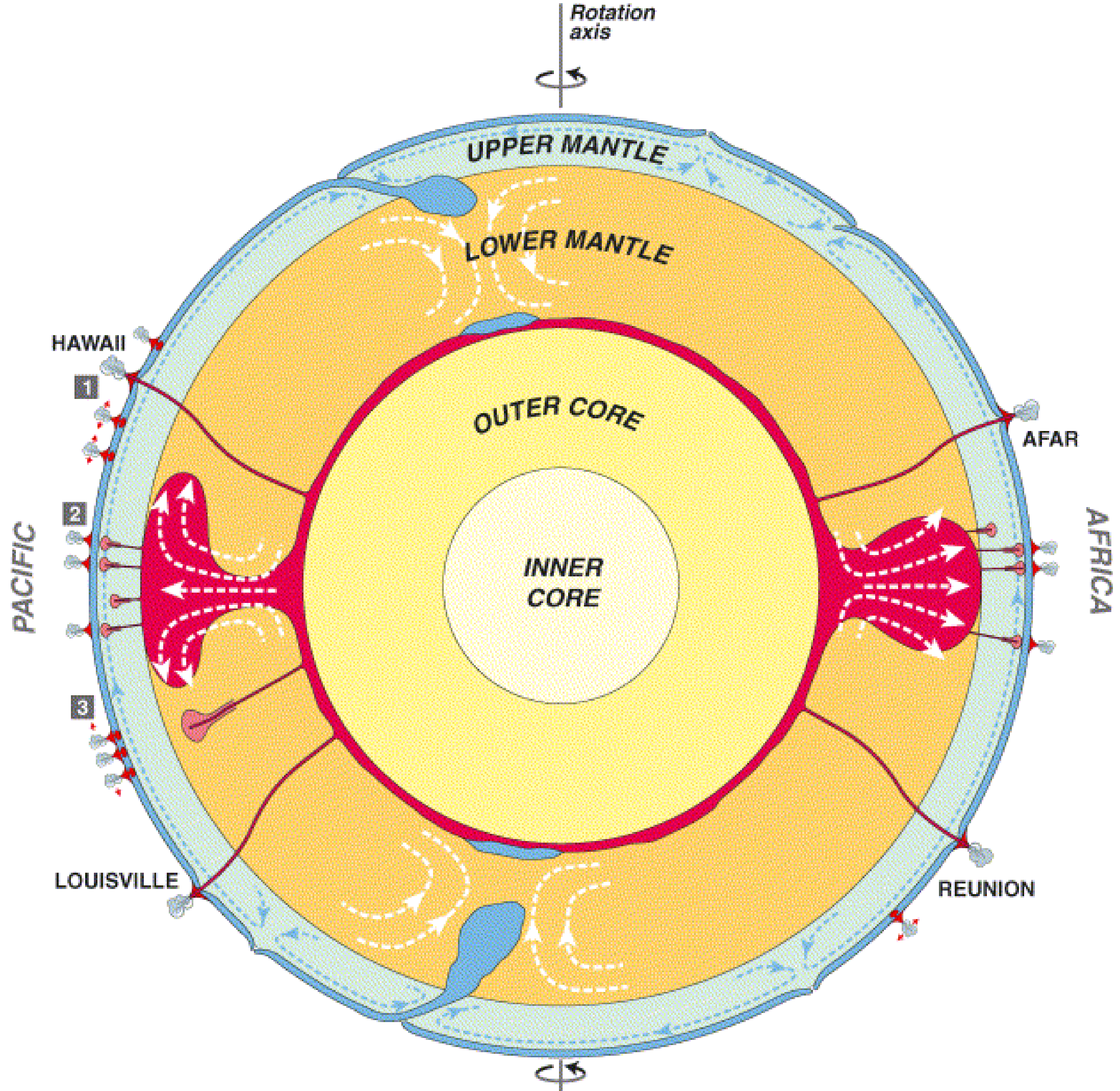



**Figure 14.23** A schematic cross-section through the Earth showing the three types of plumes/hotspots proposed by Courtillot et al. (2003). “Primary” plumes, such as Hawaii, Afar, Reunion, and Louisville are deep-seated, rising from the D'' layer at the core-mantle boundary to the surface. “Superplumes” or “superswells” are broader and less concentrated, and stall at the 660-km transition zone where the spawn a series of “secondary” plumes. “Tertiary” hotspots have a superficial origin. From Courtillot et al. (2003).