



# Plate tectonics and related magmatic processes

A global perspective

*Jörg A. Pfänder*



# Overall goal

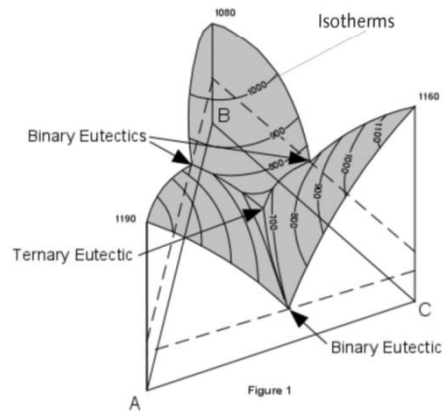
Reconstruct the **tectonic setting** and **geodynamic environment** in which a piece of **rock** formed in the geological past

Prerequisite: **ACTUALISM**



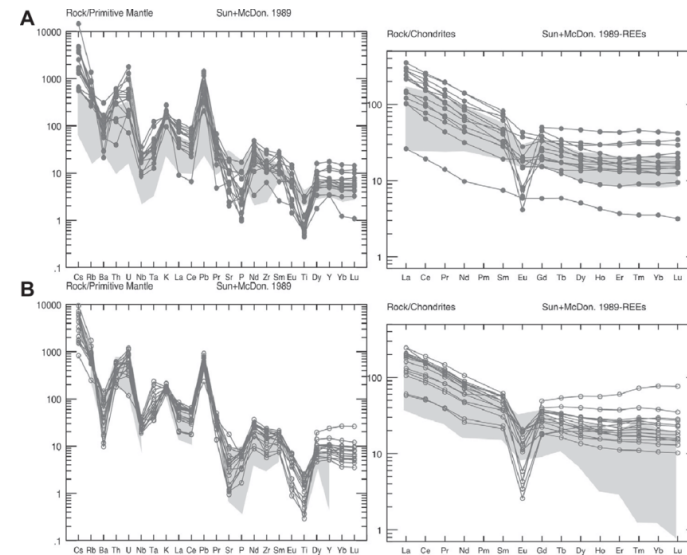
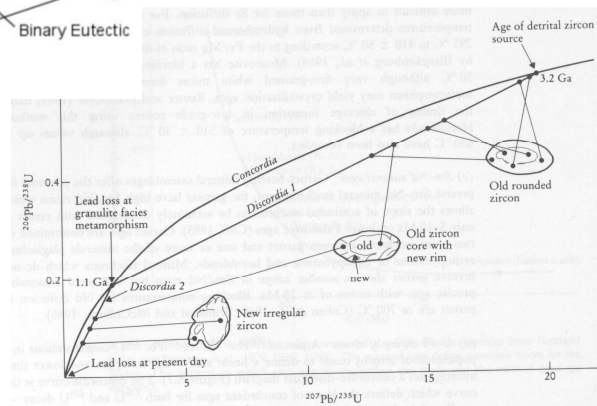
# Tools

## Geochemistry and Petrology



$$C(X) = \frac{1}{X} \int_0^X c(X') dX'$$

$$c = \frac{d}{dX}(XC)$$



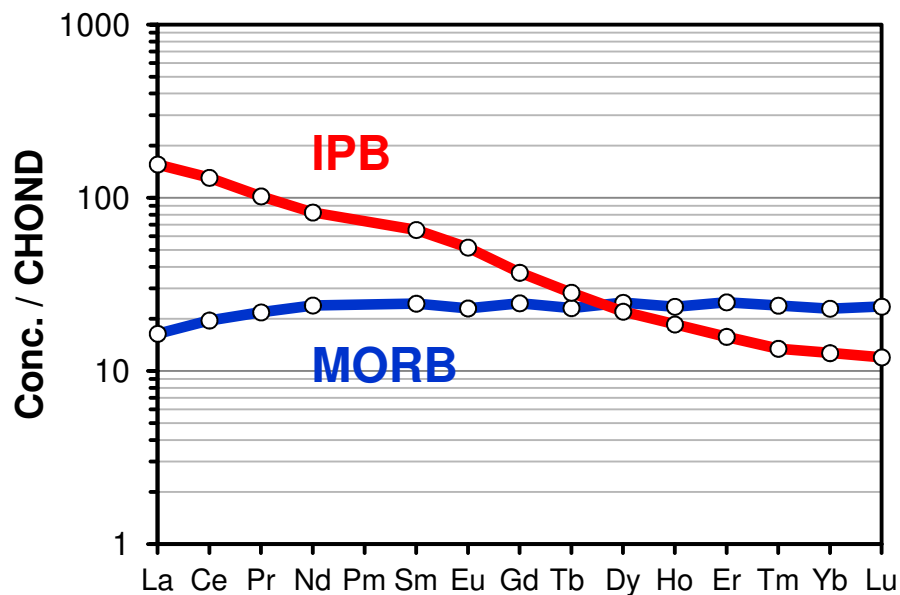


Die Ressourcenuniversität. Seit 1765.

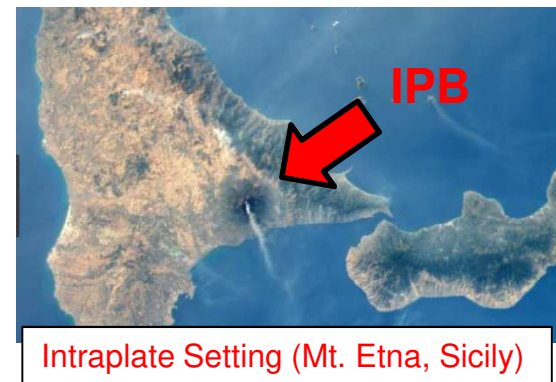
# Example



## Two basalts: „BLACK ROCKS“



Rare Earth Element Plot of a typical Mid Ocean Ridge Basalt and a typical Intraplate Basalt

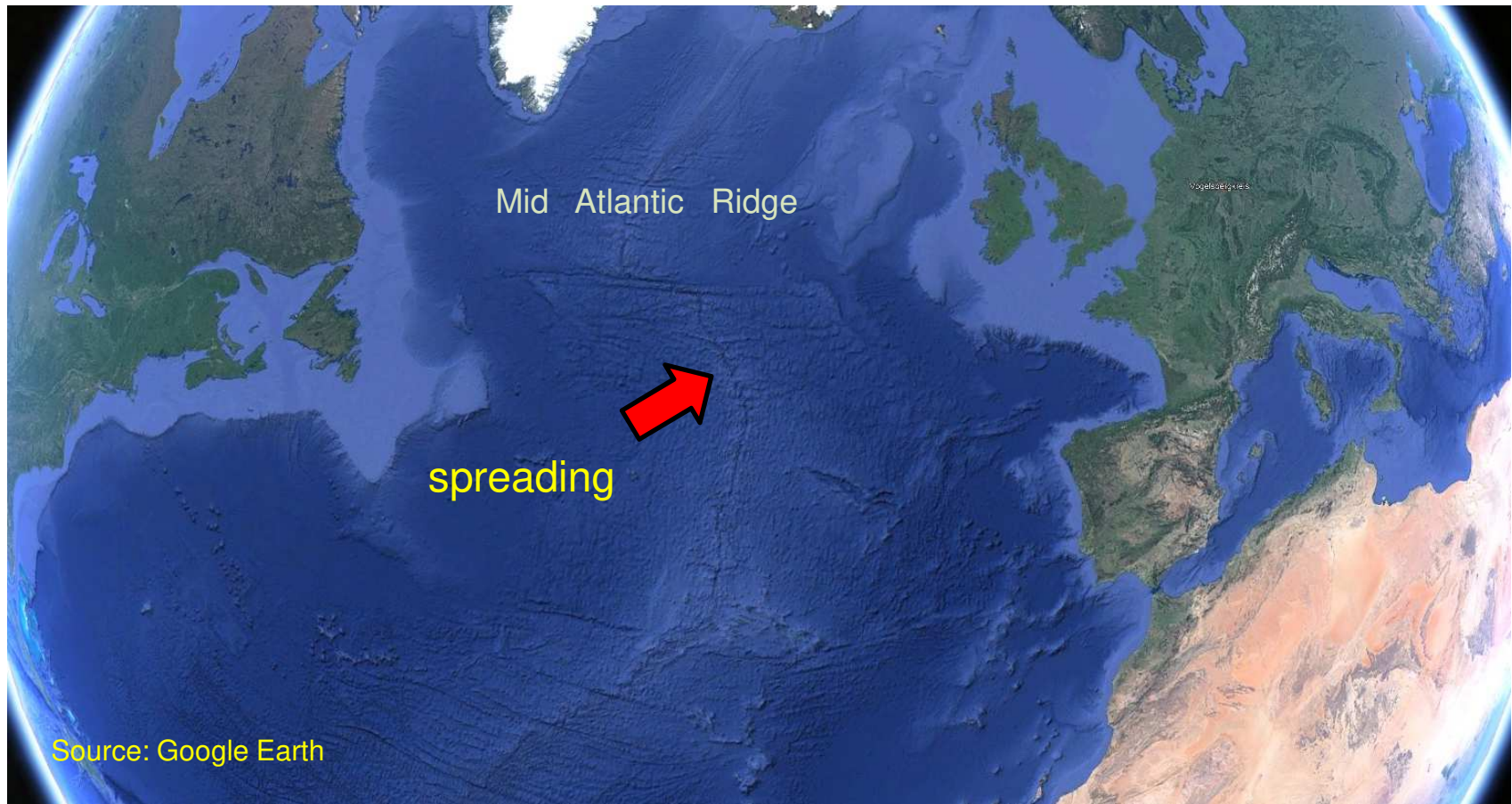




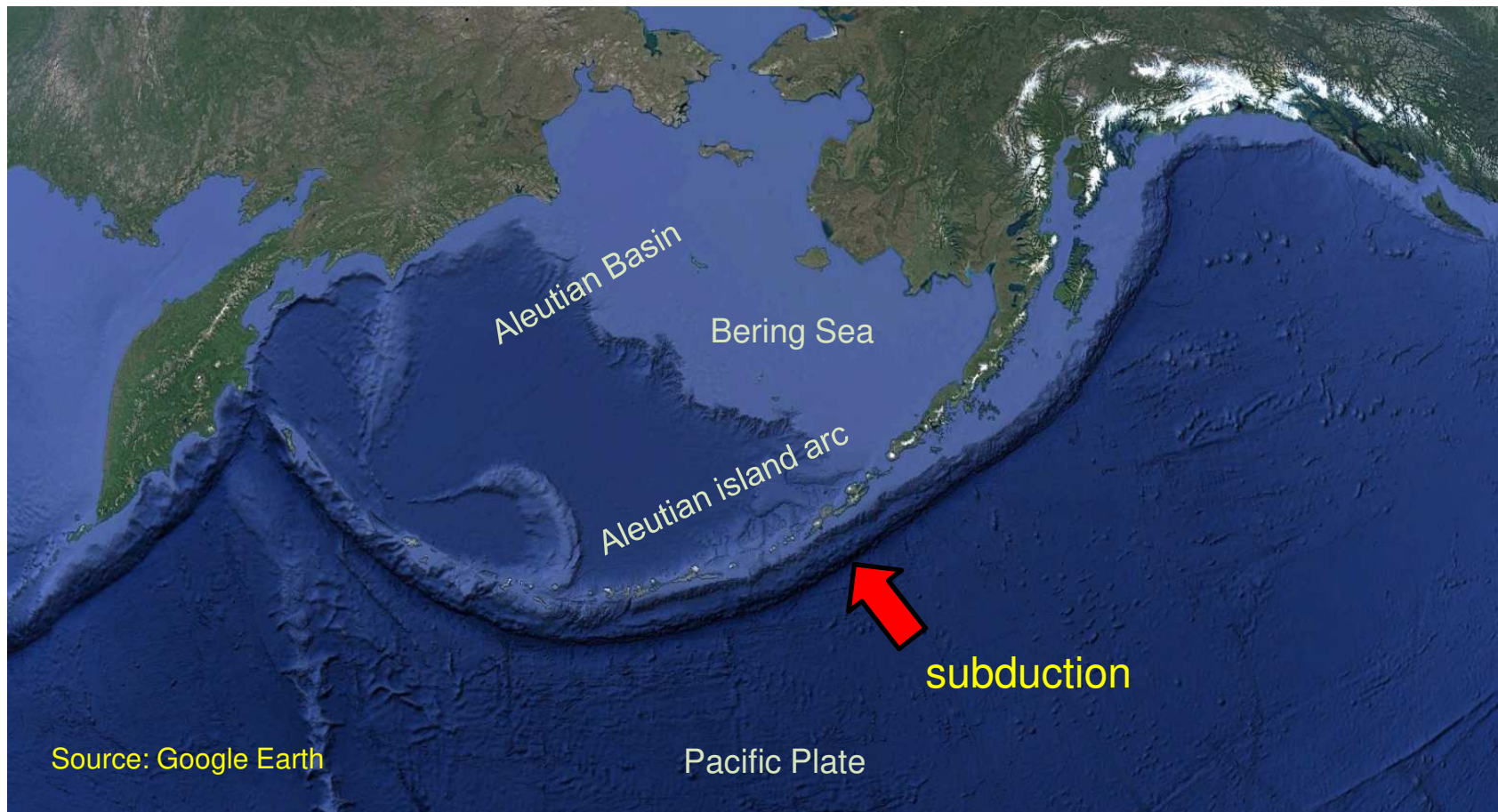
# Tectonic settings

- Mid ocean ridges
  - Oceanic island arcs
  - Active continental margins
  - Back-arc settings
  - Oceanic intraplate settings
  - Continental intraplate settings
  - Convergent plate margins
- Subduction zone settings**
- Intraplate settings**

# Mid ocean ridge



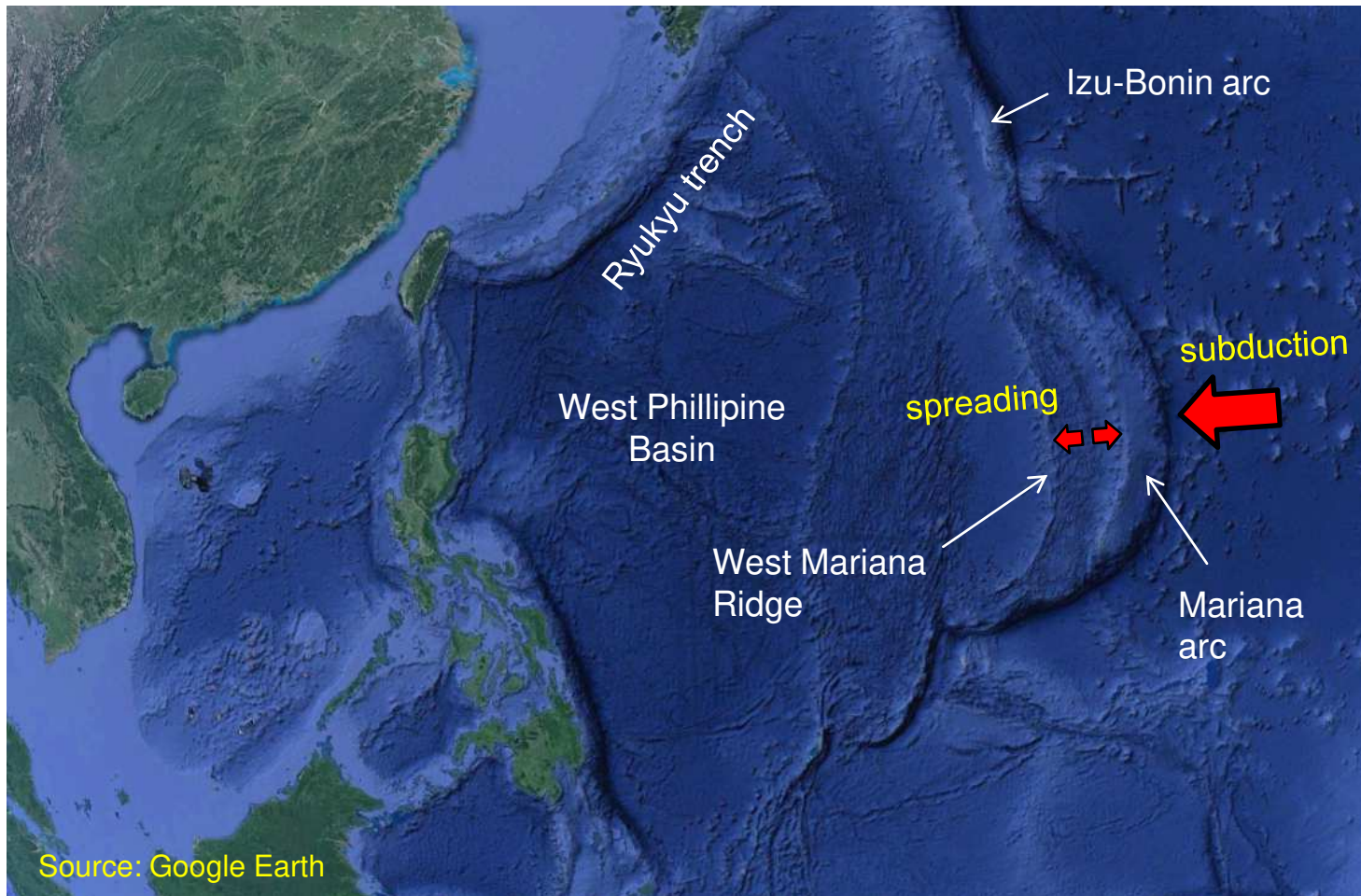
# Oceanic island arc



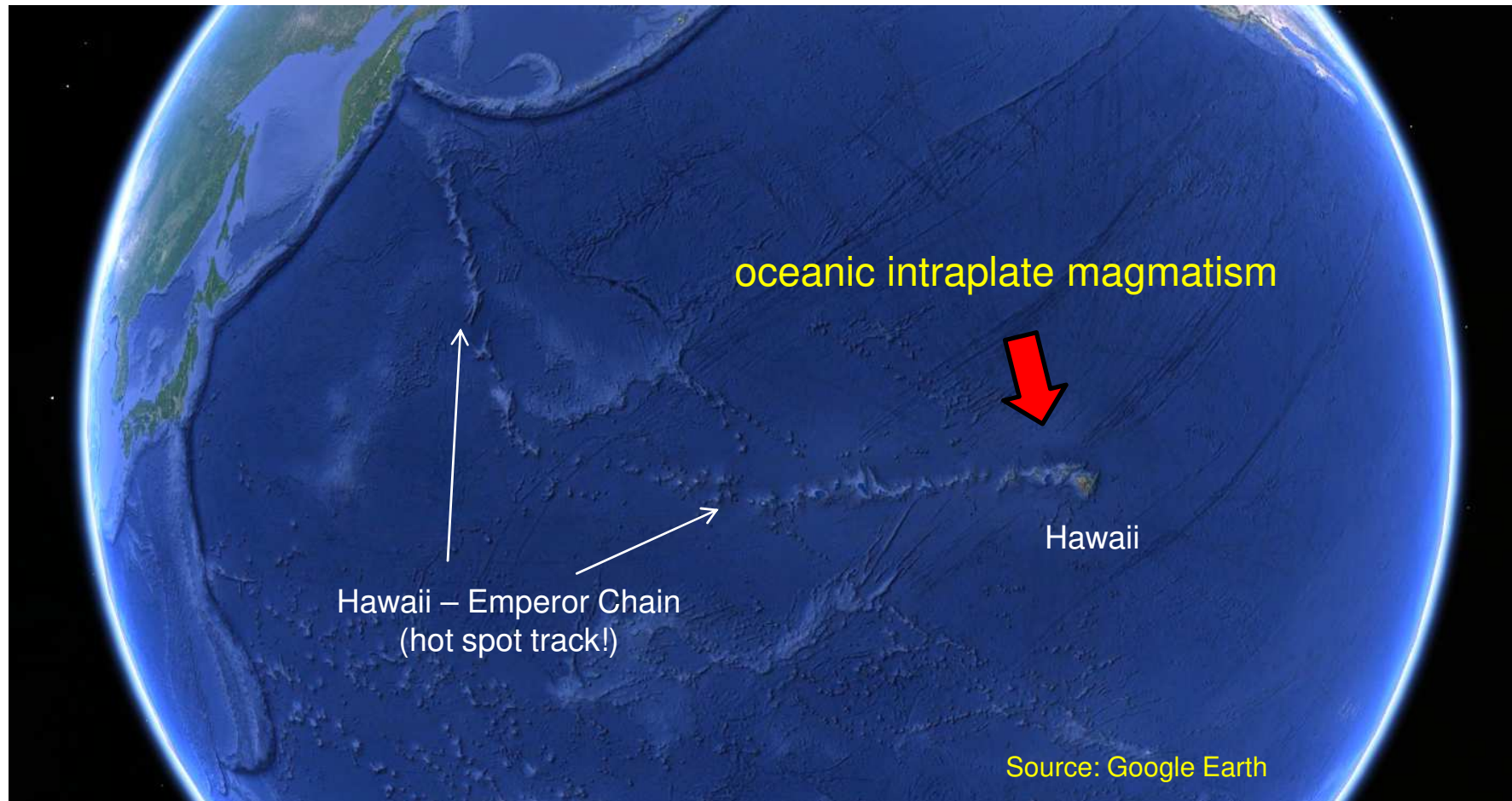
# Active continental margin



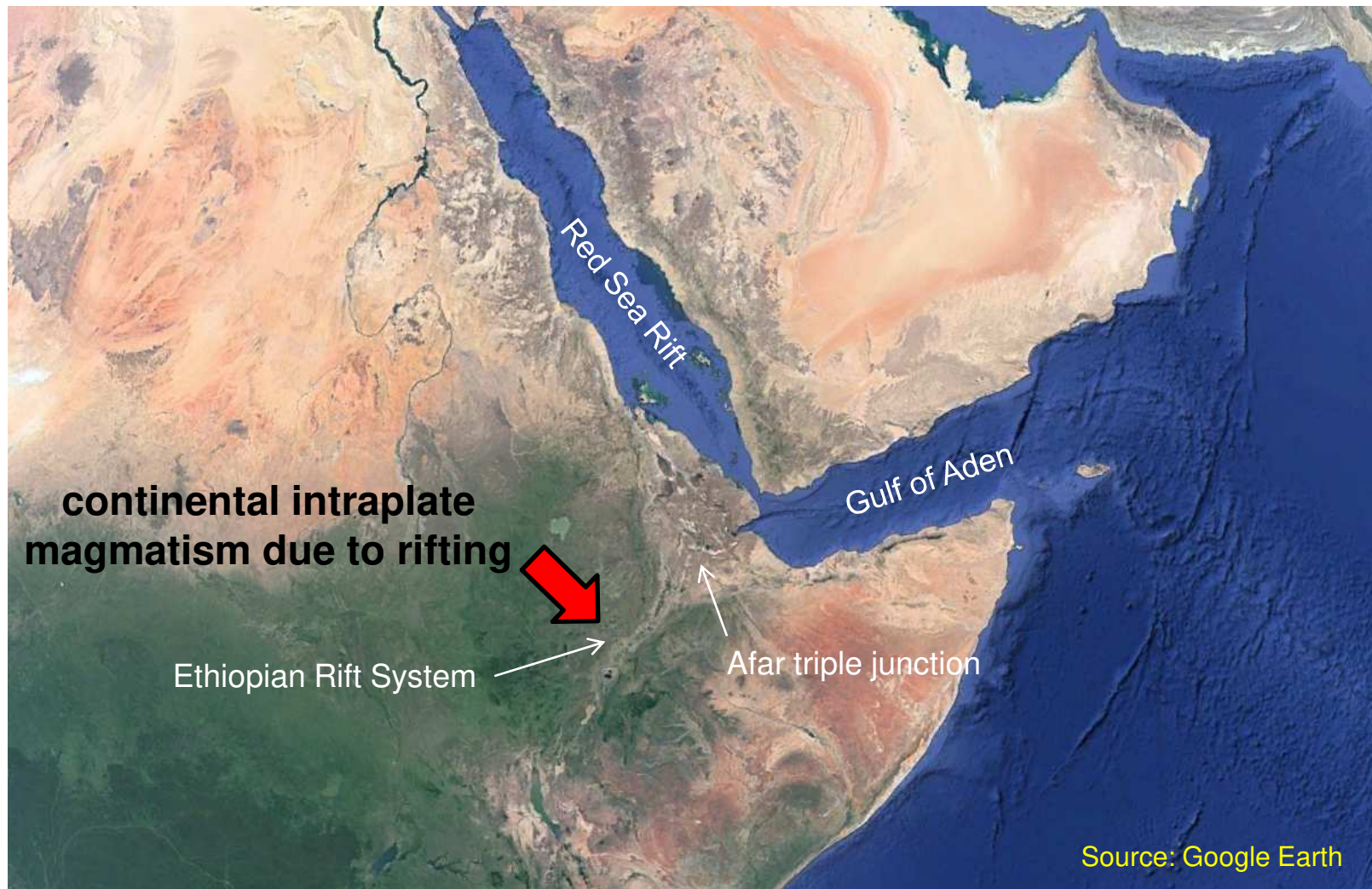
# Back arc spreading



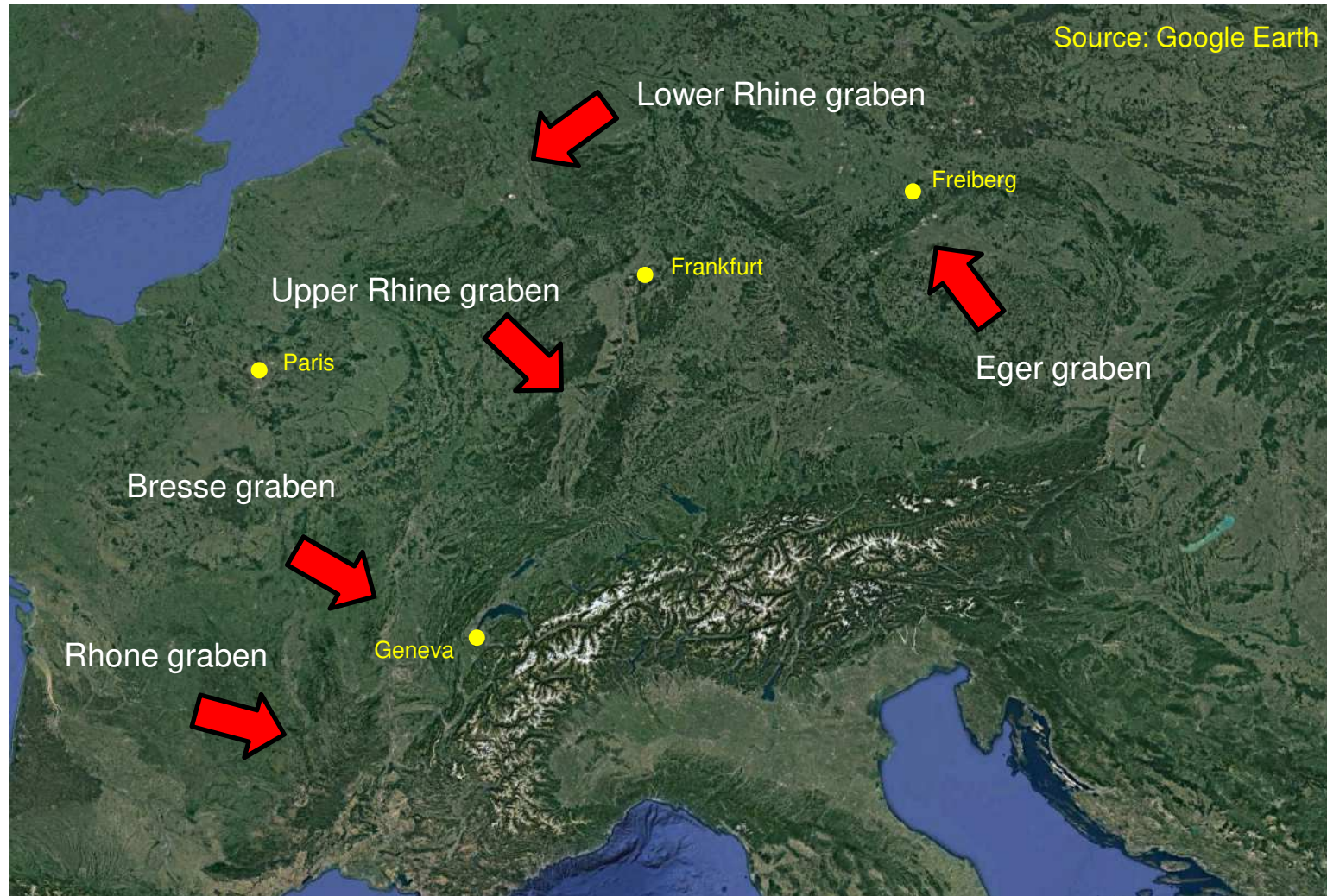
# Oceanic intraplate setting



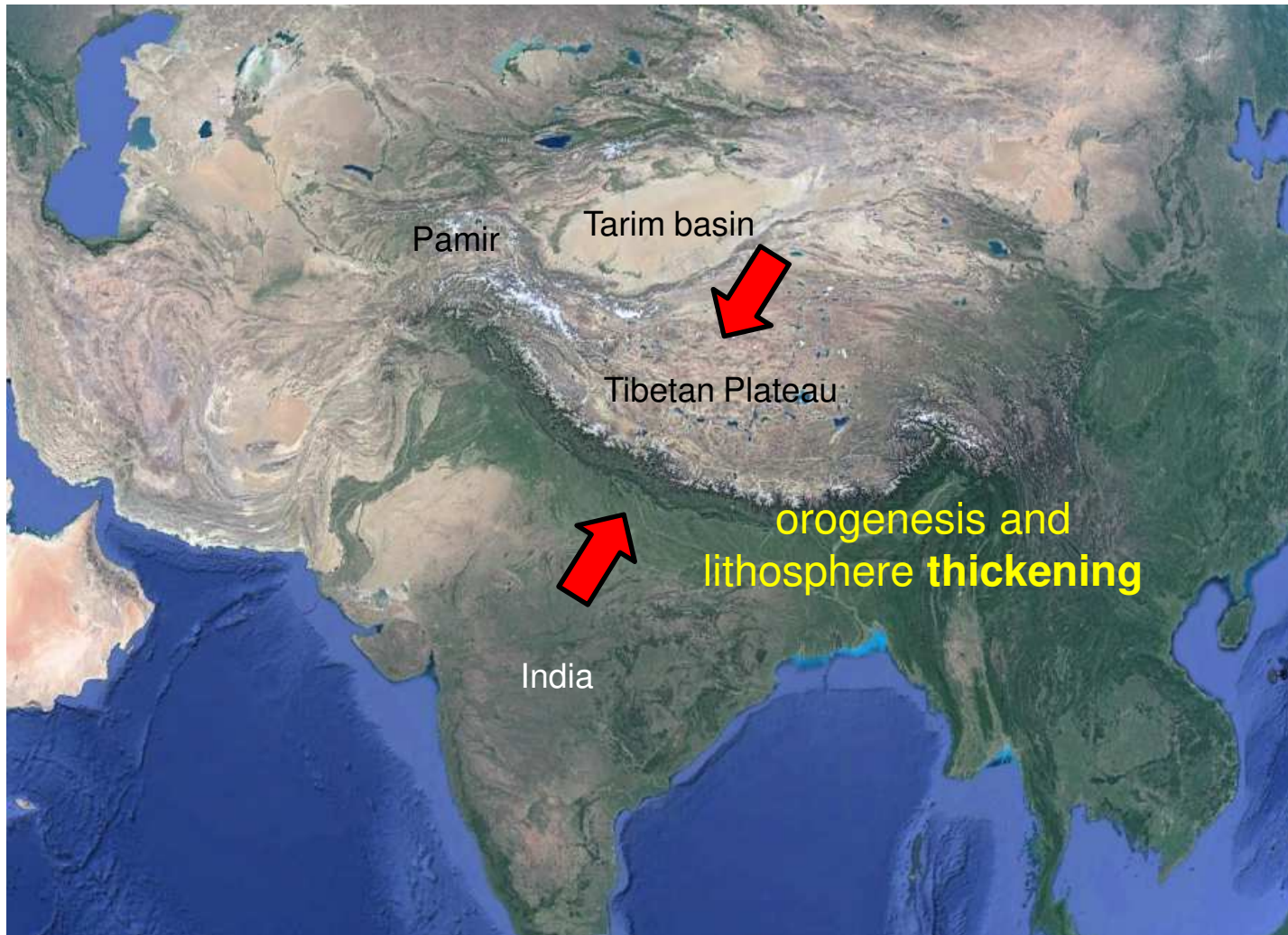
# Continental intraplate setting



# Continental intraplate setting



# Convergent plate margins

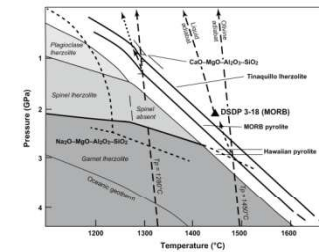
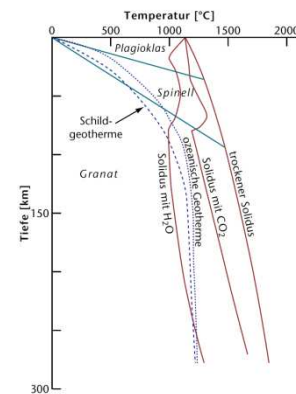
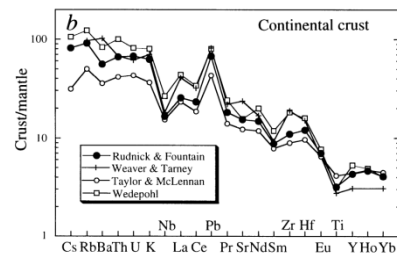


## Chapter 1

# BASICS

$$C(X) = \frac{1}{X} \int_0^X c(X') dX'$$

$$c = \frac{d}{dX}(XC)$$



# From rocks to geochemical (and geochronological) data



**ROCK**

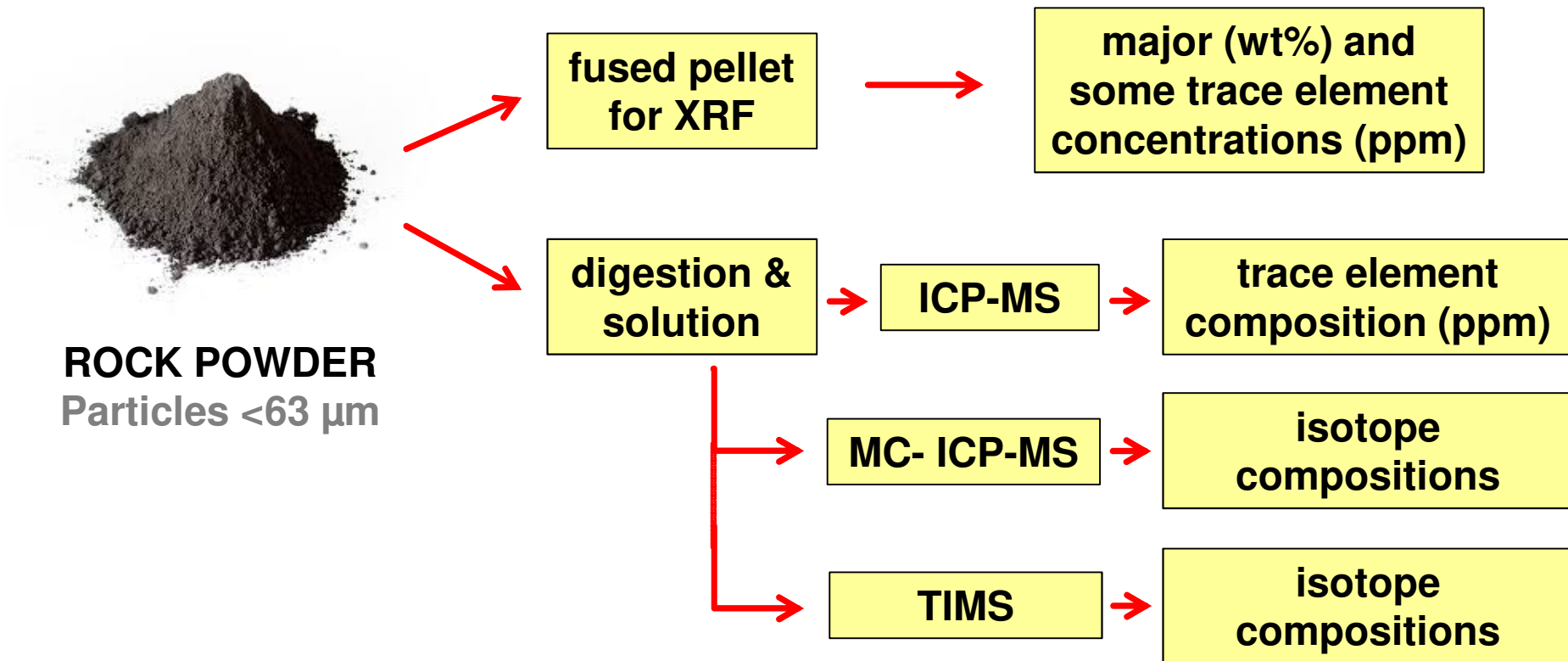
Agate ball mill  
or agate disc mill



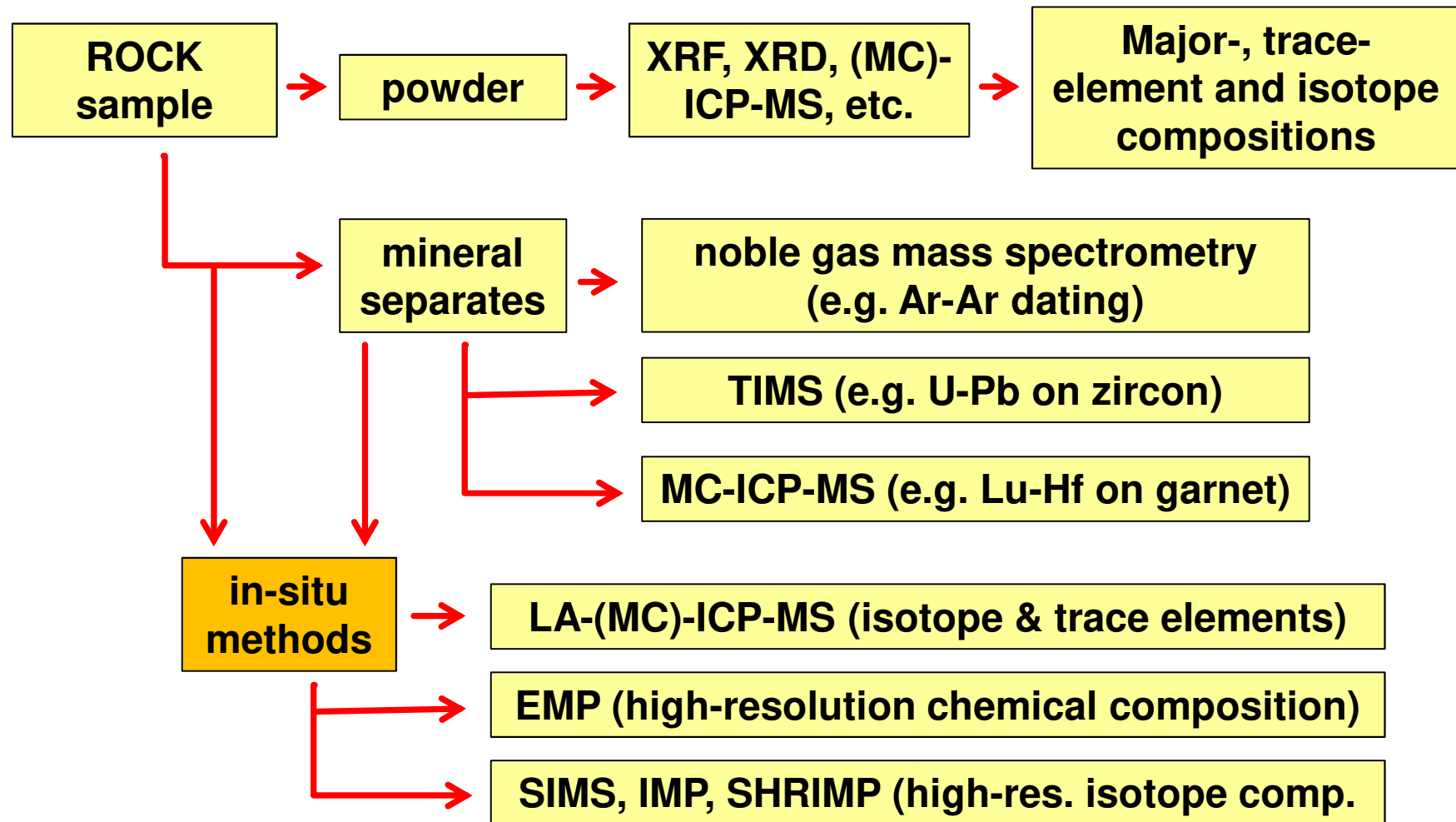
**ROCK POWDER**  
Particles <63  $\mu\text{m}$



# From rocks to geochemical (and geochronological) data



## More data acquisition methods...



# Data documentation I

**Table 1:** Major element, trace element and isotope data of SiO<sub>2</sub>-undersaturated rocks from the Heldburg region

Sample Rock type	Method	S097511 Mel	S097512 Mel	S097519 Mel	S107048 Mel	S107052 Mel	S097513 Neph	S097514 Neph	S097515 Neph	S107045 Neph
SiO <sub>2</sub>	XRF	35.5	34.9	35.4	34.5	34.2	38.4	41.0	41.3	40.0
Al <sub>2</sub> O <sub>3</sub>	XRF	9.00	9.42	9.09	9.60	9.52	9.6	11.1	11.1	10.3
Fe <sub>2</sub> O <sub>3</sub>	XRF	12.9	12.2	11.8	13.4	11.6	12.0	12.6	13.1	13.3
FeO		11.6	10.9	10.6	12.0	10.4	10.8	11.4	11.8	11.9
MnO	XRF	0.20	0.18	0.17	0.18	0.17	0.17	0.20	0.20	0.20
MgO	XRF	13.3	12.8	14.0	13.3	14.1	13.6	12.3	12.6	13.7
CaO	XRF	12.0	11.6	12.5	11.6	12.7	11.2	11.8	11.3	11.9
Na <sub>2</sub> O	XRF	3.39	3.16	3.47	3.11	3.82	2.92	4.04	3.44	3.26
K <sub>2</sub> O	XRF	1.35	1.25	0.98	1.16	0.74	0.79	1.08	1.14	0.95
TiO <sub>2</sub>	XRF	2.85	2.84	2.67	2.90	2.53	2.80	2.84	2.94	2.89
P <sub>2</sub> O <sub>5</sub>	XRF	1.00	0.99	0.84	0.76	0.82	0.61	0.82	0.77	0.86
SO <sub>3</sub>	XRF	1.11	0.84	0.71	0.28	0.46	0.31	0.03	0.05	0.06
L.O.I	XRF	7.60	9.86	8.94	9.48	9.92	8.42	2.04	2.12	2.96
SUM	XRF	100.3	99.9	100.5	100.3	100.6	100.8	99.9	100.1	100.3
Li	ICPMS	36.9	31.6	45.2	15.5	54.6	20.0	41.5	21.1	27.2
Sc	ICPMS	20.7	20.1	22.3	21.9	26.2	21.3	21.9	24.8	24.3
V	XRF	265	265	228	272	246	262	279	281	289
Cr	XRF	494	458	409	367	539	451	476	527	643
Co	ICPMS	72.8	53.5	66.5	70.0	61.0	74.4	64.9	69.0	74.6
Ni	XRF	424	206	354	309	379	385	312	325	468
Cu	XRF	188	254	133	146	144	185	113	158	178
Zn	XRF	173	193	117	137	104	163	129	163	185
Ga	ICPMS	16.4	17.8	15.9	18.0	16.2	16.6	17.8	18.0	17.1
Rb	ICPMS	33.1	42.4	25.0	30.8	22.1	14.7	18.0	32.1	16.3
Sr	XRF	1361	1104	959	815	988	807	1105	1083	982
Y	ICPMS	29.0	30.0	23.5	23.4	23.5	21.1	27.3	26.9	25.2
Zr	XRF	321	307	266	299	225	230	264	264	286
Nb	ICPMS	106.3	117.6	91.1	96.5	99.9	72.5	97.5	91.7	92.3
Cs	ICPMS	0.677	0.485	0.392	0.302	0.299	2.50	1.32	1.06	0.976
Ba	XRF	828	778	797	2961	719	657	870	828	788
La	ICPMS	109.2	109.3	80.5	75.3	77.2	62.4	96.0	89.1	81.6
Ce	ICPMS	208	208	146	147	141	125	182	168	154
Pr	ICPMS	23.8	23.9	16.8	16.8	15.3	15.1	20.6	19.2	17.7
Nd	ICPMS	88.5	88.3	65.5	63.9	55.6	59.1	76.6	73.0	67.0
Sm	ICPMS	14.3	14.5	11.1	11.2	10.4	9.8	12.5	12.1	11.2
Eu	ICPMS	4.47	4.55	3.54	3.53	3.24	3.25	3.97	3.88	3.53
Gd	ICPMS	11.88	12.18	9.48	9.31	8.44	8.90	10.58	10.36	9.45
Tb	ICPMS	1.55	1.60	1.23	1.23	1.08	1.18	1.42	1.36	1.20

(continued)

# Data documentation II

Table 1: Continued

Sample Rock type	Method	S097511 Mel	S097512 Mel	S097519 Mel	S107048 Mel	S107052 Mel	S097513 Neph	S097514 Neph	S097515 Neph	S107045 Neph
Dy	ICPMS	8.09	8.50	6.45	6.35	5.81	6.15	7.51	7.24	6.48
Ho	ICPMS	1.32	1.40	1.09	1.02	0.969	1.02	1.25	1.24	1.08
Er	ICPMS	3.09	3.31	2.54	2.48	2.35	2.36	3.11	2.98	2.63
Tm	ICPMS	0.362	0.406	0.301	0.274	0.279	0.286	0.379	0.358	0.331
Yb	ICPMS	1.73	1.89	1.48	1.40	1.54	1.32	1.84	1.78	1.67
Lu	ICPMS	0.238	0.262	0.199	0.188	0.214	0.180	0.252	0.245	0.238
Hf	ICPMS	7.11	7.17	6.18	7.03	5.46	5.62	6.10	6.06	6.44
Ta	ICPMS	5.57	5.92	4.72	5.31	4.83	4.11	4.92	4.79	5.05
Pb	ICPMS	13.36	11.99	8.72	6.73	4.89	8.20	7.75	7.93	8.77
Th	ICPMS	14.02	14.00	10.30	8.26	9.34	7.74	12.40	11.07	9.40
U	ICPMS	4.01	3.27	3.29	2.16	1.41	2.36	3.69	3.33	2.93
( <sup>206</sup> Pb/ <sup>204</sup> Pb) <sub>m</sub>		18.72	18.54	18.87	18.57	18.92	18.56	18.95	18.67	18.49
( <sup>206</sup> Pb/ <sup>204</sup> Pb) <sub>i</sub>		18.61	18.45	18.73	18.47	18.82	18.47	18.77	18.55	18.41
( <sup>207</sup> Pb/ <sup>204</sup> Pb) <sub>m</sub>		15.62	15.60	15.61	15.61	15.61	15.62	15.64	15.61	15.62
( <sup>207</sup> Pb/ <sup>204</sup> Pb) <sub>i</sub>		15.61	15.60	15.60	15.61	15.61	15.62	15.63	15.60	15.62
( <sup>208</sup> Pb/ <sup>204</sup> Pb) <sub>m</sub>		38.65	38.42	38.77	38.49	38.78	38.49	38.88	38.57	38.42
( <sup>208</sup> Pb/ <sup>204</sup> Pb) <sub>i</sub>		38.52	38.29	38.62	38.36	38.57	38.40	38.68	38.44	38.33
( <sup>176</sup> Hf/ <sup>177</sup> Hf) <sub>m</sub>		0.282881±4	0.282879±4	0.282894±5	0.282868±5	0.282922±4	–	0.282858±7	0.282867±6	0.282917±6
( <sup>176</sup> Hf/ <sup>177</sup> Hf) <sub>i</sub>		0.282878	0.282876	0.282891	0.282865	0.282919		0.282854	0.282864	0.282914
e <sub>Hf(0)</sub>		3.9	3.8	4.3	3.4	5.3	–	3.0	3.4	5.1
e <sub>Hf(i)</sub>		3.7	3.7	4.2	3.3	5.2		2.9	3.3	5.0
( <sup>143</sup> Nd/ <sup>144</sup> Nd) <sub>m</sub>		0.512858±7	0.512846±7	0.512866±10	0.512856±10	0.512879±8	–	0.512838±9	0.512844±8	0.512881±8
( <sup>143</sup> Nd/ <sup>144</sup> Nd) <sub>i</sub>		0.512839	0.512825	0.512841	0.512834	0.512855	–	0.512813	0.512824	0.512864
e <sub>Nd(0)</sub>		4.3	4.1	4.5	4.2	4.7	–	3.9	4.0	4.7
e <sub>Nd(i)</sub>		4.7	4.5	4.9	4.6	5.1		4.4	4.4	5.0
( <sup>87</sup> Sr/ <sup>86</sup> Sr) <sub>m</sub>		0.705187±12	0.704482±11	0.703682±13	0.703687±13	0.703970±32	–	0.703686±11	0.703740±11	0.703517±21
( <sup>87</sup> Sr/ <sup>86</sup> Sr) <sub>i</sub>		0.70515	0.70443	0.70364	0.703637	0.703939		0.70366	0.70370	0.70350

(continued)

# Data documentation III

**Table 2.** U–Pb analytical data of zircon from the microgranite dike at Písečný vrch, Teplice region, Czech Republic.

Sample <sup>a</sup>	Weight (mg)	Concentrations (ppm)		<sup>206</sup> Pb	<sup>232</sup> Th	Radiogenic Pb (at%) <sup>c</sup>				Atomic ratios <sup>c</sup>			Apparent ages (Ma) <sup>d</sup>		
				<sup>204</sup> Pb	<sup>238</sup> U	<sup>206</sup> Pb	<sup>207</sup> Pb	<sup>208</sup> Pb	<sup>238</sup> U	<sup>206</sup> Pb	<sup>207</sup> Pb	<sup>207</sup> Pb	<sup>206</sup> Pb	<sup>207</sup> Pb	<sup>207</sup> Pb
		U	Pb <sub>tot</sub>	Measured ratios <sup>b</sup>	(at) <sup>c</sup>					<sup>238</sup> U	<sup>235</sup> U	<sup>206</sup> Pb	<sup>238</sup> U	<sup>235</sup> U	<sup>206</sup> Pb
<i>Abraded zircon</i>															
1	0.067	220.6	11.90	988	0.51	82.48	4.35	13.16	29100	0.04964	0.3613	0.05279	312	313	320
2	0.103	203.2	11.65	646	0.55	81.64	4.31	14.05	14600	0.05034	0.3665	0.05281	317	317	321
3	0.033	188.0	9.97	1010	0.53	81.94	4.32	13.74	77800	0.04977	0.3614	0.05266	313	313	314
4	0.038	197.3	12.58	350	0.59	80.87	4.29	14.84	8020	0.05253	0.3845	0.05309	330	330	333
5	0.052	211.1	13.04	549	0.52	82.14	4.39	13.47	12640	0.05419	0.3997	0.05339	340	341	350
6	0.087	221.4	14.44	694	0.54	81.59	4.48	13.92	13400	0.05754	0.4360	0.05496	360	367	411
7	0.110	184.1	10.75	727	0.59	80.70	4.25	15.04	16600	0.05119	0.3720	0.05270	322	321	316
8	0.088	92.7	5.32	750	0.56	81.36	4.30	14.34	24000	0.05154	0.3752	0.05279	324	324	320
9	0.087	215.4	14.88	200	0.56	81.46	4.25	14.29	3840	0.04986	0.3587	0.05217	314	311	293
10	0.024	506.4	30.27	723	0.62	80.16	4.25	15.59	18400	0.05249	0.3837	0.05301	330	330	329
11	0.095	216.7	12.47	326	0.56	81.46	4.30	14.24	7200	0.04623	0.3365	0.05280	292	295	320
12	0.096	269	18.86	405	0.52	82.10	4.55	13.35	7100	0.05833	0.4455	0.05539	366	374	428
13	0.132	169	19.36	69.6	0.58	80.99	4.53	14.66	1030	0.05041	0.3735	0.05374	317	322	360
<i>Weakly abraded and broken zircon</i>															
14	0.133	125.6	18.46	52.3	0.59	80.71	4.26	15.04	680	0.05082	0.3697	0.02753	320	319	318
15	0.128	131.3	41.4	29.95	0.68	79.18	4.03	16.79	240	0.0494	0.347	0.0509	311	302	236
16	0.121	182.2	13.35	172	0.59	80.88	4.27	14.85	3200	0.05017	0.3650	0.05277	316	316	319
<i>Non-abraded zircon</i>															
17															
18	0.146	277.3	17.25	184	0.70	77.99	4.95	17.06	4030	0.04221	0.3695	0.06350	267	319	725
19	0.222	953	123.9	54.0	0.58	81.09	4.22	14.69	778	0.04619	0.3315	0.05204	291	291	287
20	0.222	180.3	10.2	199	0.61	79.61	5.04	15.35	4660	0.03984	0.3476	0.06327	252	303	717
21	0.154	356.8	30.95	87.3	0.54	81.79	4.37	13.84	1570	0.04434	0.3268	0.05344	280	287	348
22	0.445	232.5	14.04	226	0.66	78.75	4.99	16.26	4860	0.04335	0.3785	0.06332	274	326	719

<sup>a</sup> Zircon concentrates were cleaned in warm 7N HNO<sub>3</sub> and H<sub>2</sub>O before dissolution in concentrated HF in teflon-lined autoclaves at 210 °C for 4 days. A mixed <sup>205</sup>Pb–<sup>235</sup>U tracer was added before sample dissolution. Lead and U were separated using ion-exchange chromatographic procedures (e.g., ROMER et al. 1996; SCHMID et al. 2003). Lead and U were loaded together on single Re-filaments using a silica-gel emitter and H<sub>3</sub>PO<sub>4</sub> (GERSTENBERGER & HAASE 1997) and measured at 1200–1260 °C and 1350–1400 °C, respectively, on a Finnigan MAT262 multicollector mass-spectrometer (modified with an ion source by Spectromat) using Faraday collectors and ion counting.

<sup>b</sup> Lead isotope ratios corrected for fractionation (0.1 ‰ / a.m.u.).

<sup>c</sup> Lead corrected for fractionation, blank, tracer contribution, and initial lead (<sup>206</sup>Pb/<sup>204</sup>Pb = 18.1 ± 0.1, <sup>207</sup>Pb/<sup>204</sup>Pb = 15.60 ± 0.02, <sup>208</sup>Pb/<sup>204</sup>Pb = 38.2 ± 0.2). During the measurement period, total blanks were < 15 pg for lead and < 1 pg for uranium. <sup>232</sup>Th/<sup>238</sup>U ratios calculated for an age of 320 Ma from <sup>208</sup>Pb<sub>rad</sub>/<sup>206</sup>Pb<sub>rad</sub>.

<sup>d</sup> Apparent ages were calculated using the constants recommended by IUGS (STEIGER & JÄGER 1977).

# Databases

Large datasets need an adequate infrastructure:  
**DATABASES**

**Available geochemical databases are:**

**EarthChem** (PetDB, SedDB, Geochron,...)

<https://earthchem.org>

**GEOROC** (Geochemistry of Rocks of the Oceans and Continents): <http://georoc.mpch-mainz.gwdg.de/georoc/>

**GERM** (Geochemical Earth Reference Model)

<https://earthref.org/GERM/>

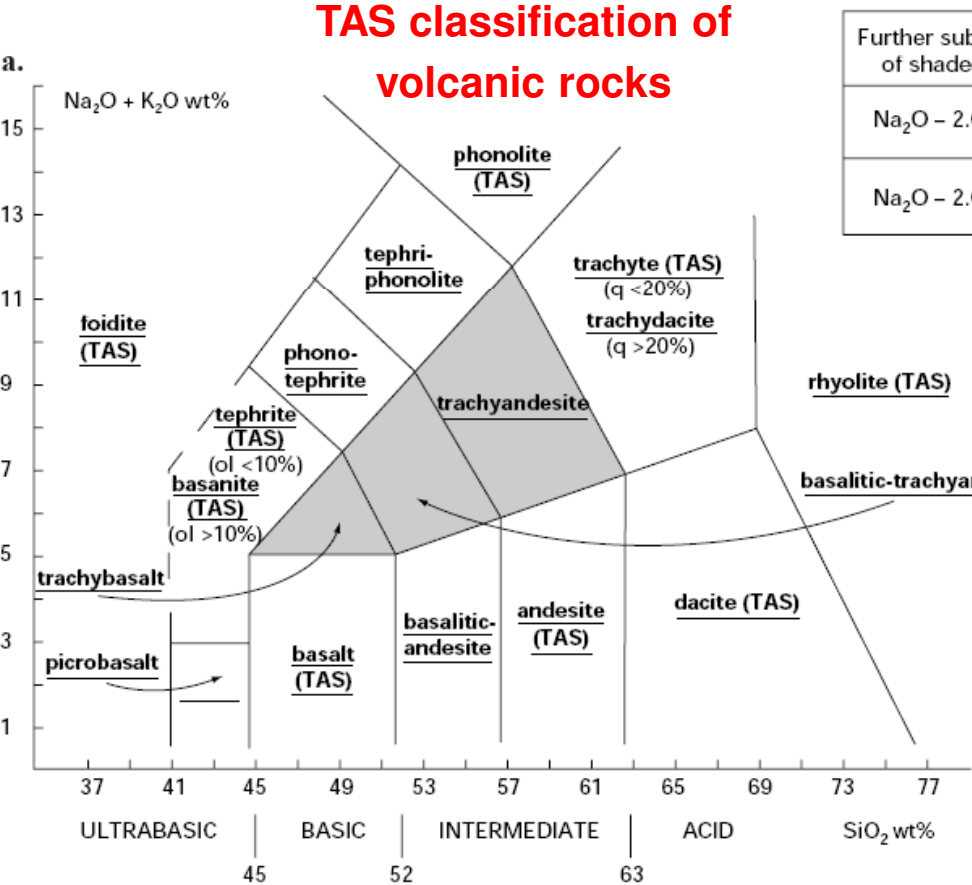
# How can we use geochemical data?

- Rock classification !
- Process identification !
- Reconstruct ancient Tectonic  
Settings... !(?)


# Use of geochemical data – rock classification

## Common types of magmatic rocks

- Classification of the most common rock types by chemistry



Further subdivisions of shaded fields	<u>trachybasalt</u>	<u>basaltic-trachyandesite</u>	<u>trachyandesite</u>
$\text{Na}_2\text{O} - 2.0 \geq \text{K}_2\text{O}$	<u>hawaiite</u>	<u>mugearite</u>	<u>benmoreite</u>
$\text{Na}_2\text{O} - 2.0 \leq \text{K}_2\text{O}$	<u>potassic-trachybasalt</u>	<u>shoshonite</u>	<u>latite (TAS)</u>


  
 Note that trachybasalts, basaltic-trachyandesites and trachyandesites are further subdivided based on their **Na/K ratio** [ f(p) ]

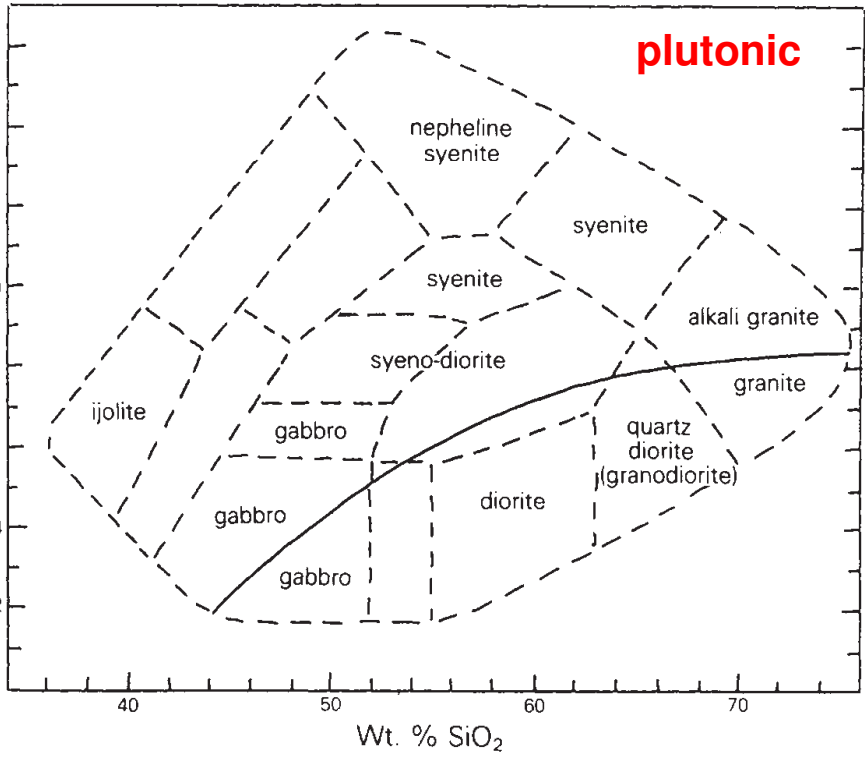
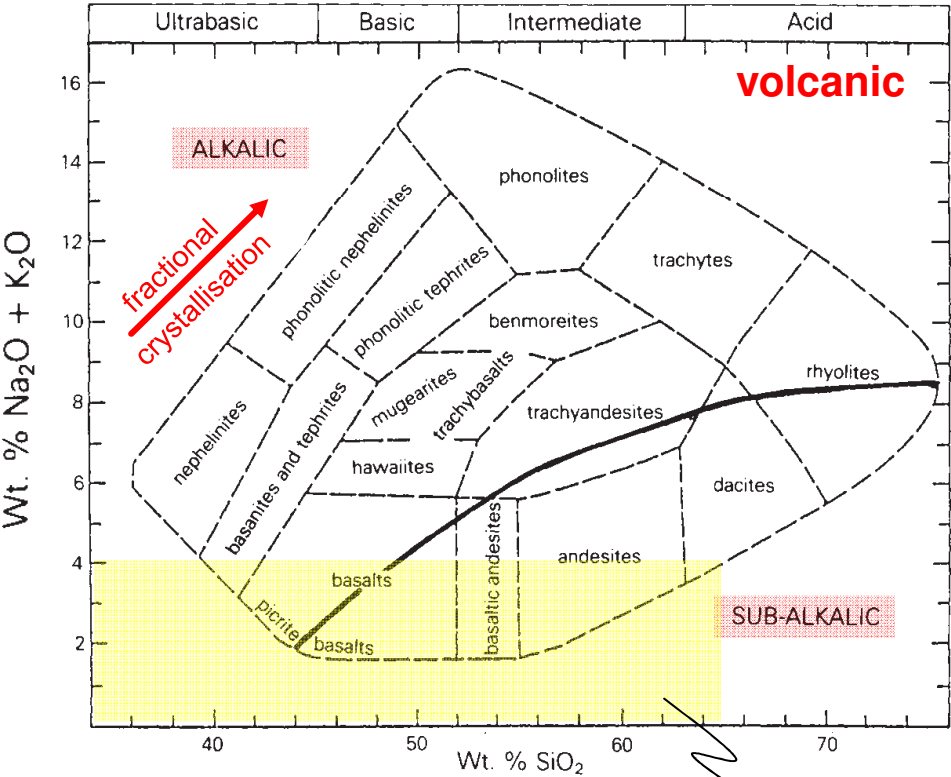
*From: Gillespie & Styles, BGS (British Geological Survey) Rock Classification – Vol 1 – Classification of igneous rocks*

# Use of geochemical data – rock classification

## Common types of magmatic rocks

- Classification of the most common rock types by chemistry

### TAS classification for volcanic and plutonic rocks

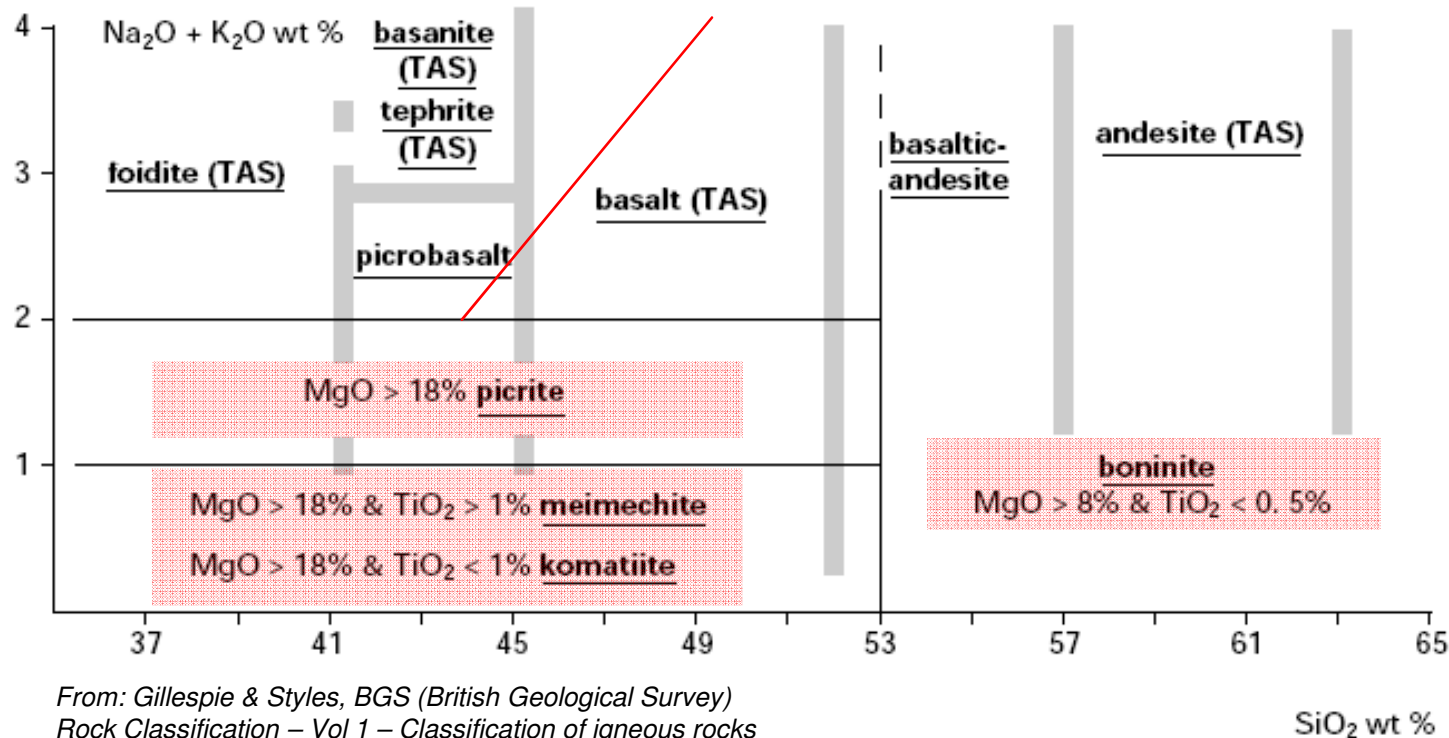


range shown in next slide....

From: Wilson, M.: *Igneous Petrogenesis*, Springer.

## Common types of magmatic rocks

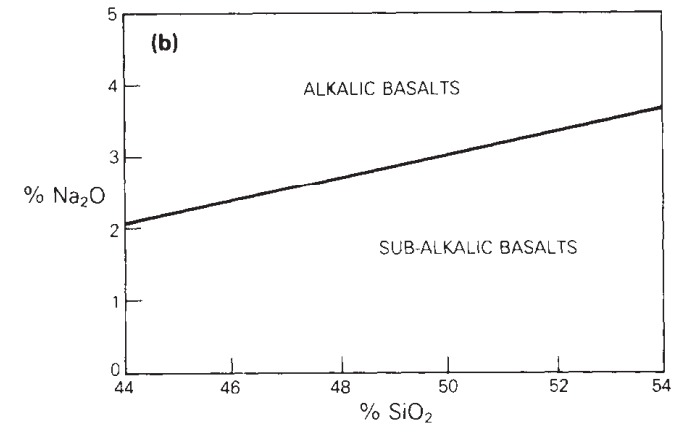
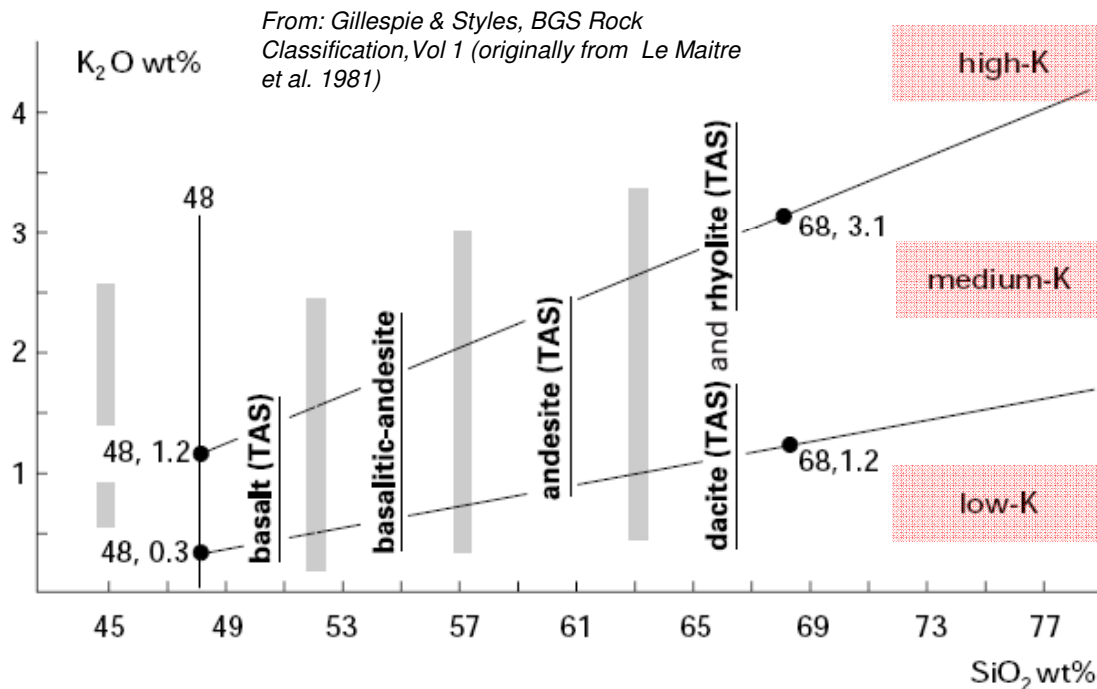
- Classification of the most common rock types by chemistry



**Picrites, Meimechites, Komatiites and Boninites** are defined by combining the TAS nomenclature with MgO and TiO<sub>2</sub> contents

## Common types of magmatic rocks

- Classification of the most common rock types by chemistry



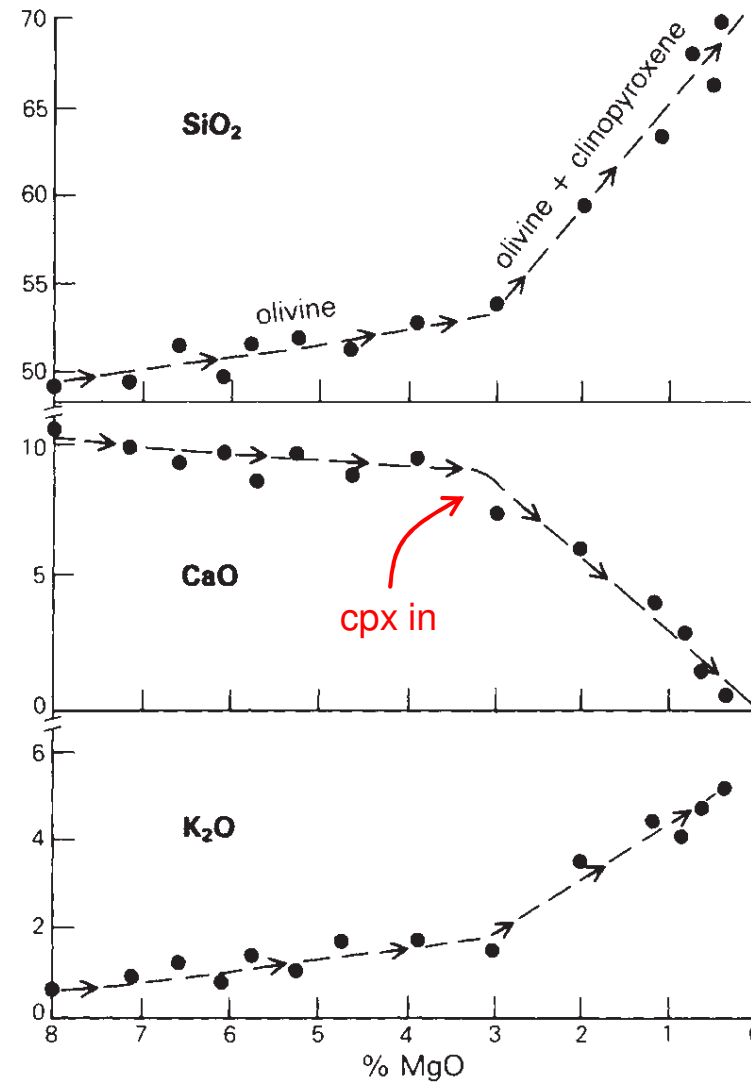
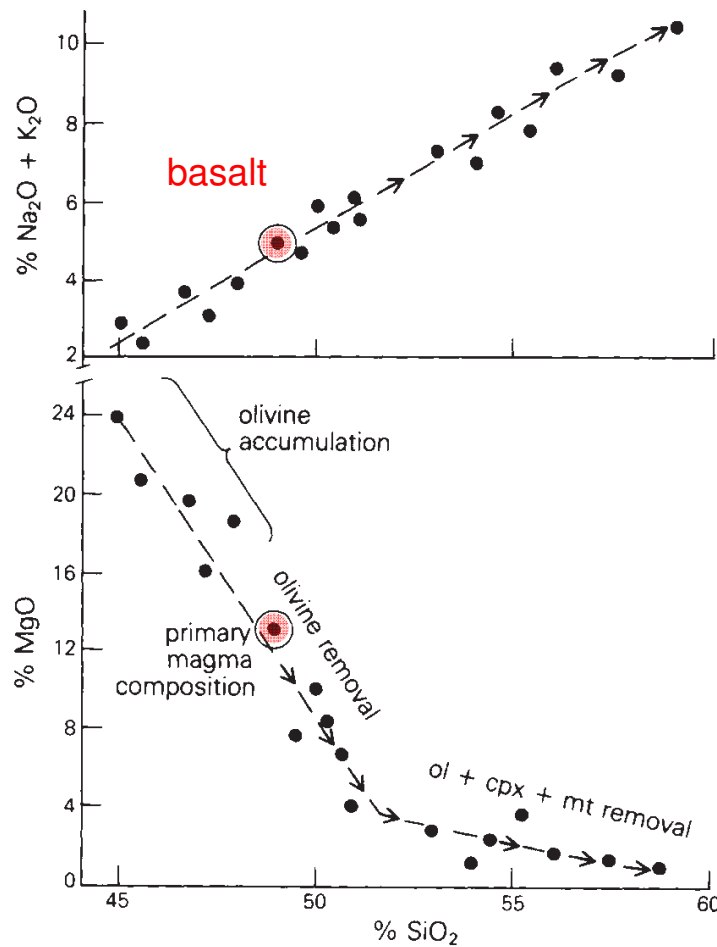
Subdivision of basalts in  
alkalic and subalkalic

Refinement of the TAS classification by using the K<sub>2</sub>O (and Na<sub>2</sub>O) contents

## Process identification

- Fractional crystallisation

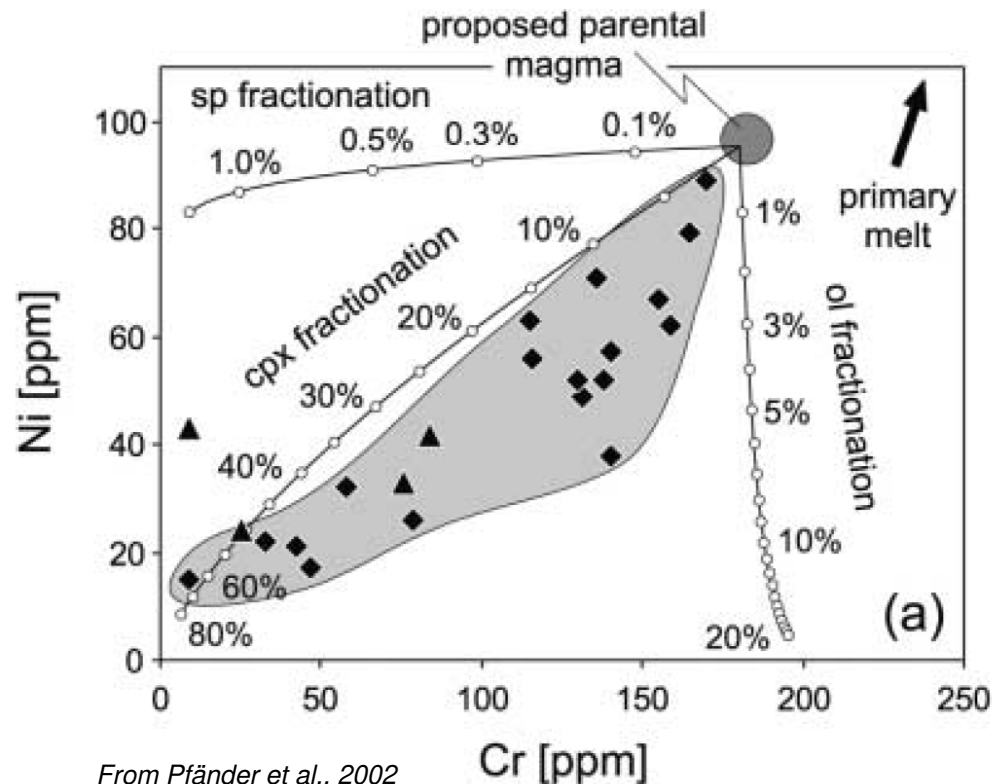
From: Wilson, M.: *Igneous Petrogenesis*, Springer.



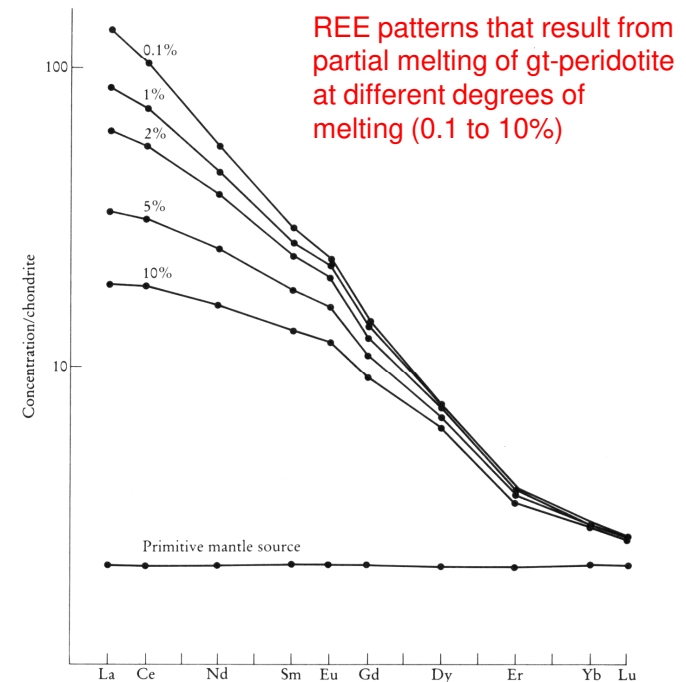
## Process identification

- Fractional crystallisation (and partial melting)

How do REEs behave upon **fractional crystallisation** and **partial melting**?



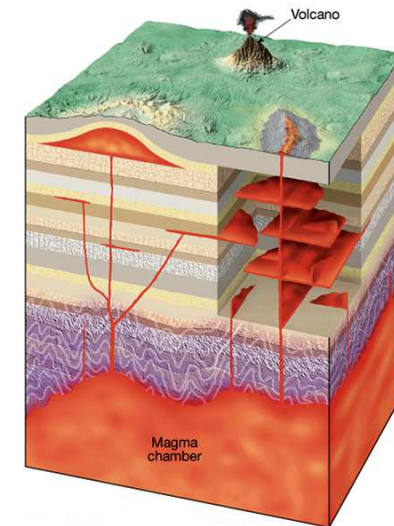
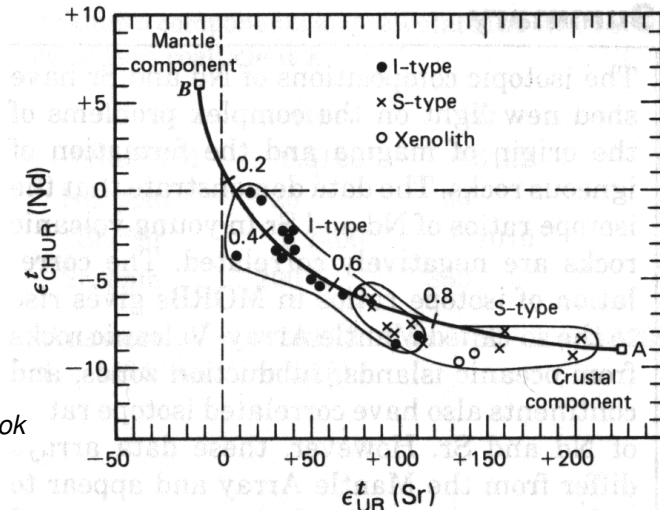
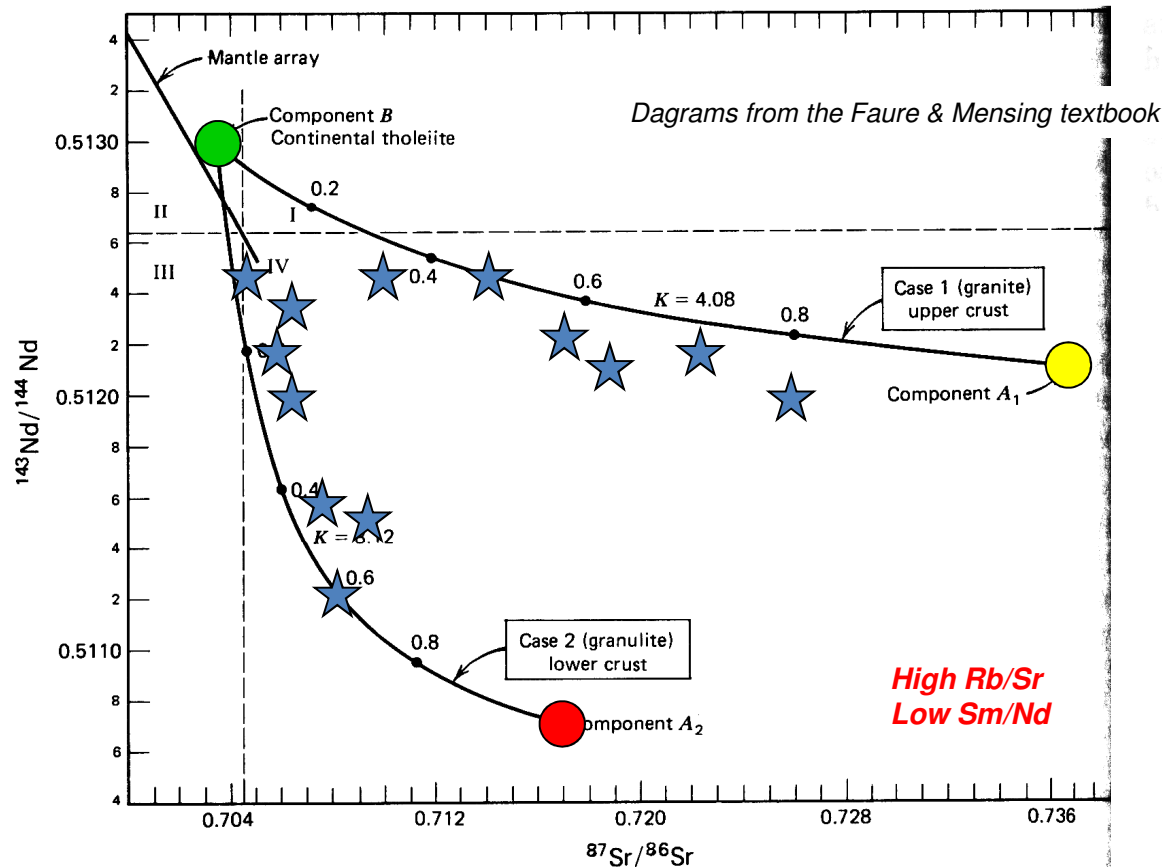
From Pfänder et al., 2002



From Rollinson, Using geochemical data

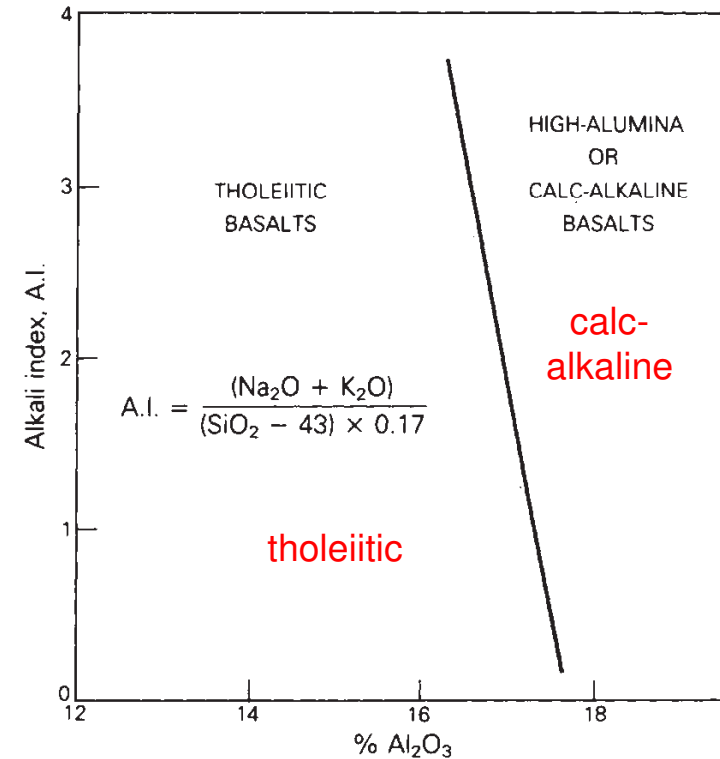
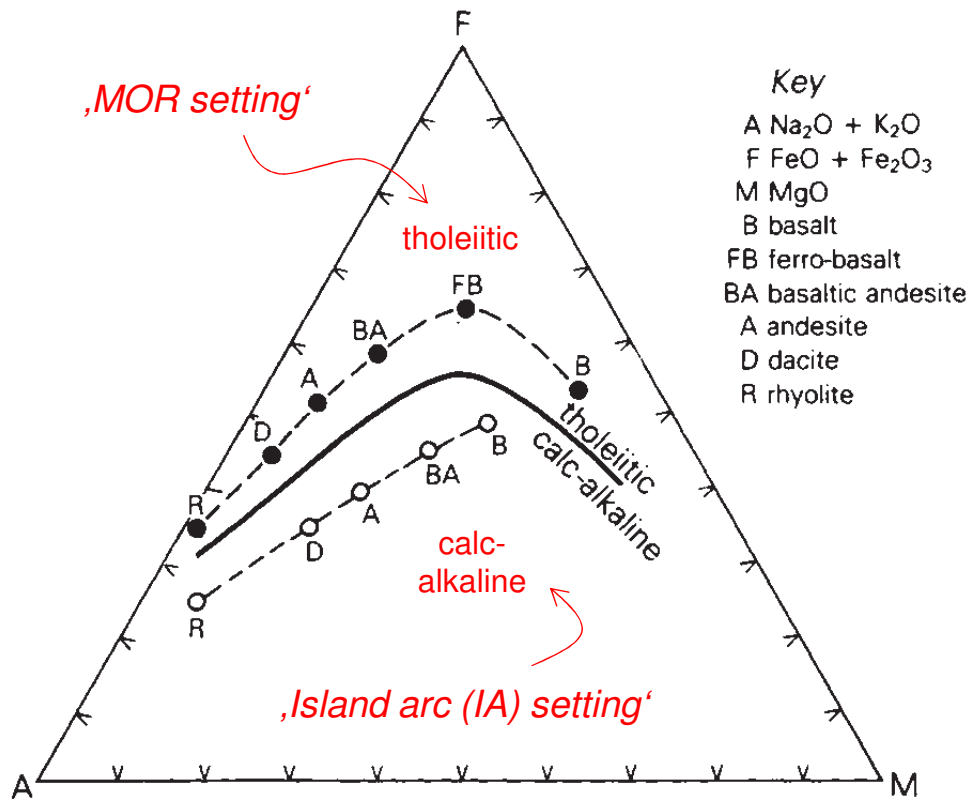
## Process identification

- Crustal assimilation of mantle derived primary magmas



# Tectonic setting

Fe/Mg ratio and crystallisation trends in the ternary plot  
 FeO – MgO – Na<sub>2</sub>O+K<sub>2</sub>O



Both taken from: Wilson, M.: Igneous Petrogenesis, Springer. A.I. vs. Al<sub>2</sub>O<sub>3</sub> from Middlemost 1975)

## Tectonic setting

- Magma series and their first-order tectonic affinity

**Table 1.3** Characteristic magma series associated with specific tectonic settings.

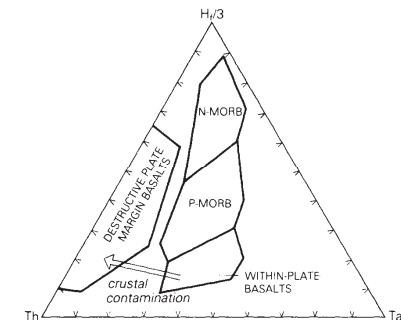
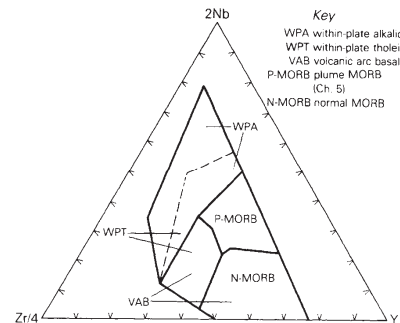
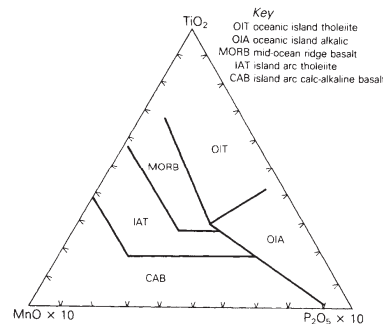
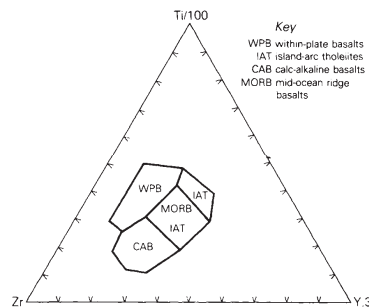
Tectonic setting	Plate margin		Within plate	
	Convergent (destructive)	Divergent (constructive)	Intra-oceanic	Intra-continental
volcanic feature	island arcs, active continental margins	mid-oceanic ridges back-arc spreading centres	oceanic islands	continental rift zones continental flood- basalt provinces
characteristic magma series	tholeiitic calc-alkaline alkaline	tholeiitic — —	tholeiitic — alkaline	tholeiitic — alkaline
SiO <sub>2</sub> range	basalts and differentiates	basalts	basalts and differentiates	basalts and differentiates

*From: Wilson, M.: Igneous Petrogenesis, Springer.*

# How else can we use geochemical data?

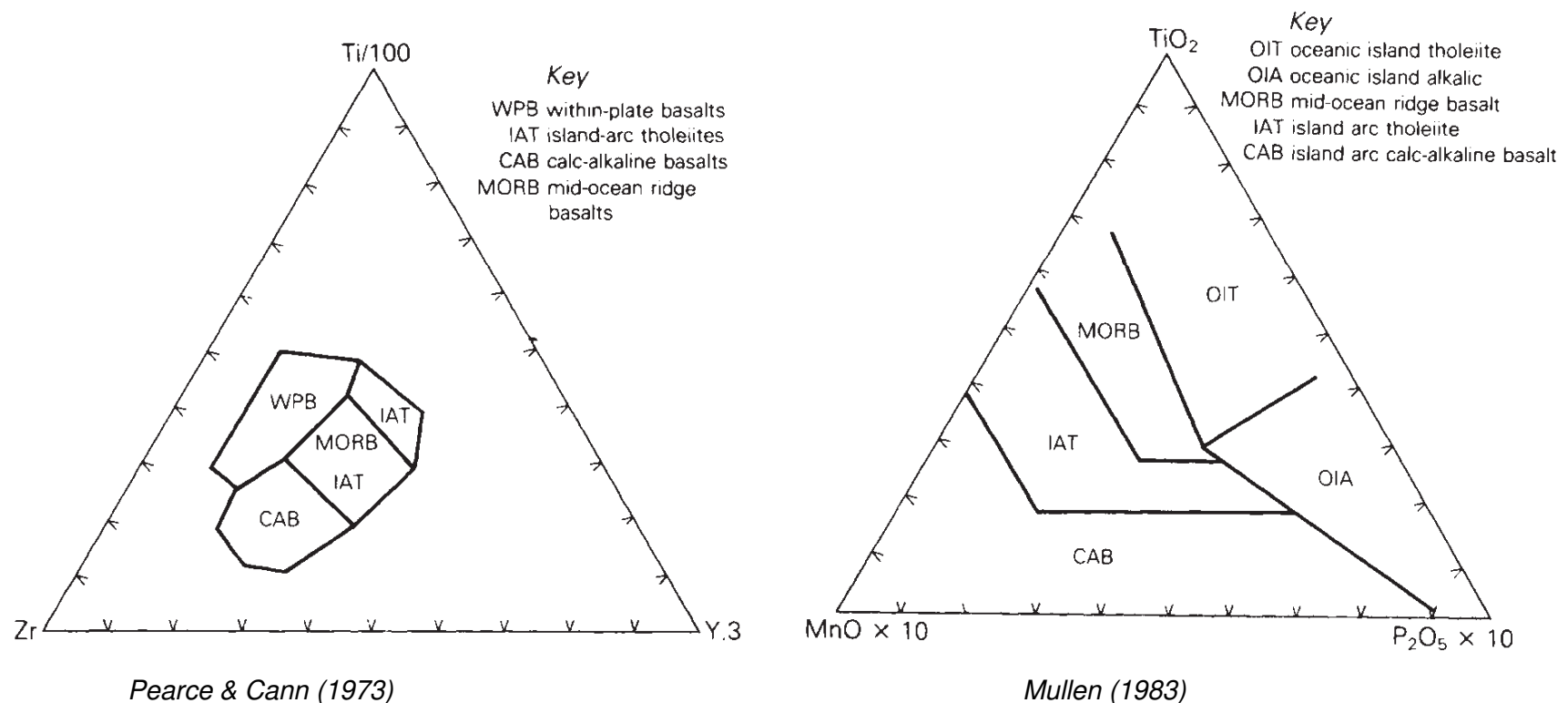
And how we should not?

Or why we should better not use the 'classical' tectono-magmatic discrimination diagrams



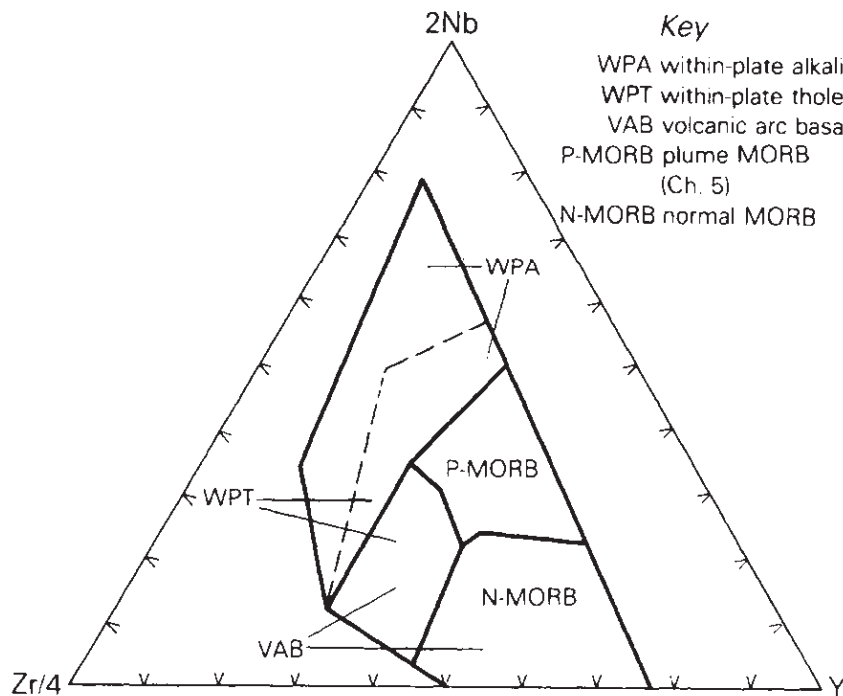
# Use and abuse of geochemical data – tectonic indiscrimination

## Rationale behind the 'classical' tectonomagmatic discrimination diagram



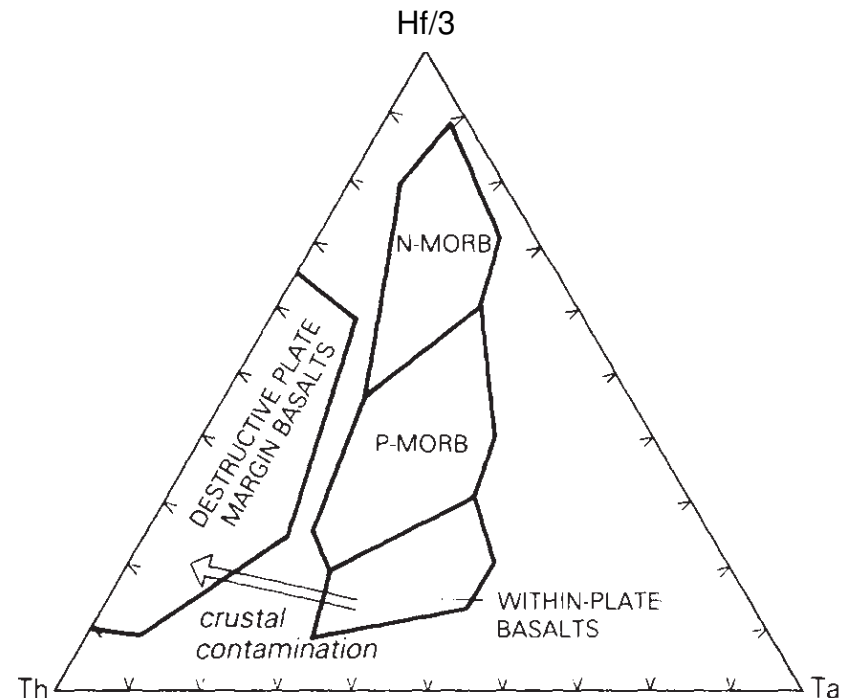
**EASY TO USE**, no detailed process understanding required (simply measure, plot, read the result). All elements are easy to determine (XRF, INAA).

## Rationale behind the ‘classical’ tectonomagmatic discrimination diagram



Meschede (1986)

Key  
WPA within-plate alkalic  
WPT within-plate tholeiite  
VAB volcanic arc basalt  
P-MORB plume MORB  
(Ch. 5)  
N-MORB normal MORB



Wood et al. (1979)

**EASY TO USE**, no detailed process understanding required (simply measure, plot, read the result). All elements are easy to determine (XRF, INAA).

# Use and abuse of geochemical data – tectonic indiscrimination

Lithos 232 (2015) 76–83



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## Trace element indiscrimination diagrams



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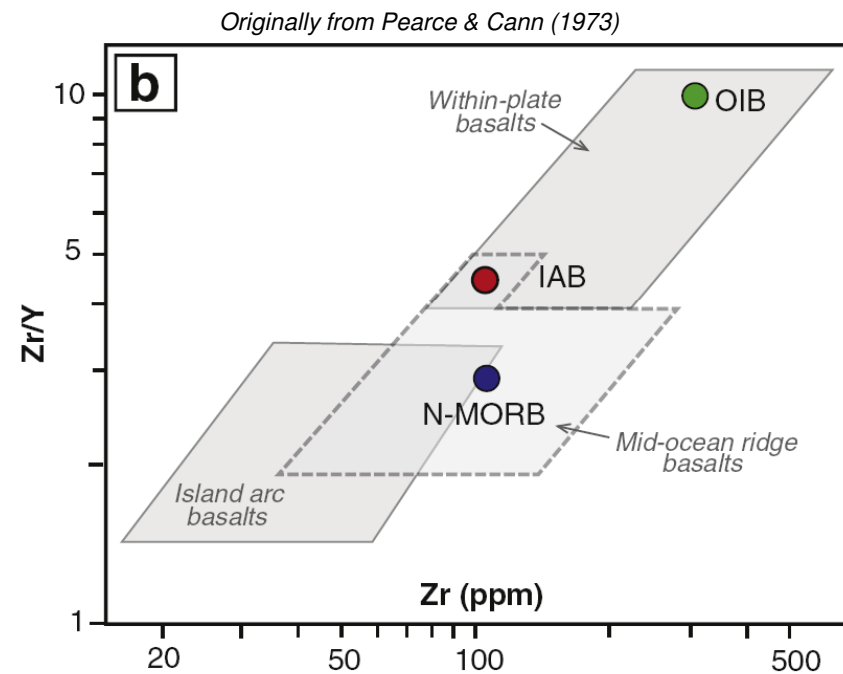
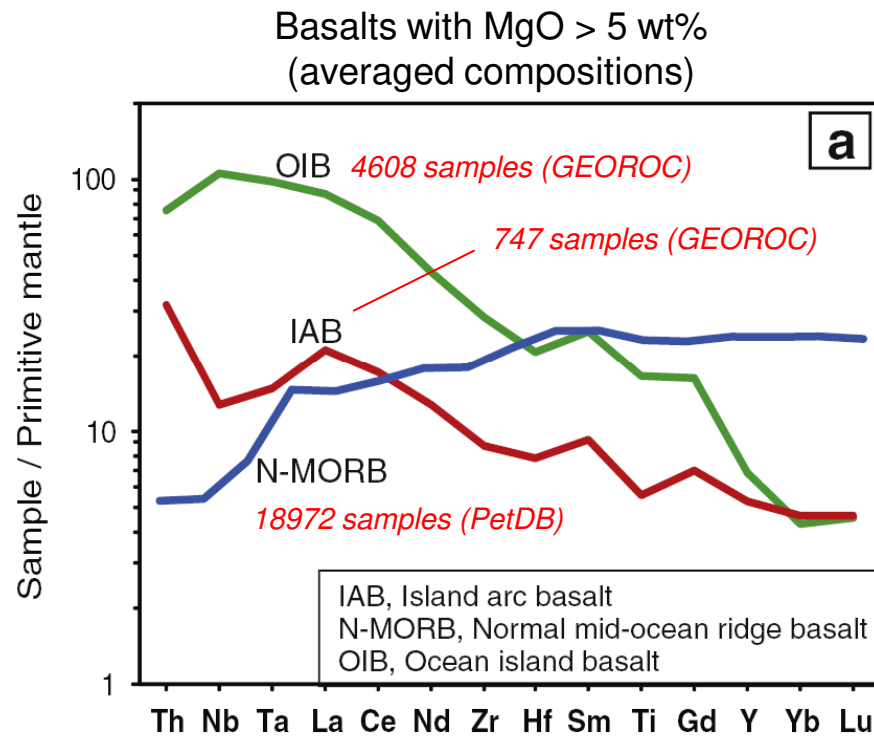
Petrological database utilization

### ABSTRACT

We tested the accuracy of trace element discrimination diagrams for basalts using new datasets from two petrological databases, PetDB and GEOROC. Both binary and ternary diagrams using Zr, Ti, V, Y, Th, Hf, Nb, Ta, Sm, and Sc do a poor job of discriminating between basalts generated in various tectonic environments (continental flood basalt, mid-ocean ridge basalt, ocean island basalt, oceanic plateau basalt, back-arc basin basalt, and various types of arc basalt). The overlaps between the different types of basalt are too large for the confident application of such diagrams when used in the absence of geological and petrological constraints. **None of the diagrams we tested can clearly discriminate between back-arc basin basalt and mid-ocean ridge basalt, between continental flood basalt and oceanic plateau basalt, and between different types of arc basalt** (intra-oceanic, island and continental arcs). Only ocean island basalt and some mid-ocean ridge basalt are generally distinguishable in the diagrams, and even in this case, mantle-normalized trace element patterns offer a better solution for discriminating between the two types of basalt.

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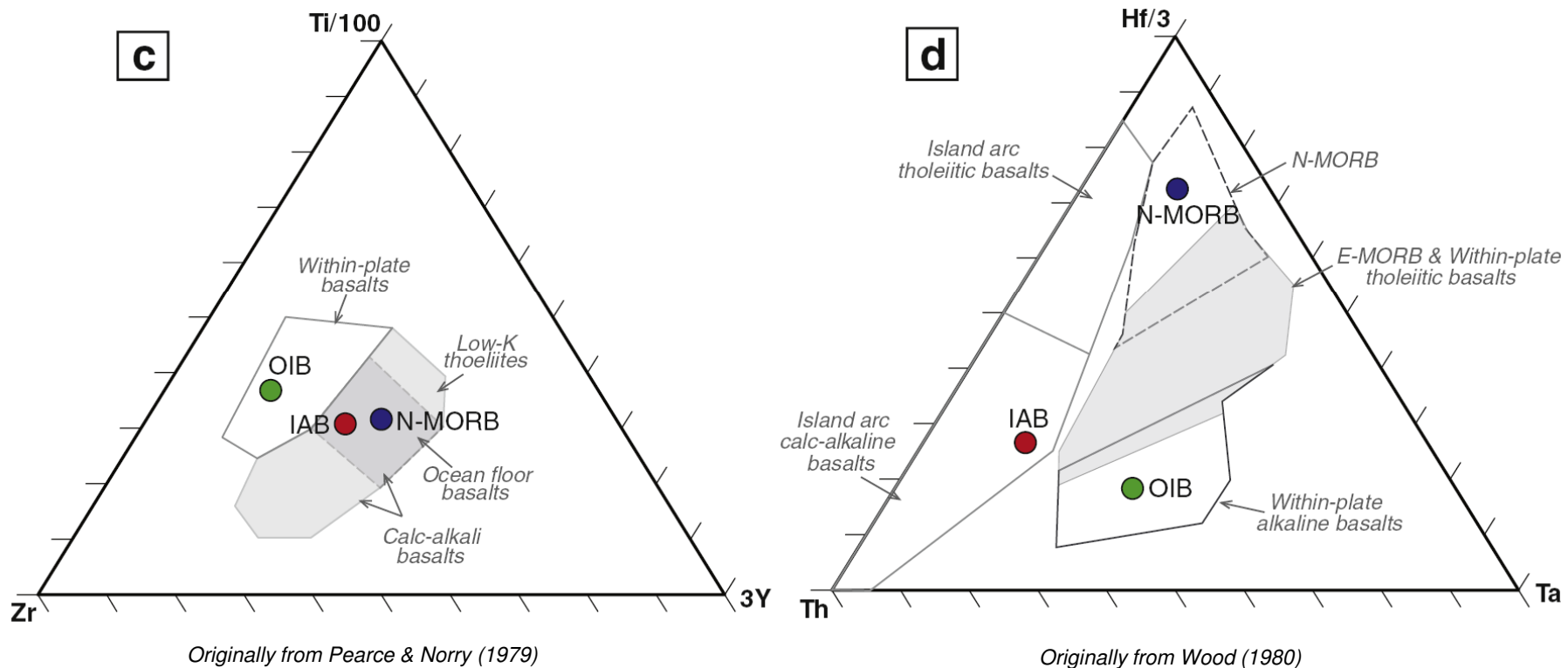
## Tectonic indiscrimination...



Li et al. (2015)

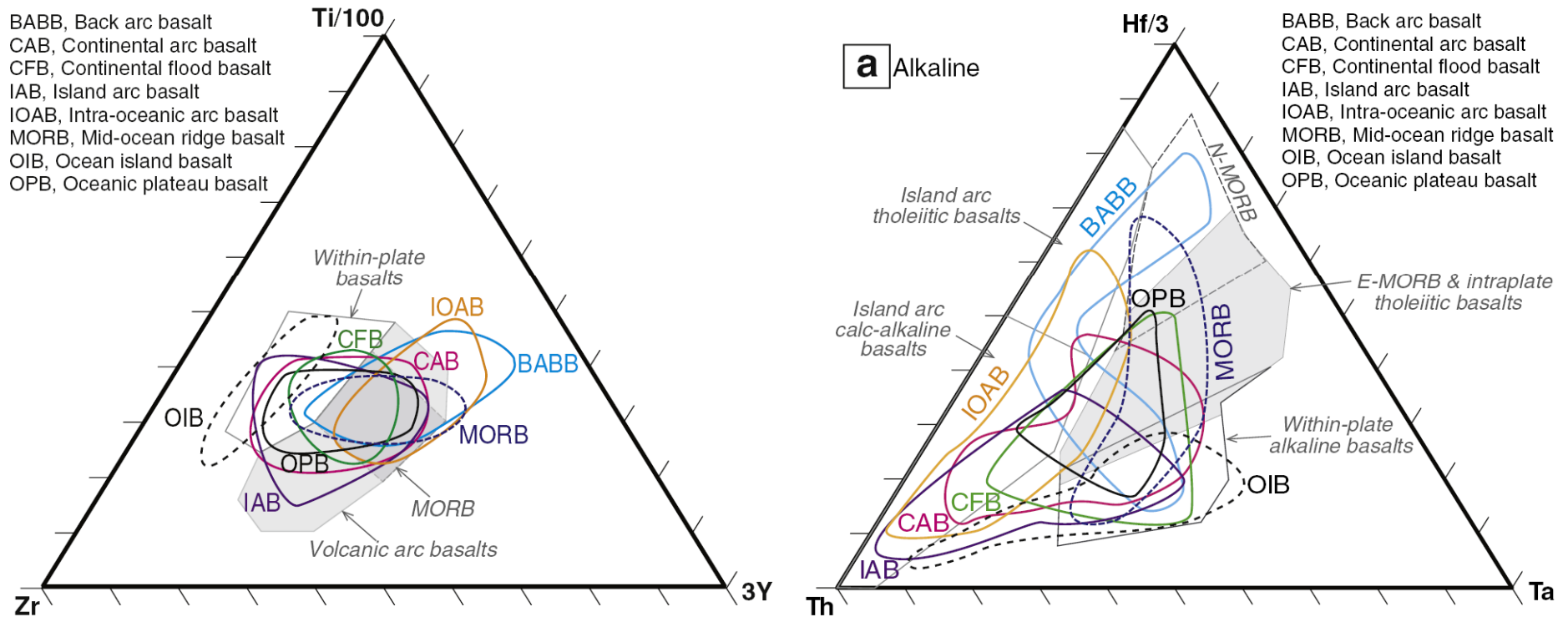
The **averaged** compositions of **OIB** and **MORB** fit relatively well into the fields proposed by Pearce & Cann (1973), but the **IAB** composition is completely off.

## Tectonic indiscrimination...



The **averaged** compositions of **OIB**, **IAB** and **MORB** fit into the fields proposed by Pearce & Norry (1979), but are very close to each other. By contrast, the settings are well discriminated by the fields defined by Wood (1980)

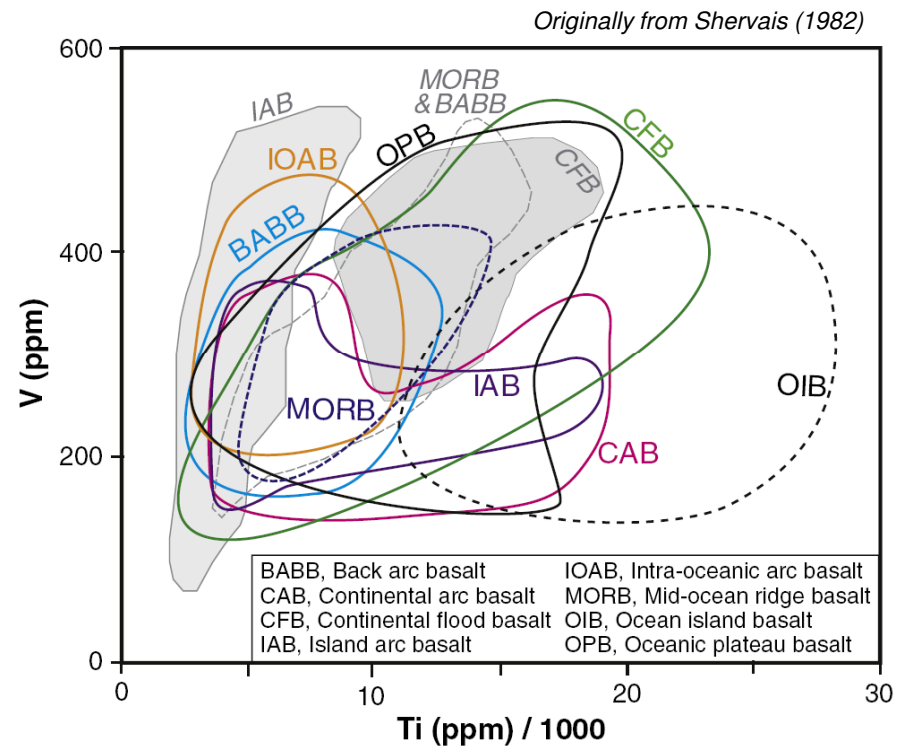
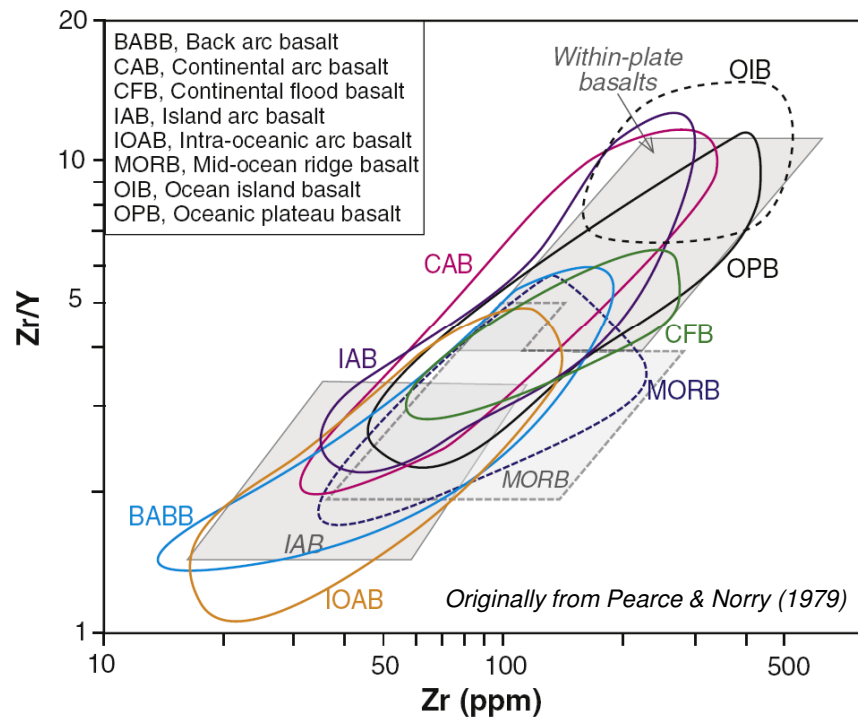
## Tectonic indiscrimination... a closer look



Using a very **large dataset**, the classical discrimination diagrams fail to discriminate different types of basalt from different tectonic settings (MORB, BABB, IOAB, IAB, CAB, OPB, CFB), only **OIBs** are slightly offset!

# Use and abuse of geochemical data – tectonic indiscriminatio

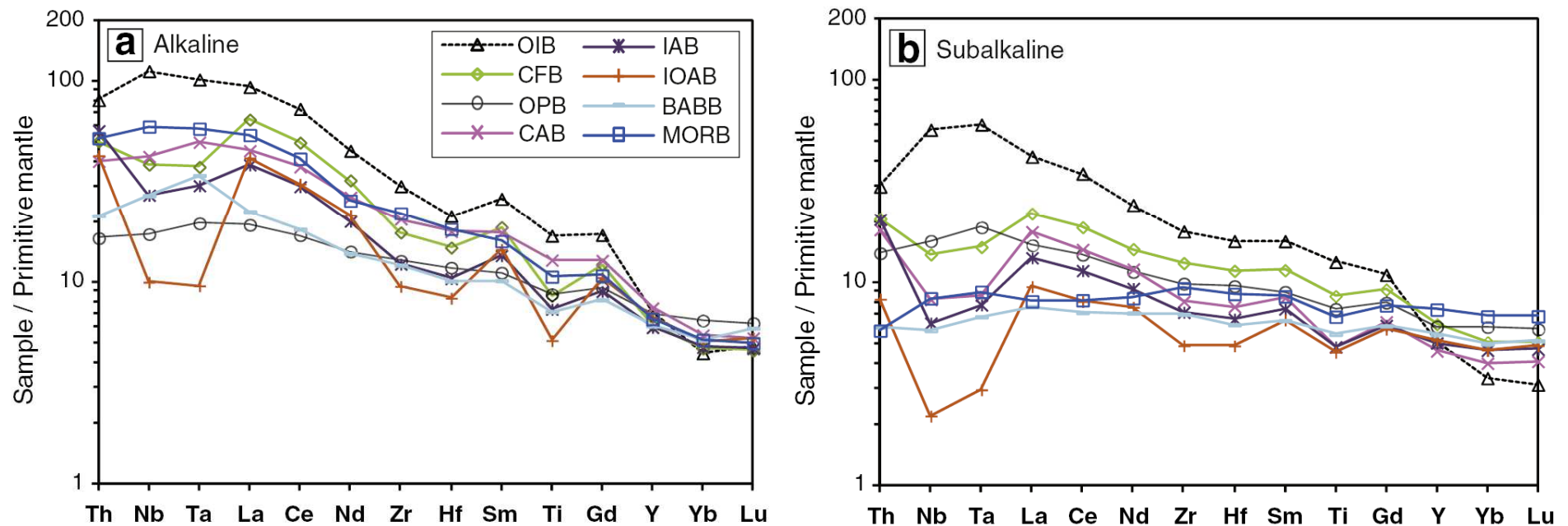
## Tectonic indiscriminatio... a closer look



This validates for any other of the classical discrimination diagrams, without any exception! Therefore... more sophisticated approaches are required!

## Tectonic indiscrimination... what to do?

Multi-element concentration diagrams (normalized to the „primitive mantle“ composition) do a better job



Including more elements improves the **resolution!** Alkaline and subalkaline series have **similar patterns**, but distinctly **different concentrations**.

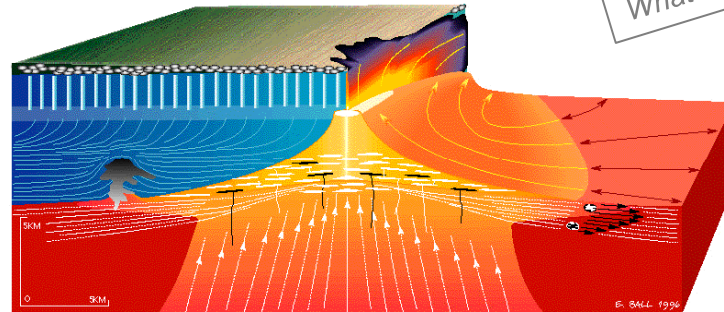
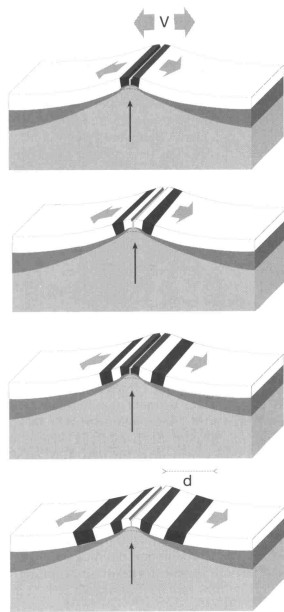
**BUT – NEVER IGNORE THE GEOLOGICAL CONTEXT!**

Chapter 2

# Mid ocean ridges

What is the composition of primary basalts generated at mid-ocean ridges?

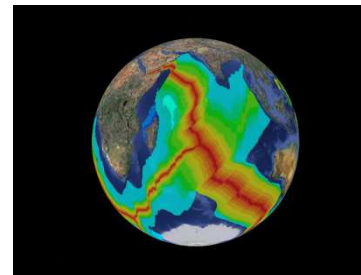
What are the degrees of melting?



Melting at a mid-ocean ridge (from A. Nicolas)

What is the composition of the asthenospheric mantle source?

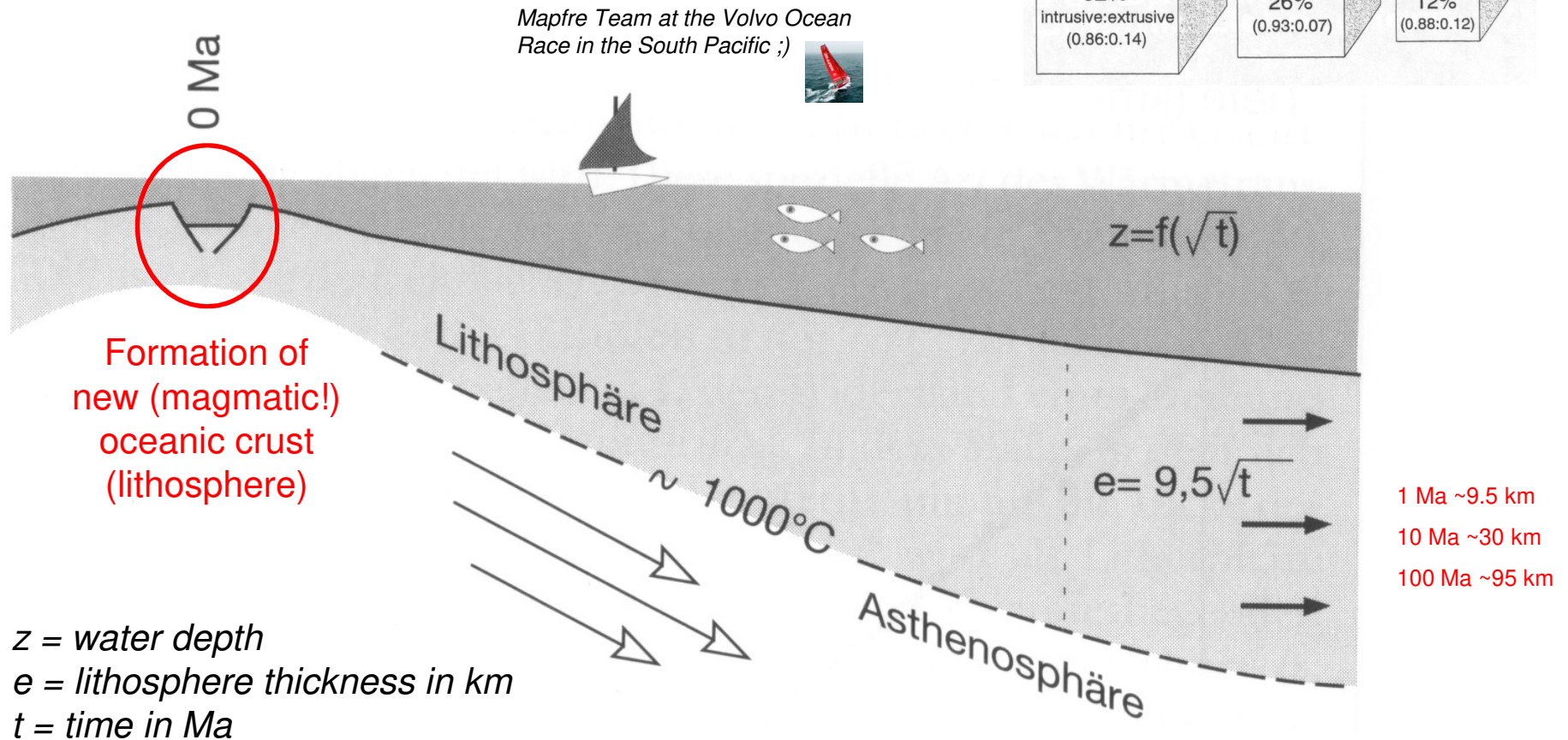
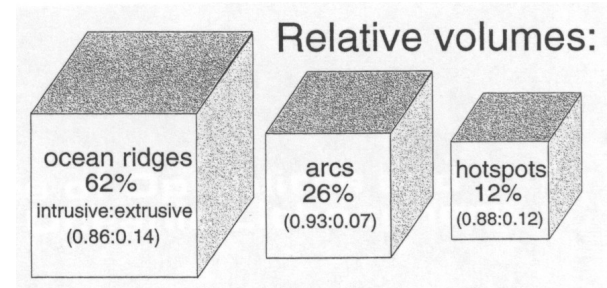
What is the thermal state of the oceanic mantle?



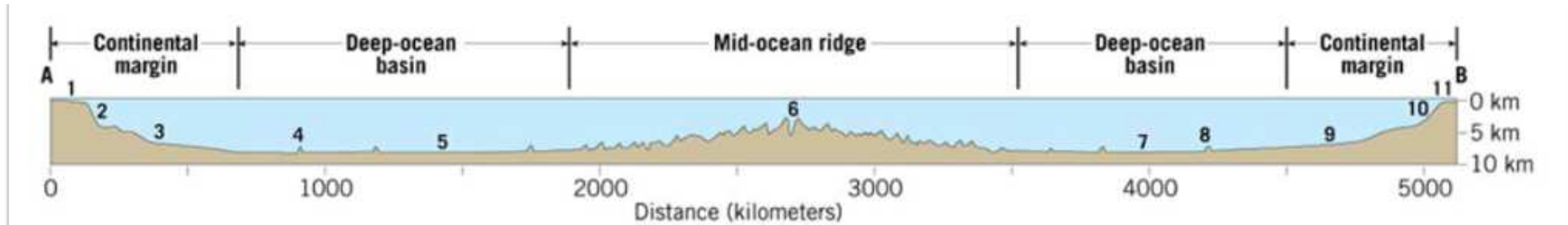
What are the spreading rates, do they correlate to melt production rates and ridge topography??

# Very Simple profile through the oceanic lithosphere

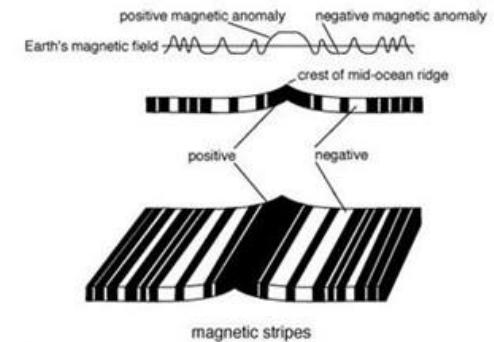
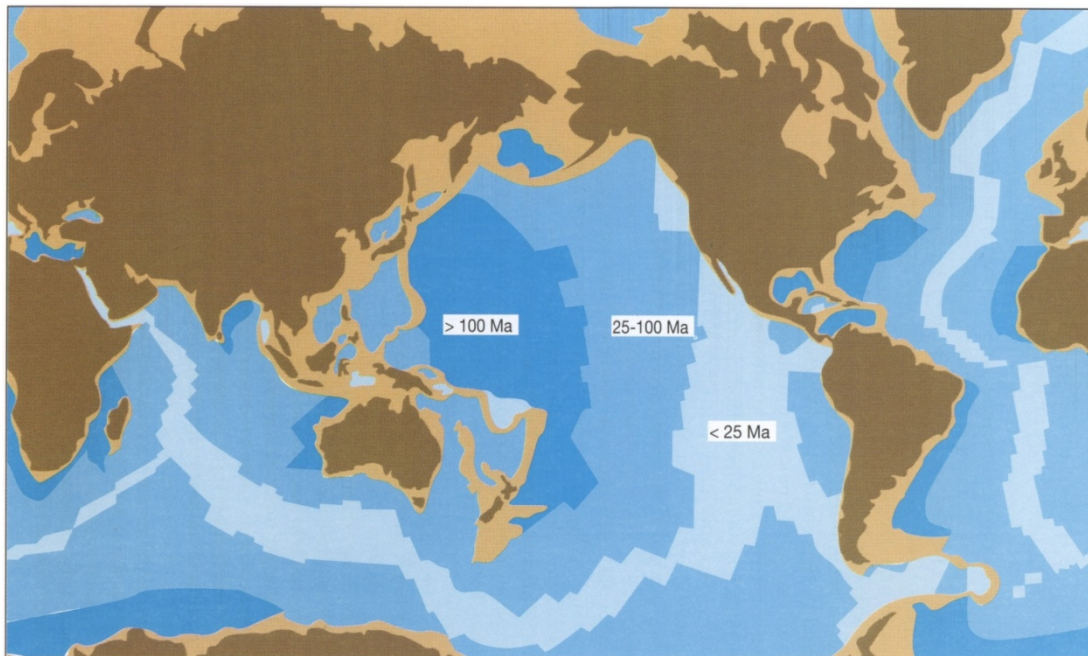
Temperature, water depth and lithosphere thickness as a function of lithosphere age



## Schematic profile across an ocean ridge...



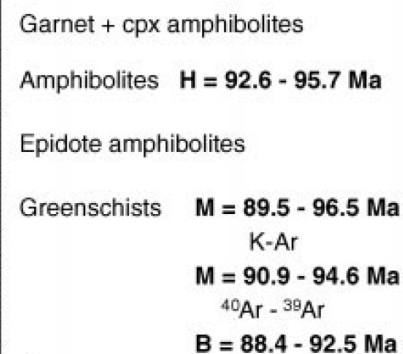
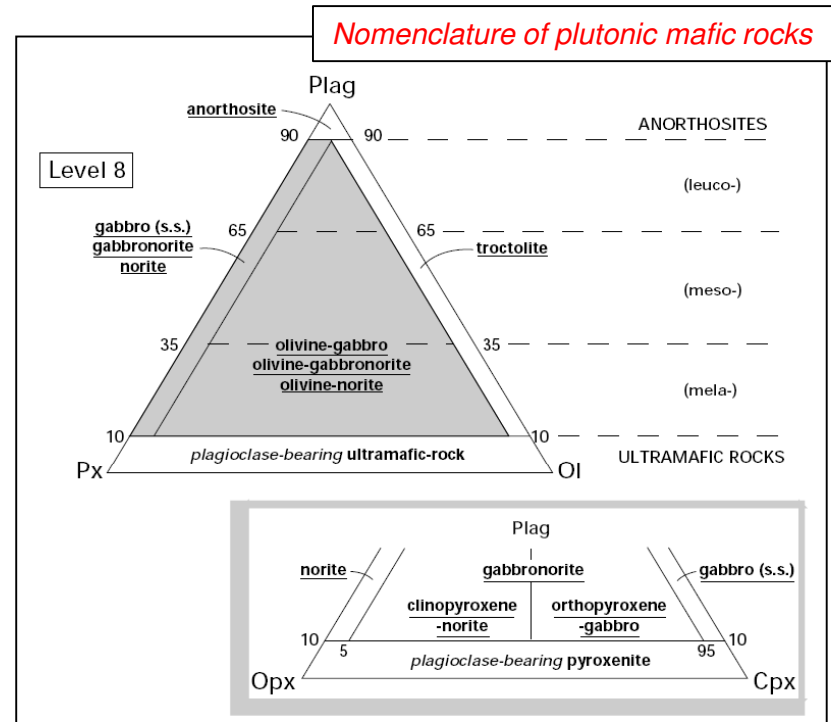
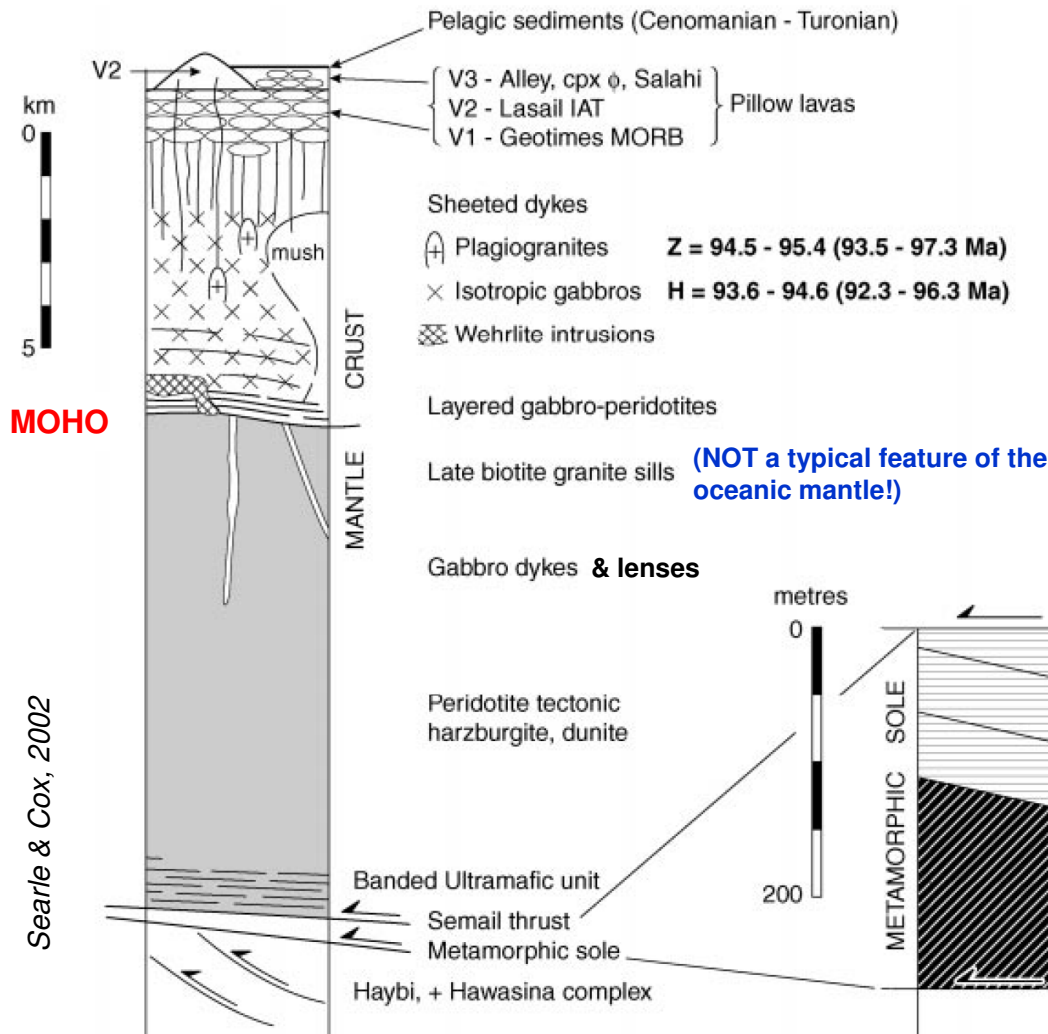
## ...and simplified age distribution of the ocean floor



Magnetic structure of the basaltic ocean floor preserves the **magnetic field reversals**

This is due to the **remanent magnetisation** of Fe-bearing phases (Ti-magnetite, etc.) upon cooling

## Lithology of the oceanic lithosphere as deduced from ophiolites (here: Oman)



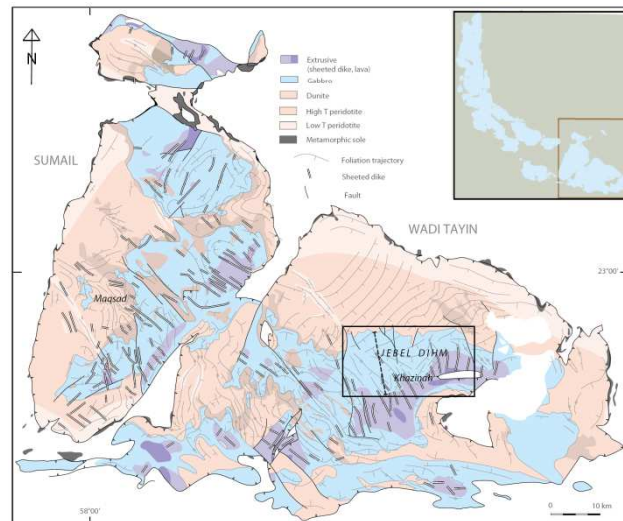
hotter

inverse temperature gradient

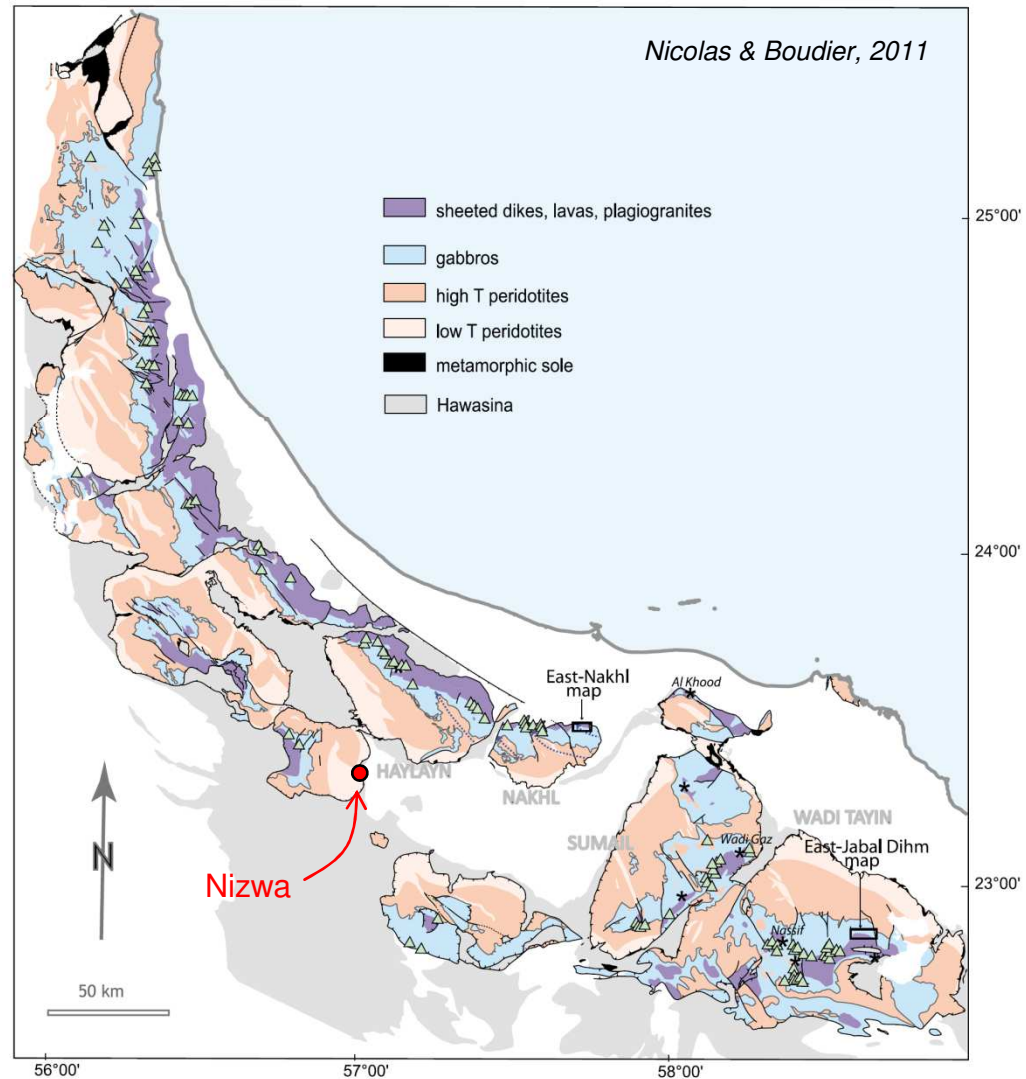
colder

# Lithology of the oceanic lithosphere as deduced from ophiolites

Ophiolite example: OMAN

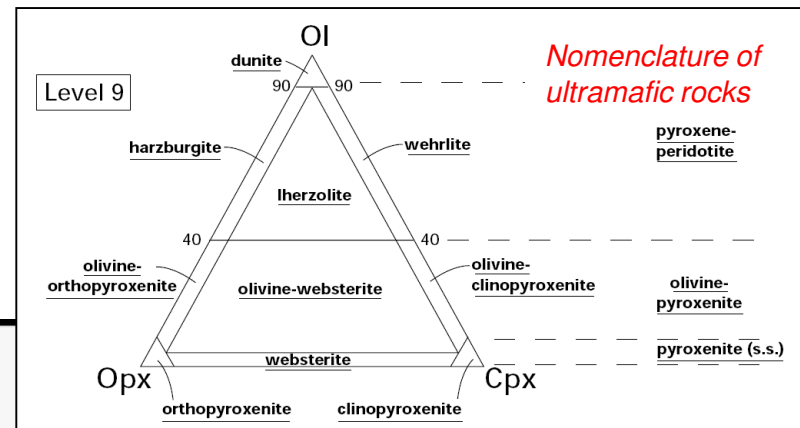
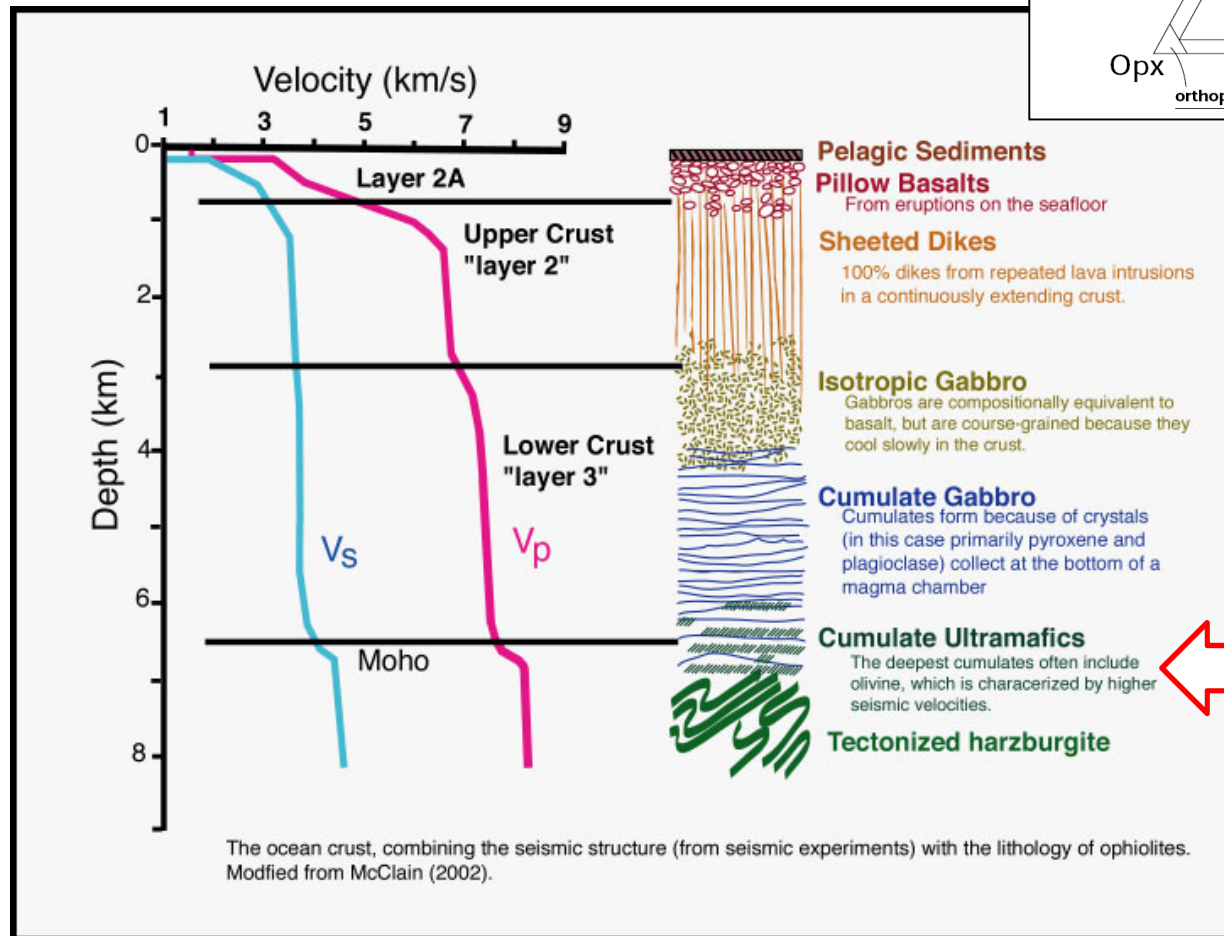


Nicolas & Boudier, 2015



Nicolas & Boudier, 2011

## Seismic structure of the oceanic crust... ... and its lithological interpretation



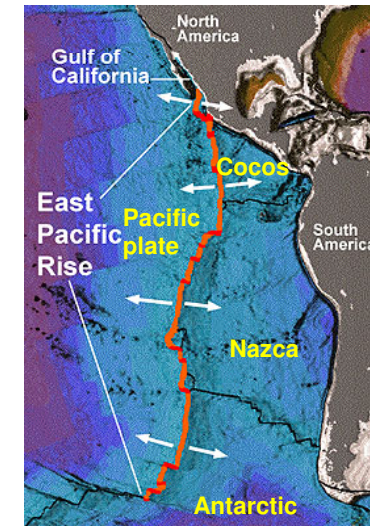
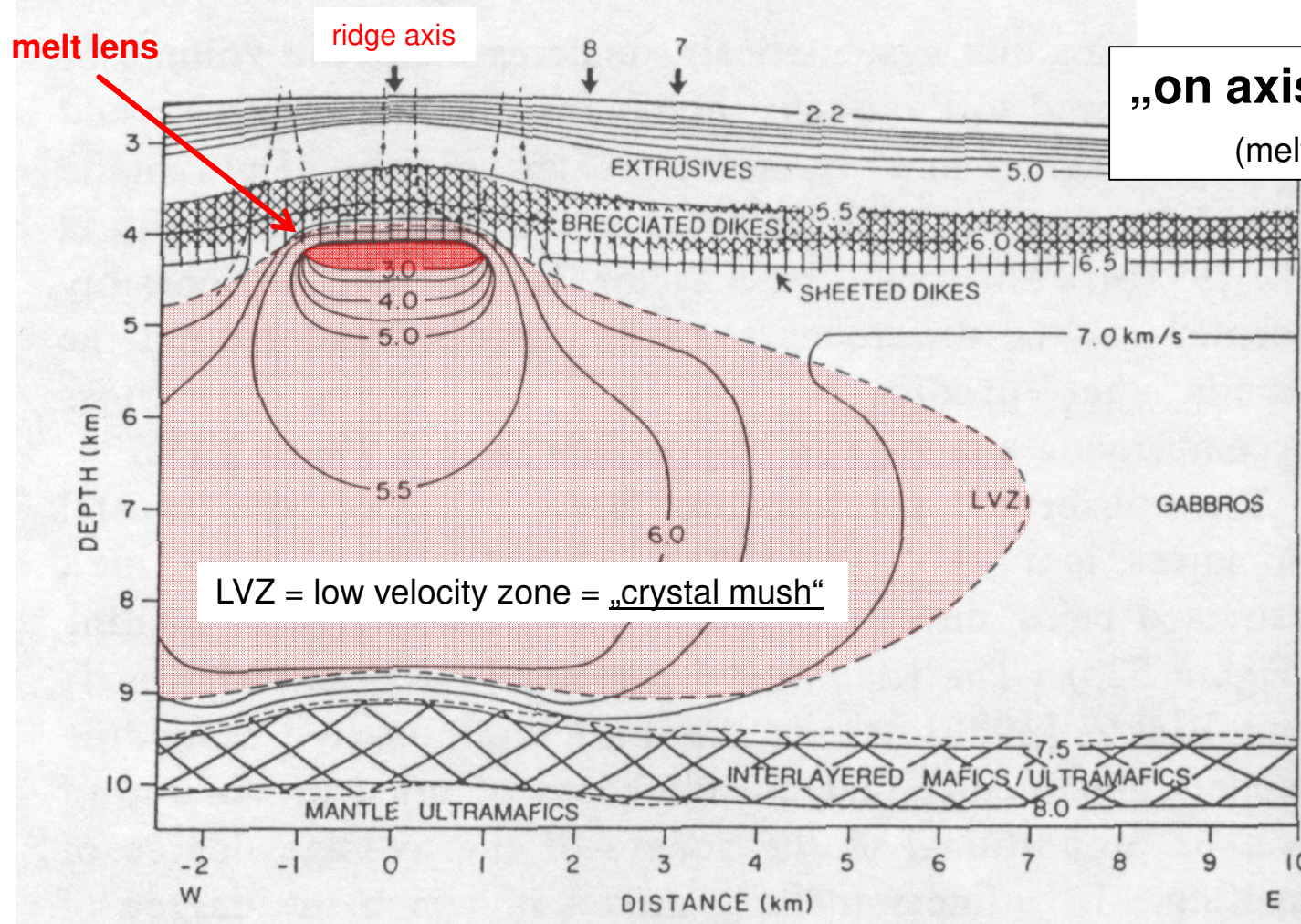
„off axis profile“  
(no melt lens/magma chamber!)

These ,ultramafic cumulates‘ are sometimes **wehrlites**, but mostly **residual dunites**.

But note, that these ,ultramafic cumulates‘ are often **NO cumulates** but **residual rocks** or **olivine mush** dragged by basaltic melt!

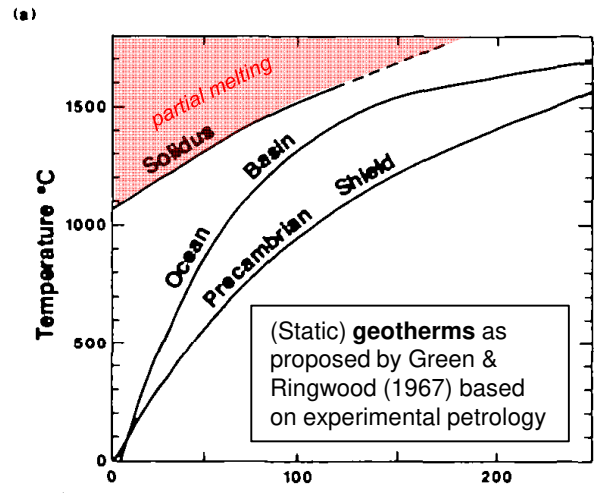
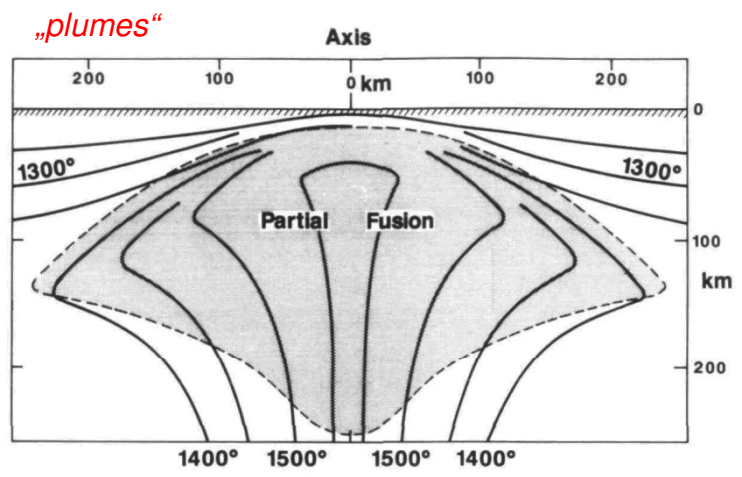
## 2D seismic structure at the East Pacific Rise (EPR)

Velocity contours perpendicular to the ridge axis

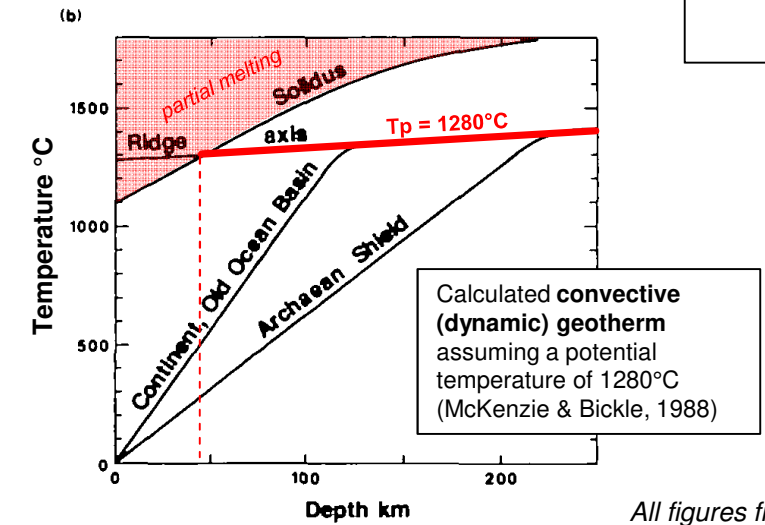
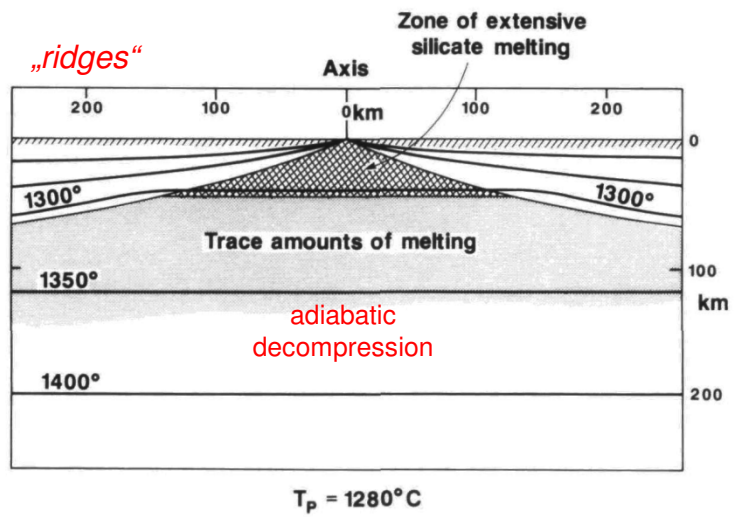


EPR is a fast spreading ridge

# Thermal state of the (oceanic) lithosphere and underlying asthenosphere



Using primitive **MORB glasses** and the thickness of the **oceanic crust**, yields a mantle potential **temperature of ~1280°C**, corresponding to a „melt thickness“ of **7 km**

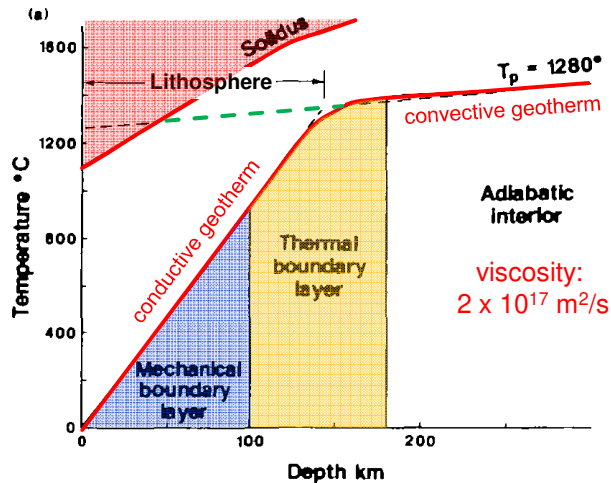


All figures from: McKenzie & Bickle, 1988. J. Petrol., 29, 625-679

# Thermal state of the (oceanic) lithosphere and underlying asthenosphere

TBL: isotopic **homogeneous**  
(i.e. convecting)

MBL: isotopic **heterogeneous**  
(i.e. non-convecting)



Melting the mantle?  
Temperature is, what matters!

$$T_p = T \exp\left(-\frac{g\alpha z}{C_p}\right)$$

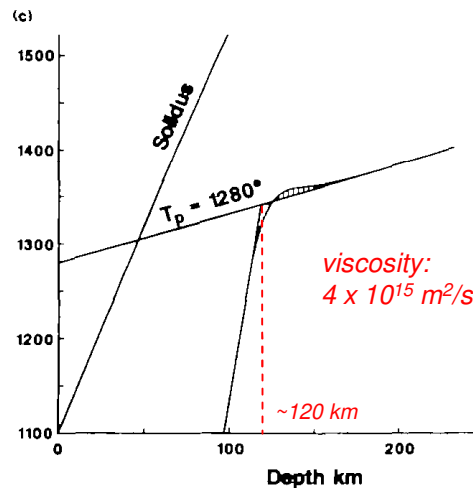
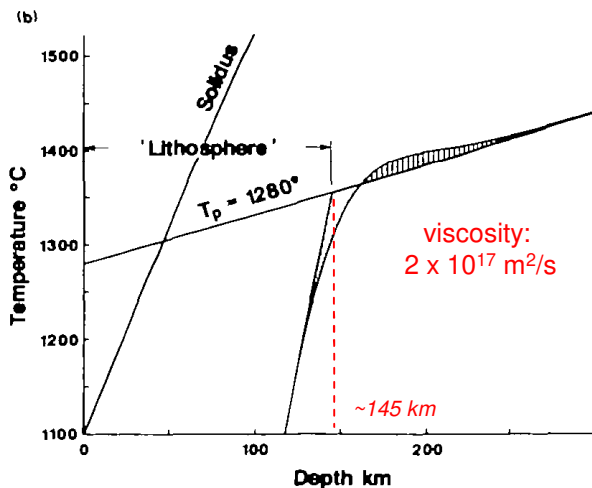
adiabatic  
temperature  
gradient

$$\left(\frac{\partial T}{\partial z}\right)_s = \frac{g\alpha_f T}{C_p}$$

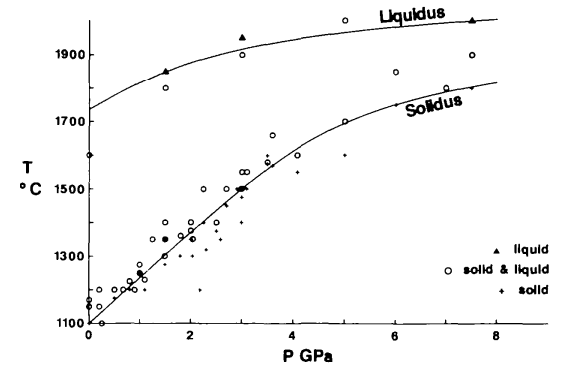
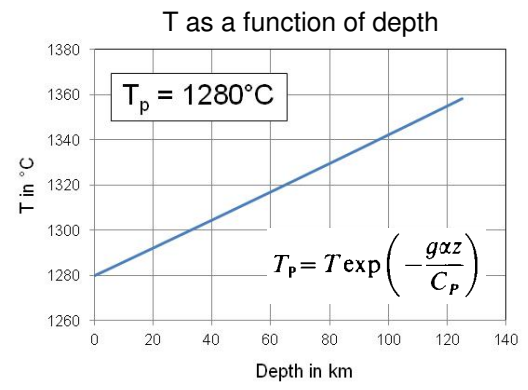
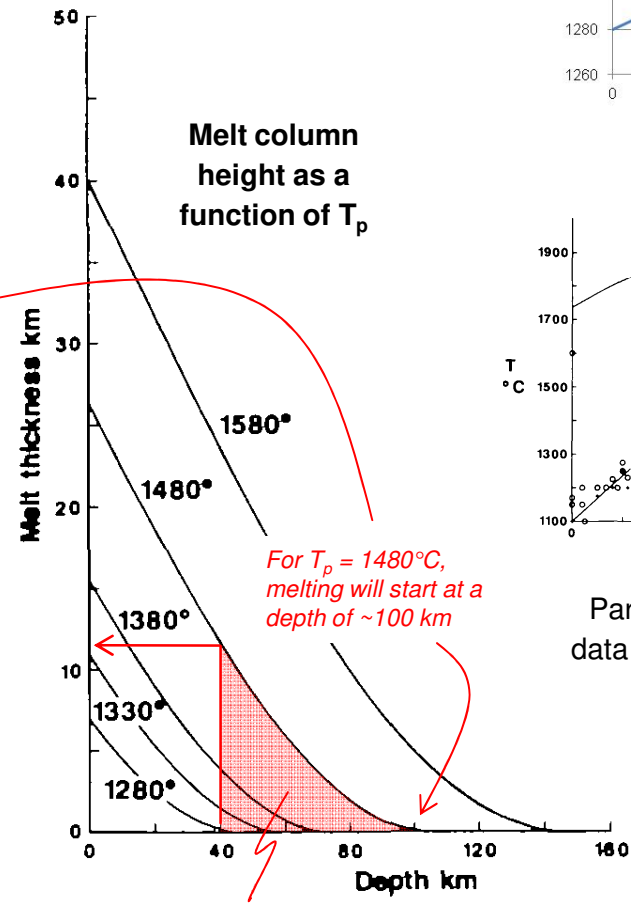
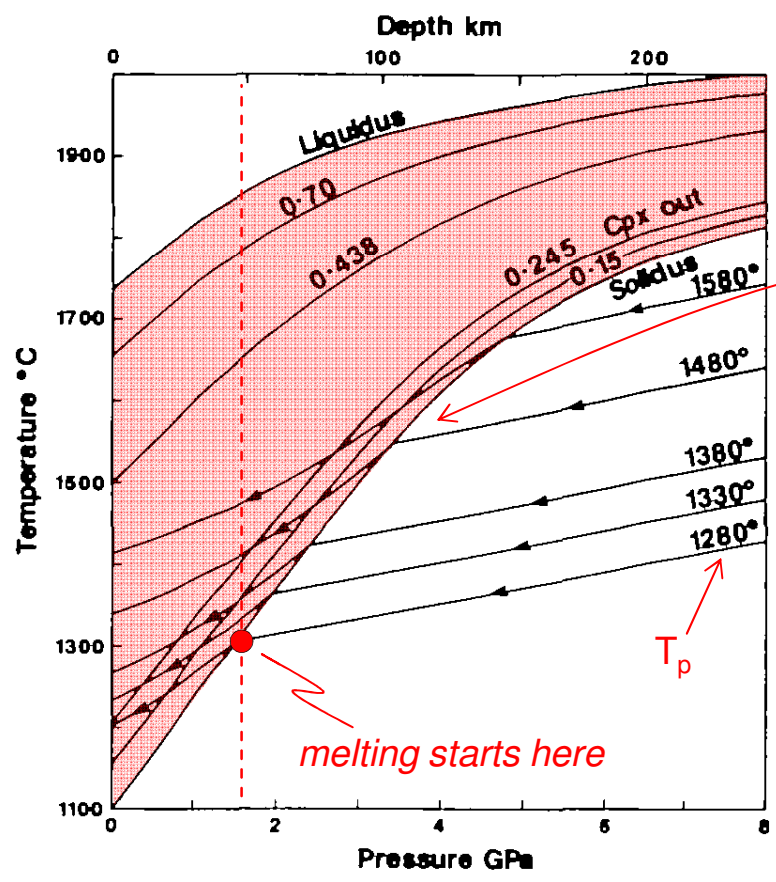
$$\left(\frac{\partial T}{\partial z}\right)_s \simeq 1^\circ\text{C km}^{-1}$$

for a melt!

$g$  = acceleration due to gravity,  
 $\alpha$  = thermal expansion coefficient,  
 $z$  = depth  
 $C_p$  = specific heat,  
 $T$  = absolute temperature,  
 $T_p$  = potential temperature



# Adiabatic decompression melting and melt volume as a function of $T_p$

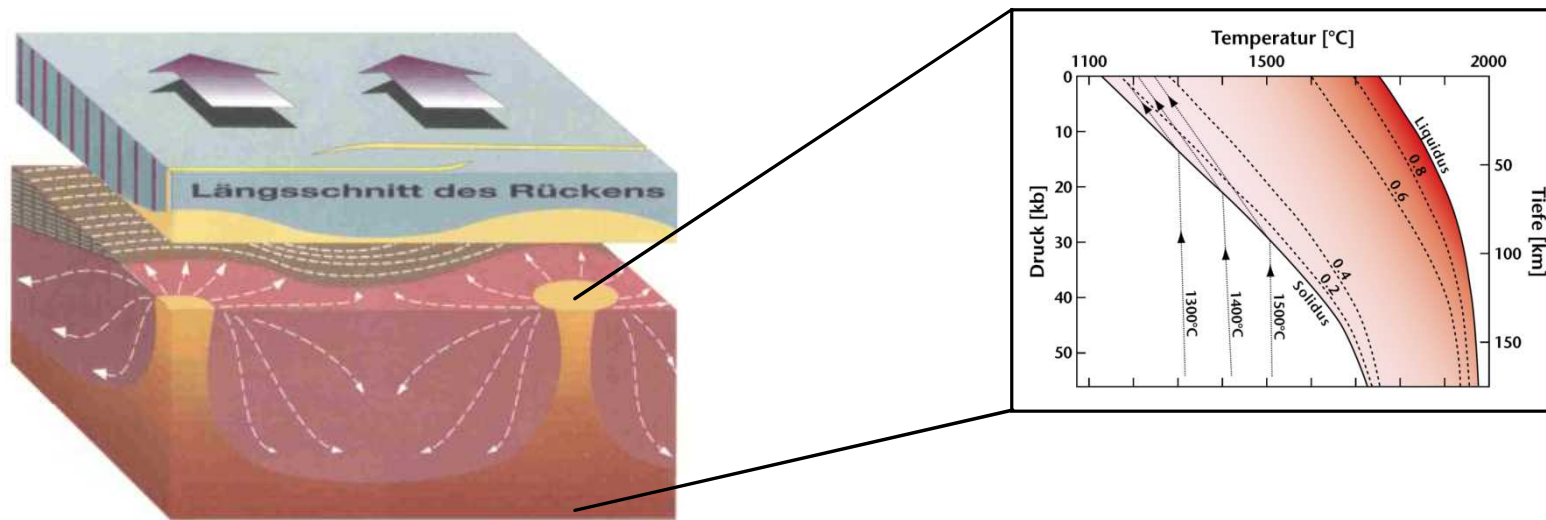


Parameterization of experimental data to obtain **solidus** and **liquidus** temperatures

From: McKenzie & Bickle, 1988. J. Petrol., 29, 625-679

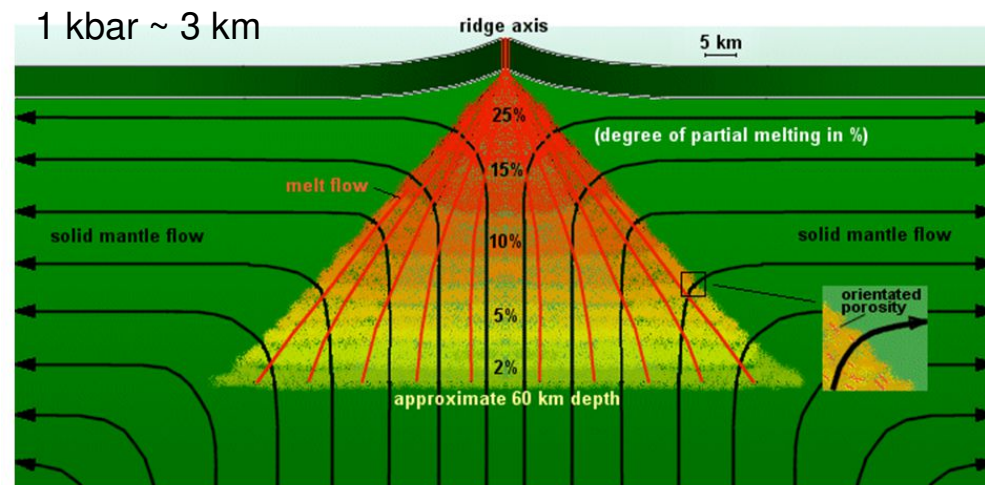
Example: If melting starts at ~100 km depth (this is the case for  $T_p = 1480^\circ\text{C}$ ), and ends at ~40 km depth, this produces a melt column of about 12 km thickness

# Oceanic asthenosphere: Melting in „diapirs“



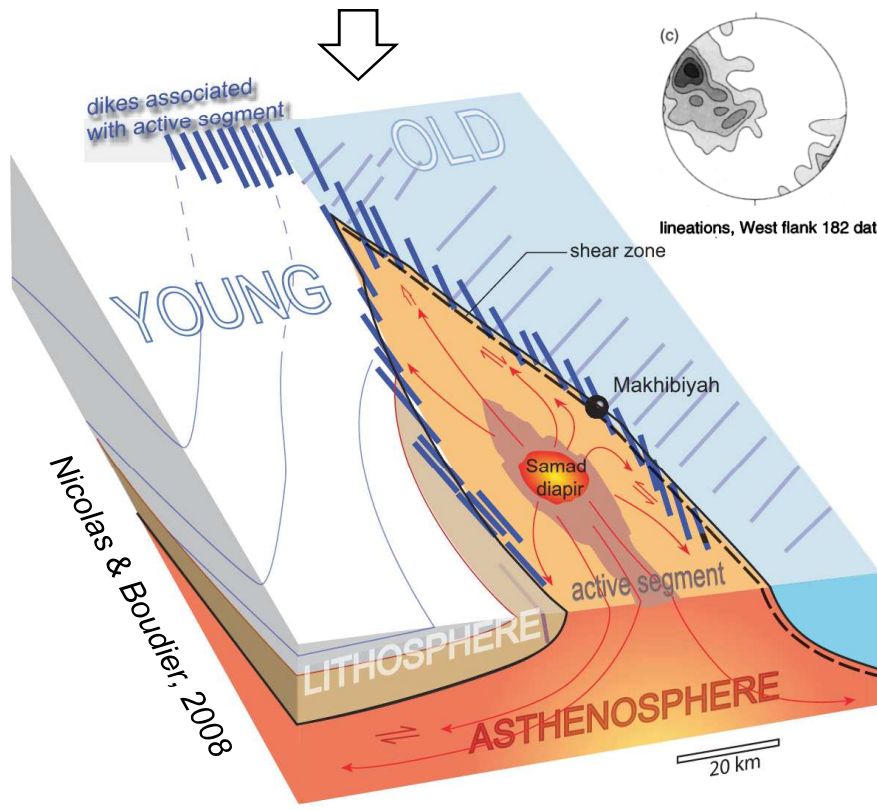
Melting beneath mid-ocean ridges

Upwelling and **decompression melting** likely does occur in „**diapirs**“, not along linear structures

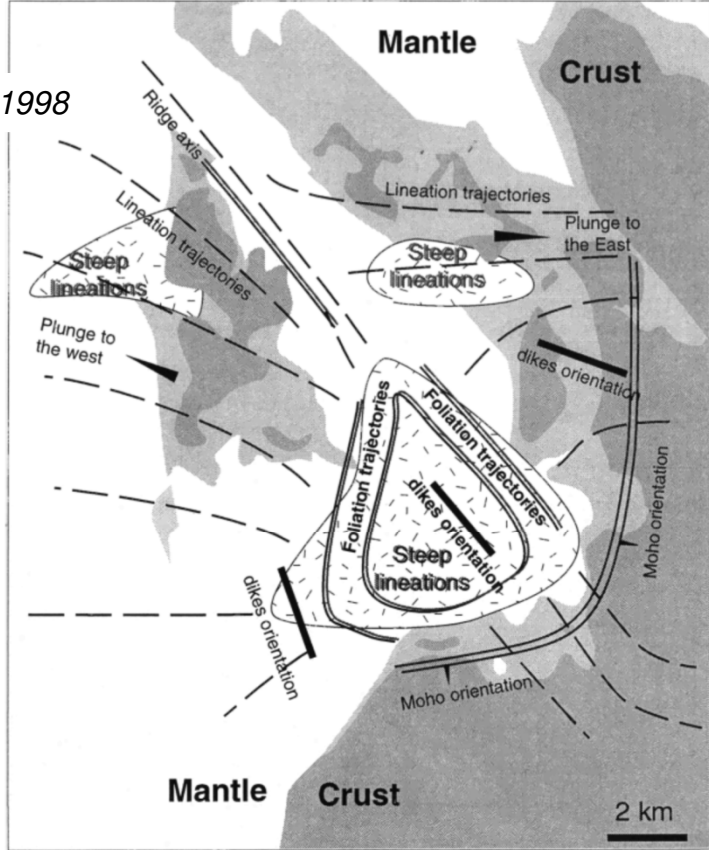
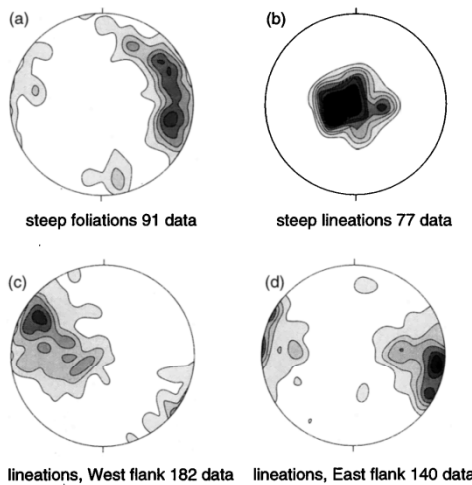


# Melting occurs in up-welling „diapirs“

Mantle diapir and propagating ridge segment (structural reconstruction, Oman ophiolite)



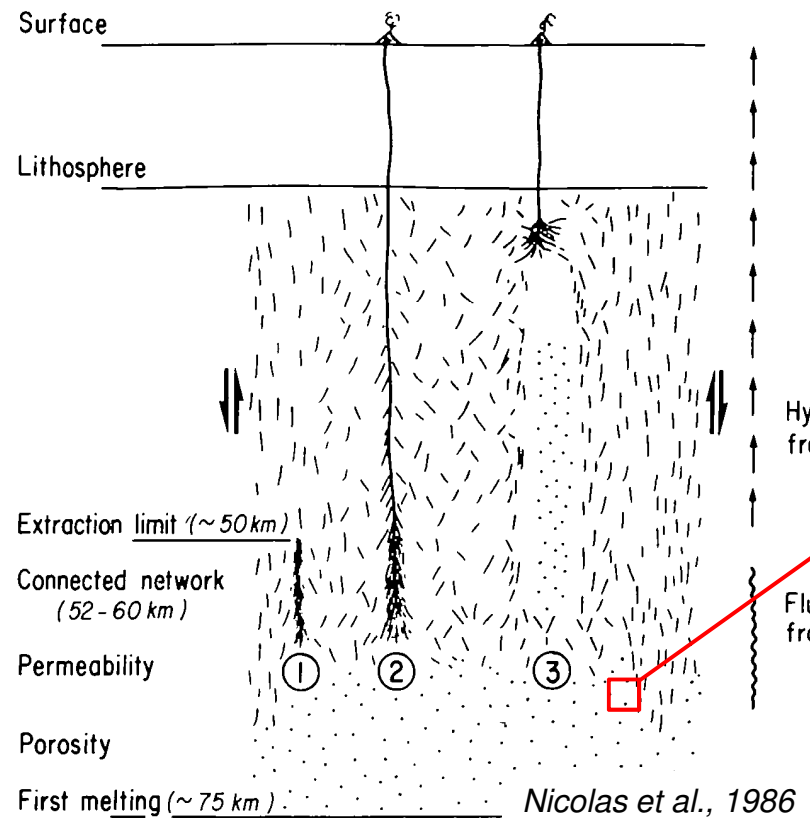
Jousselin et al., 1998



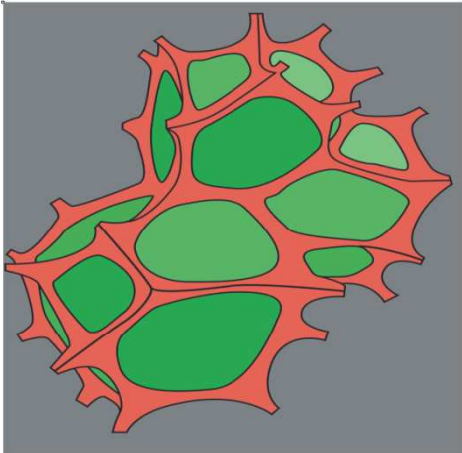
Maqusad mantle diapir and inferred ridge axis (Oman ophiolite) as reconstructed from structural data (foliation planes and lineations in mantle peridotite and lower gabbros)

# How does the melt leave the source (mantle)?

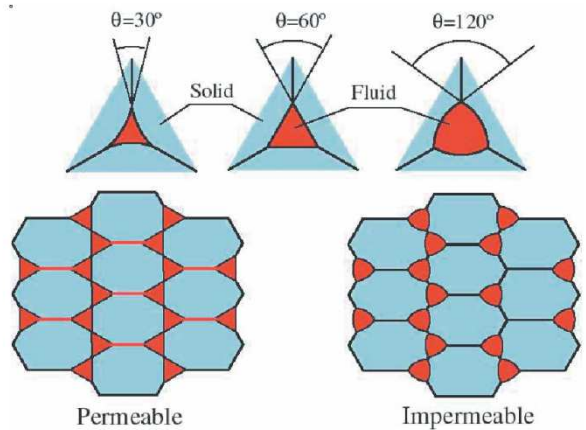
Porous flow vs. hydraulic fracturing



Nicolas et al., 1986



Schematic drawing of an interconnected melt-network



**Three stages of melt extraction:** 1) Formation of a connected melt network, 2) melt overpressure and hydrofracturing, 3) dying conduit

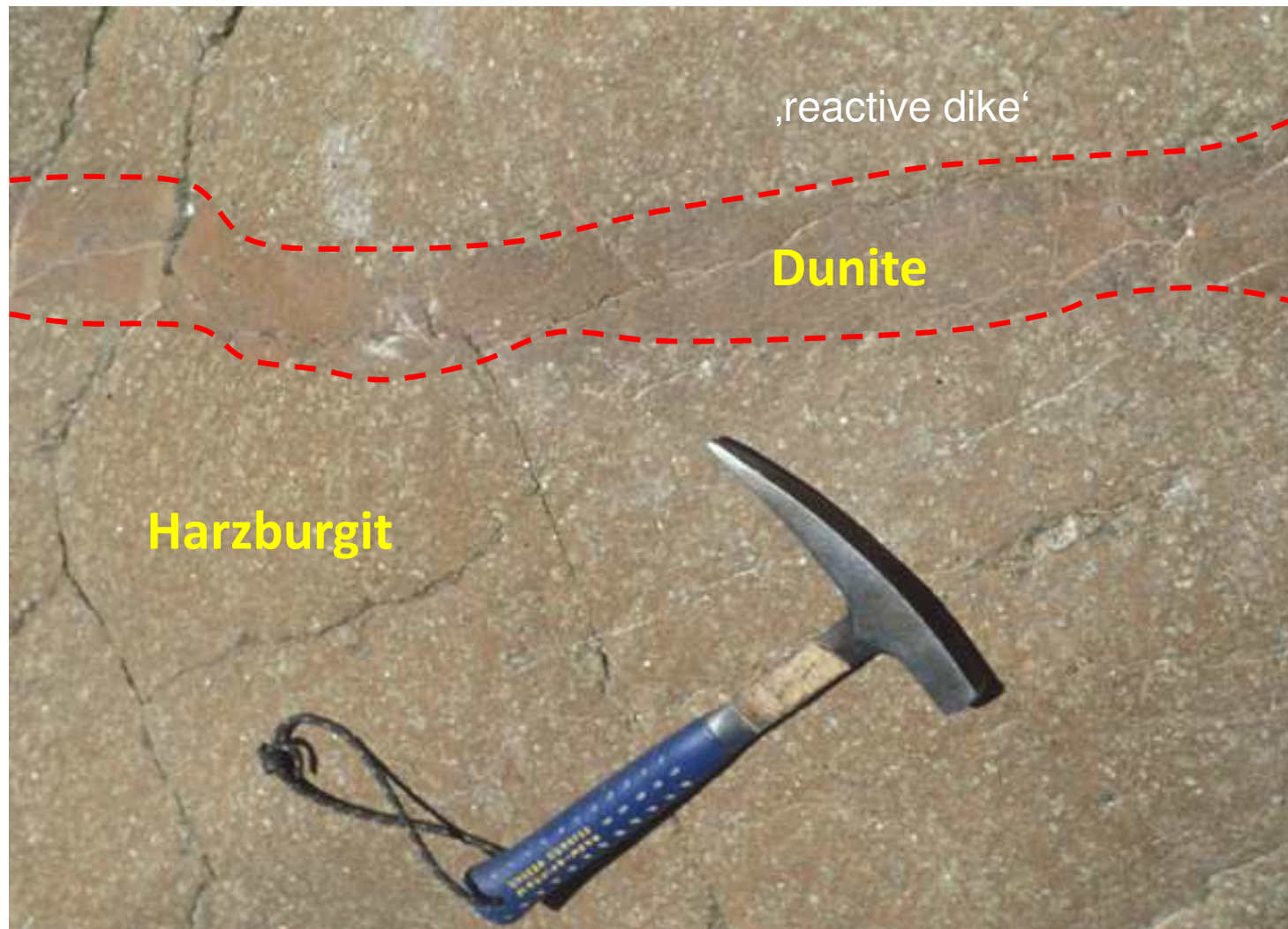
## How does the melt leave the source (mantle)?

Reactive porous flow – Oman ophiolite



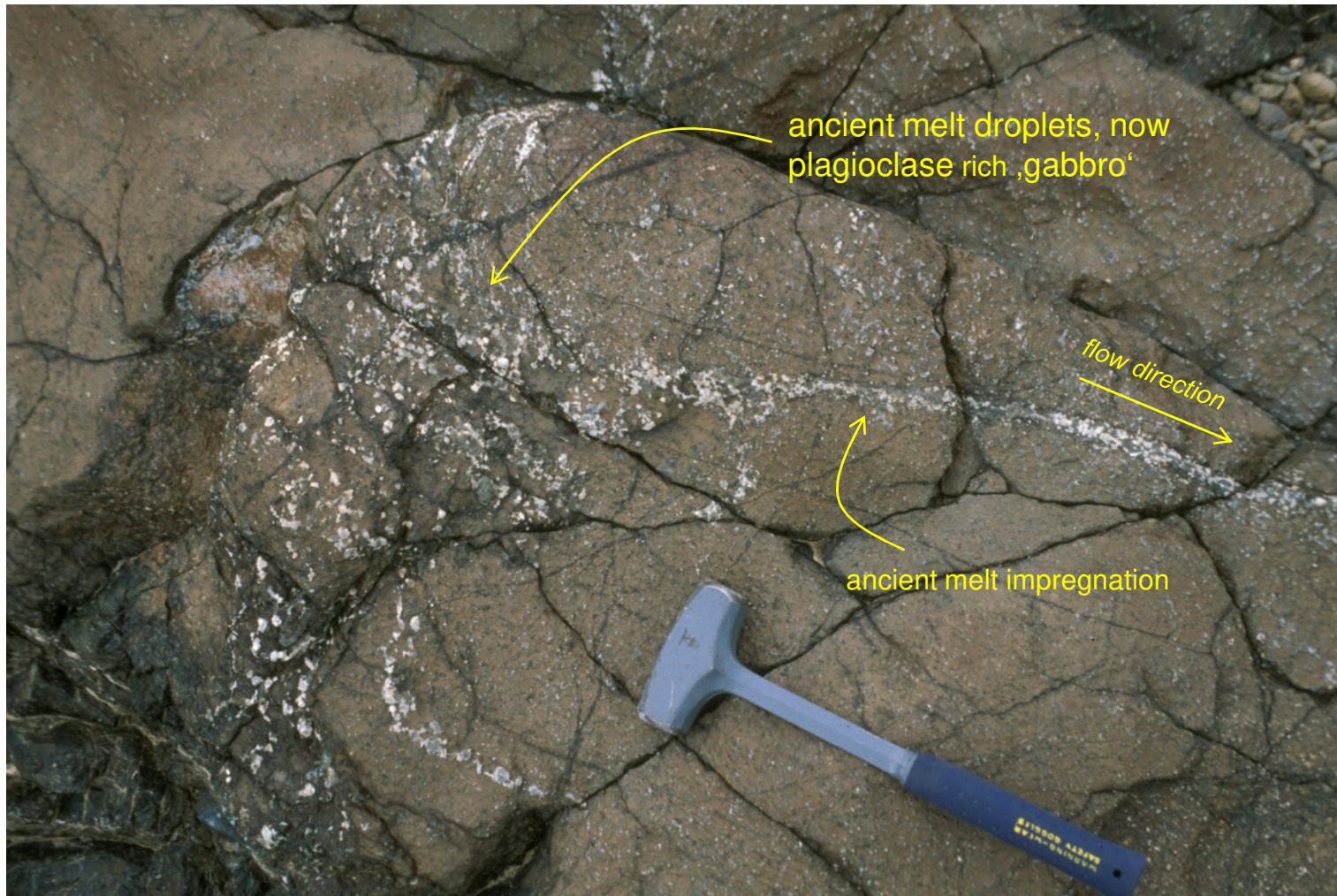
## How does the melt leave the source (mantle)?

Residual dunite – Oman ophiolite



## How does the melt leave the source (mantle)?

Transition to hydraulic fracturing – Oman ophiolite



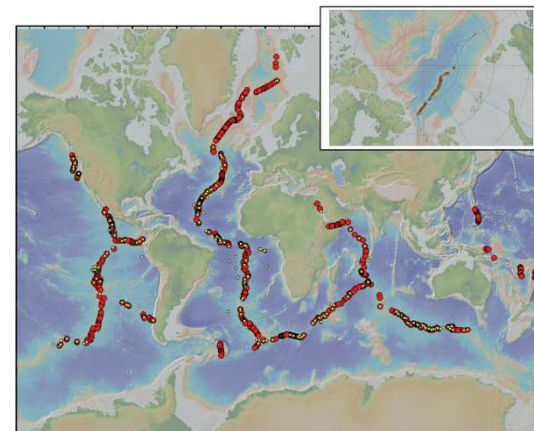
## Melt compositions at mid ocean ridges - Major elements

Table 4. Primitive MORB compositions.

	Iceland <sup>1</sup>	Global Range <sup>2</sup>
SiO <sub>2</sub>	48.72	47.2 – 51.4
TiO <sub>2</sub>	0.93	0.50 – 1.29
Al <sub>2</sub> O <sub>3</sub>	15.99	15.0 – 17.58
FeO	8.96	7.21 – 8.97
MnO	0.16	0.06 – 0.21
MgO	10.56	8.39 – 10.90
CaO	14.01	11.56 – 13.35
Na <sub>2</sub> O	1.72	1.62 – 2.49
K <sub>2</sub> O	0.05	0.00 – 0.28

39 samples

*Presnall et al., 2002*



430 samples

<sup>1</sup> Most magnesian published Icelandic glass (Breddam, 2002).  
<sup>2</sup> 39 primitive MORB glasses from Atlantic and Pacific Oceans (Presnall and Hoover, 1987).

**Table 1.** The Composition of ALL MORB<sup>a</sup>

	n	ALL MORB Mean <sup>b</sup>	± (95% conf)	ALL MORB Log-Normal Mean <sup>c</sup>	± (95% conf)	Upper Bound	Lower Bound
MgO	430	<b>7.58</b>	0.12	7.69	0.08	9.04	6.21
SiO <sub>2</sub>	430	<b>50.47</b>	0.08	50.41	0.08	51.75	49.22
FeO	430	<b>10.43</b>	0.21	10.07	0.14	13.48	8.12
CaO	430	<b>11.39</b>	0.09	11.35	0.07	12.49	10.15
Na <sub>2</sub> O	430	<b>2.79</b>	0.03	2.76	0.04	3.69	2.06
Al <sub>2</sub> O <sub>3</sub>	430	<b>14.70</b>	0.12	14.95	0.10	16.74	13.25
TiO <sub>2</sub>	430	<b>1.68</b>	0.05	1.54	0.03	2.32	1.06
K <sub>2</sub> O	430	<b>0.160</b>	0.014	0.144	0.011	0.558	0.045
P <sub>2</sub> O <sub>5</sub>	409	<b>0.184</b>	0.012	0.169	0.008	0.400	0.082
MnO	379	<b>0.184</b>	0.005	0.173	0.004	0.231	0.105

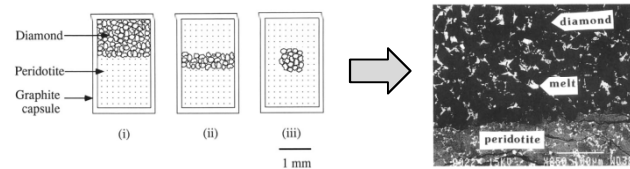
*Gale et al., 2013, The mean composition of ocean ridge basalts. Geochemistry, Geophysics, Geosystems.*

>> This publication also provides **trace-element** and **isotope** data!

## Melt compositions at mid ocean ridges

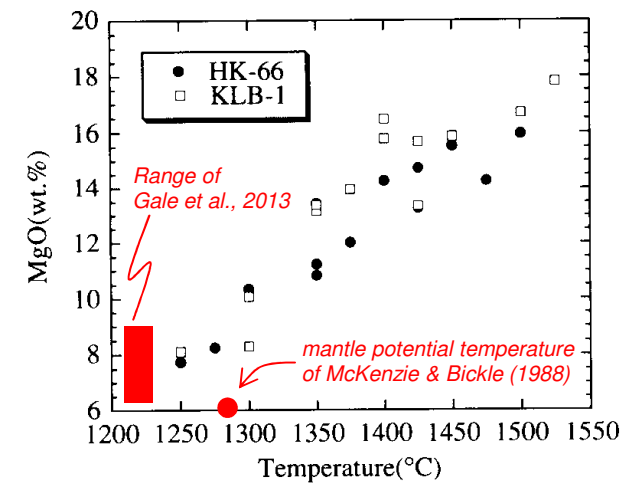
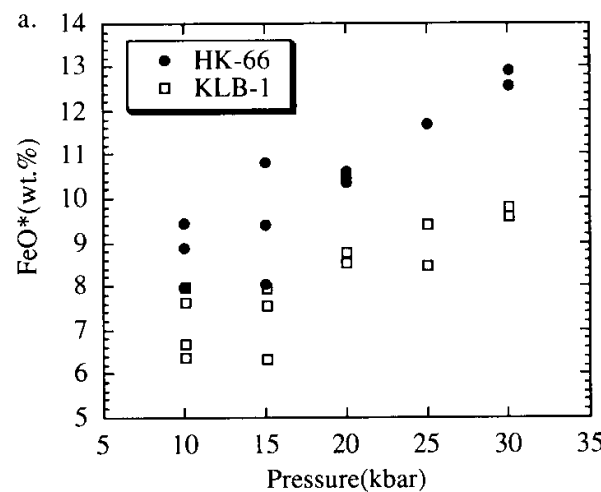
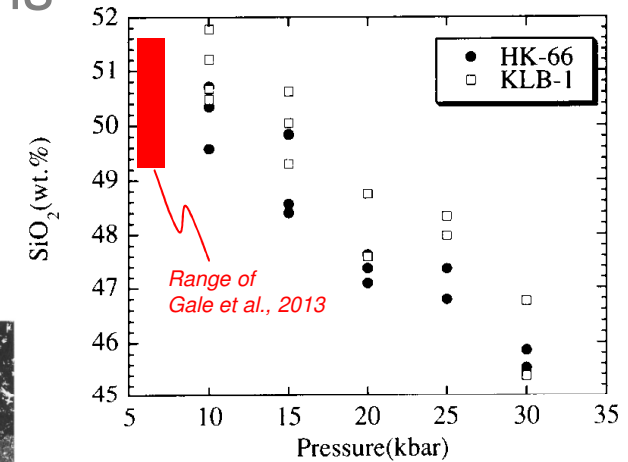
Experimental constraints on p-T conditions

Melting at MOR's is basically **dry** and at relatively **low pressure** (thin lithosphere!)



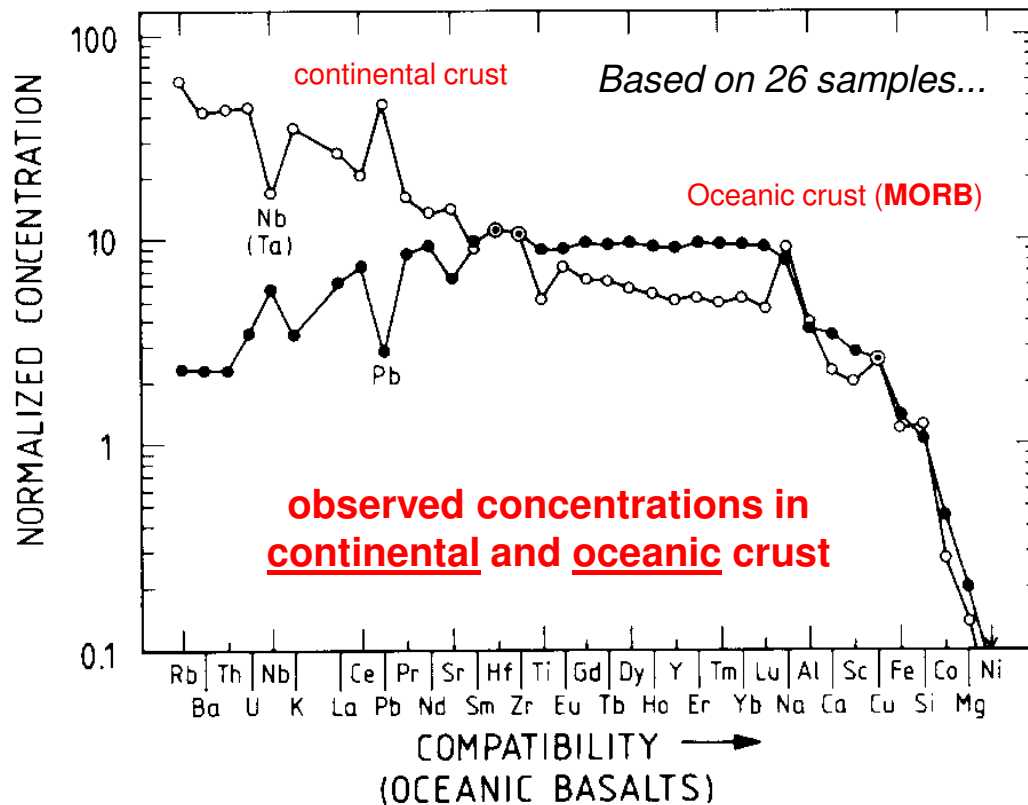
Modal composition of KLB-1 (wt%):  
ol: 58, opx: 25, cpx: 15, sp: 2

	KLB-1	HK-66
SiO <sub>2</sub>	44.48	48.02
TiO <sub>2</sub>	0.16	0.22
Al <sub>2</sub> O <sub>3</sub>	3.59	4.88
FeO*	8.10	9.90
MnO	0.12	0.14
MgO	39.22	32.35
CaO	3.44	2.97
Na <sub>2</sub> O	0.30	0.66
K <sub>2</sub> O	0.02	0.07
P <sub>2</sub> O <sub>5</sub>	0.03	0.07
Cr <sub>2</sub> O <sub>3</sub>	0.31	0.25
NiO	0.25	n.d.
Total	100.02	99.53



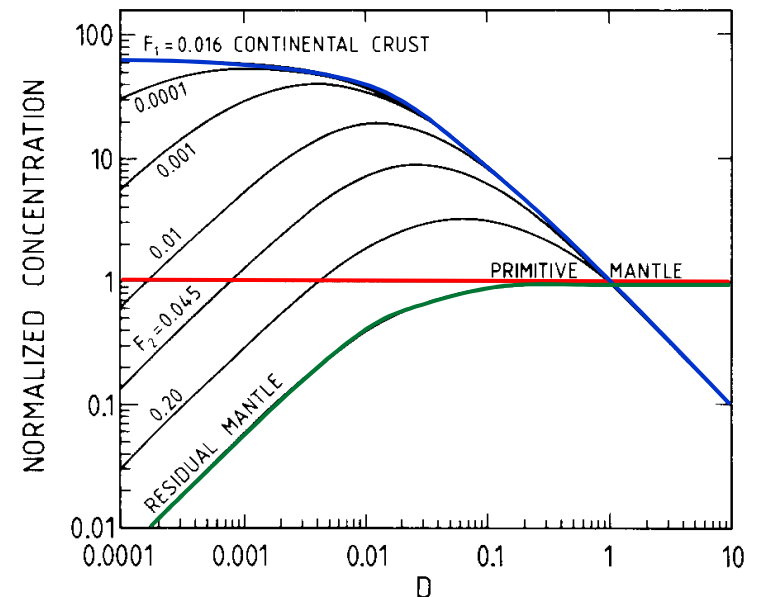
## Melt compositions at mid ocean ridges

Trace element compositions, early work, principal implications



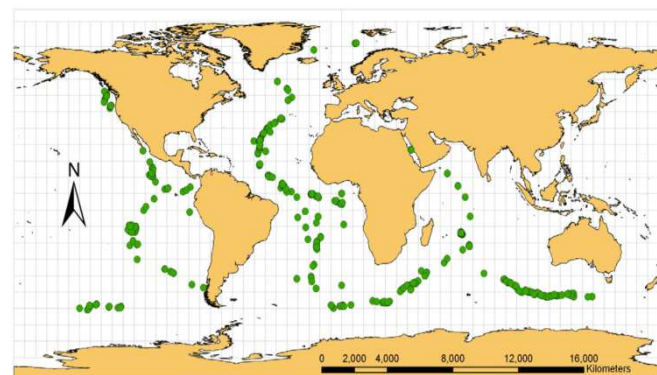
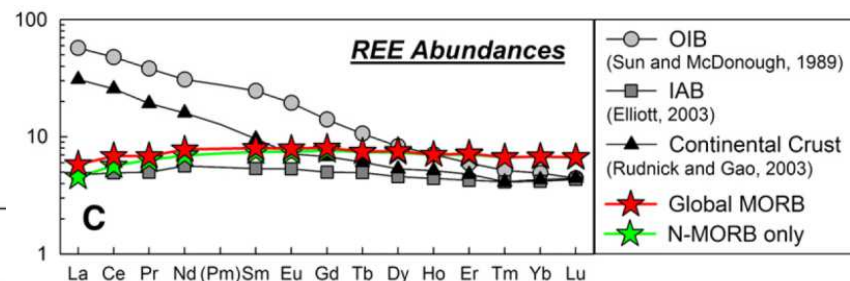
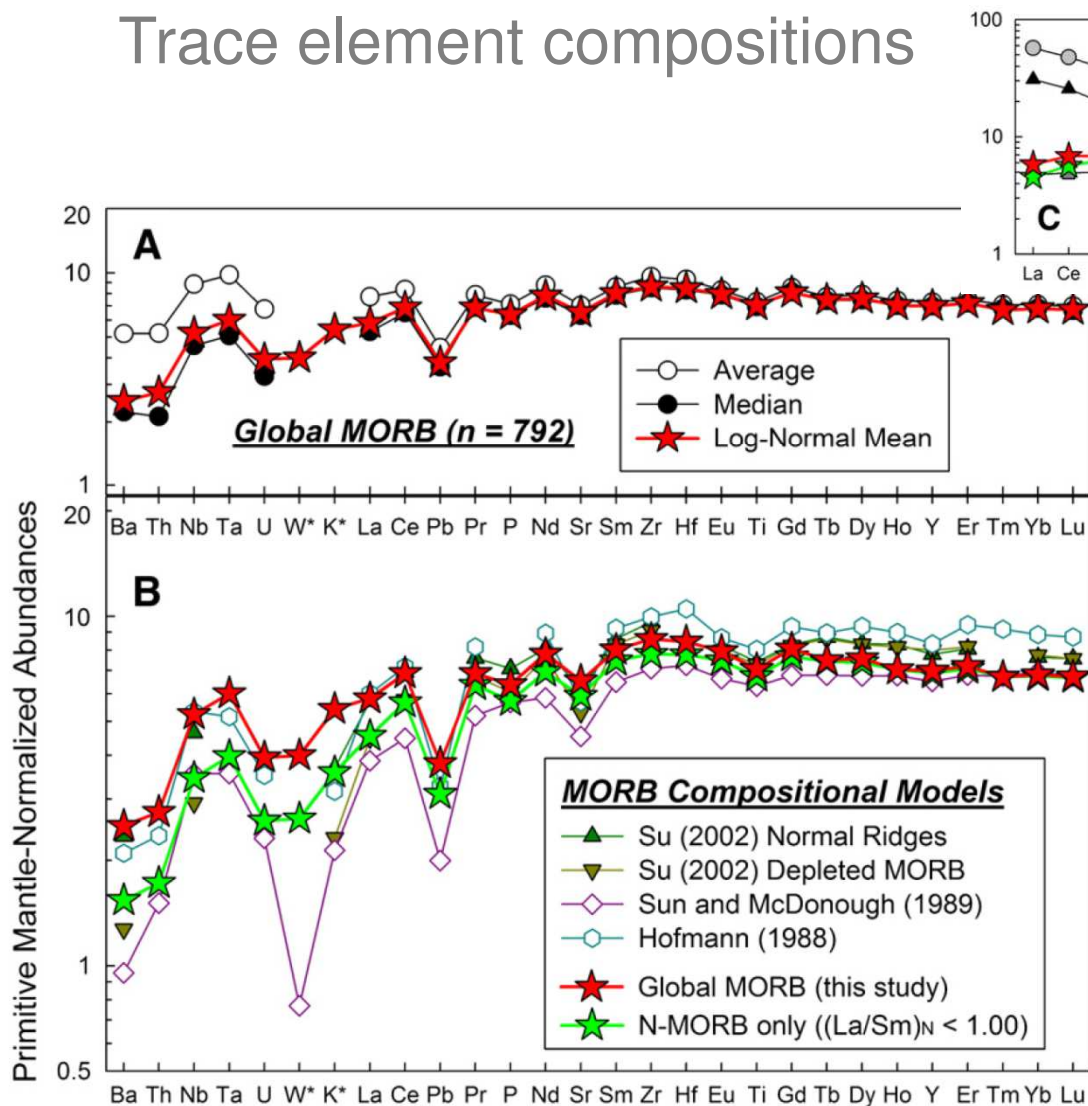
Hofmann, 1988: *Chemical differentiation of the Earth: the relationship between mantle, continental crust, and oceanic crust.* *Earth Planet. Sci. Lett.*, 90, 297-314.

Explained by the extraction of **1.5% melt** from **PRIMA** to form the **CC**, and subsequent **8-10% melt** from **DM** to form the **OC**



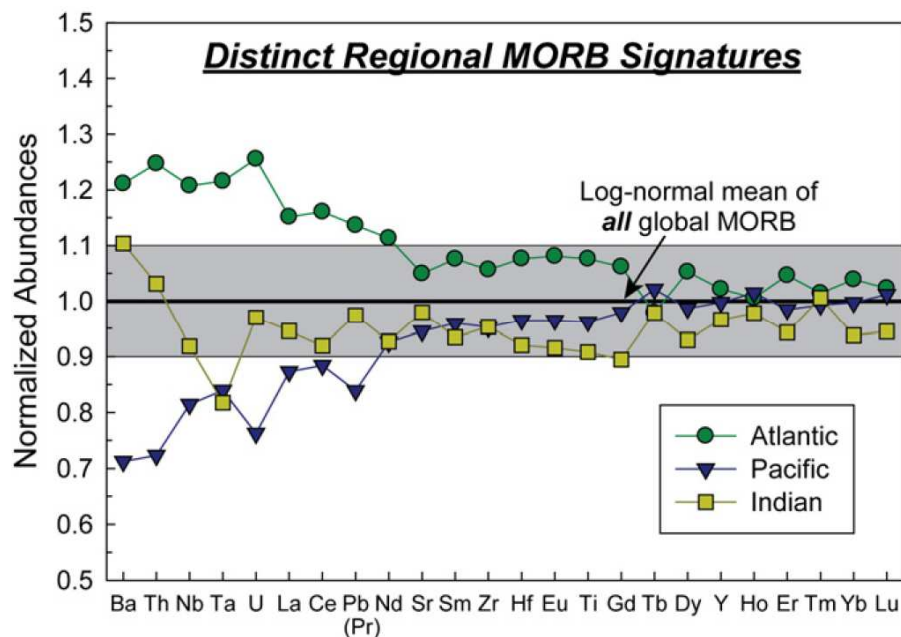
## Melt compositions at mid ocean ridges

Trace element compositions



## Melt compositions at mid ocean ridges

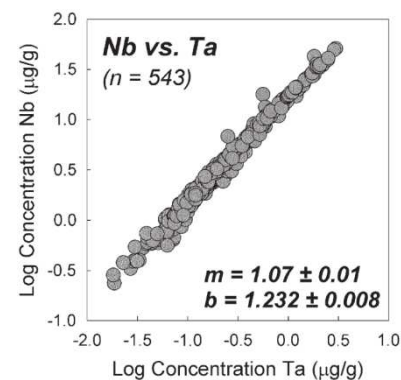
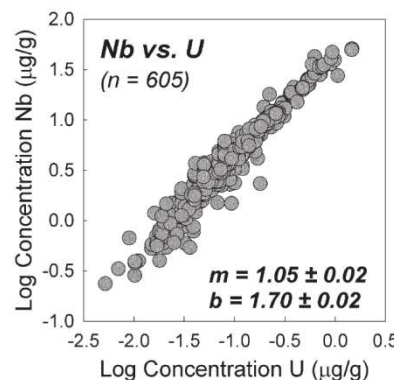
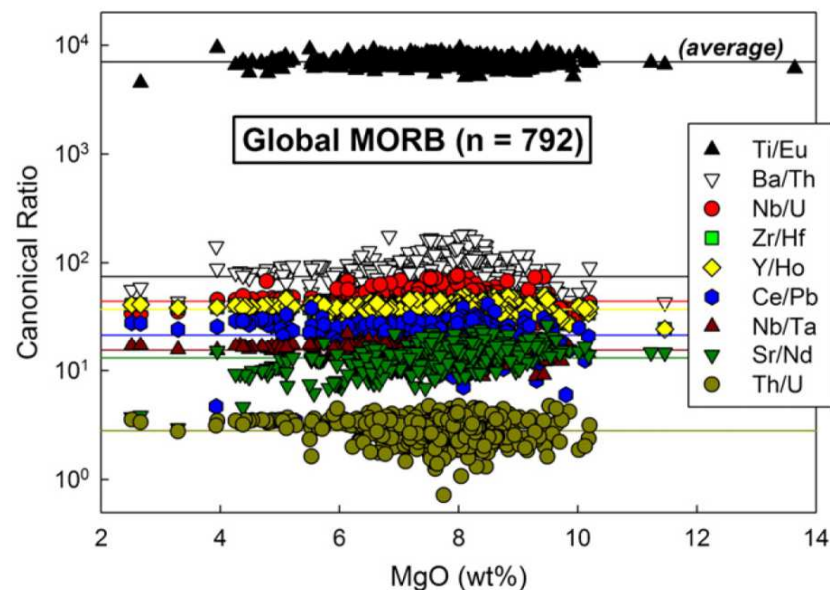
### Trace element compositions



▲ Differences between Atlantic, Pacific and Indian MORB

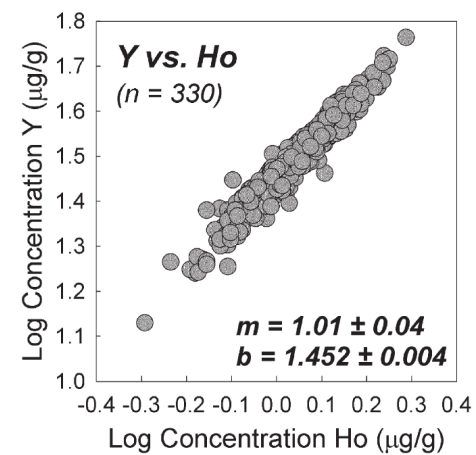
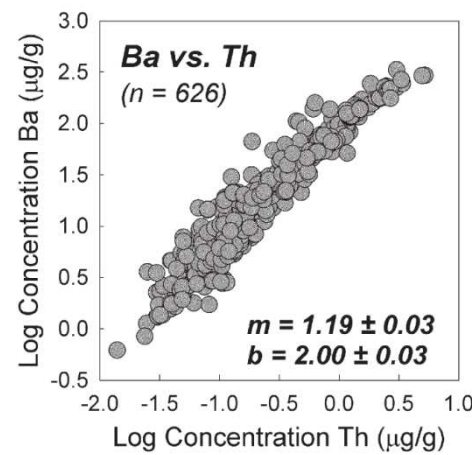
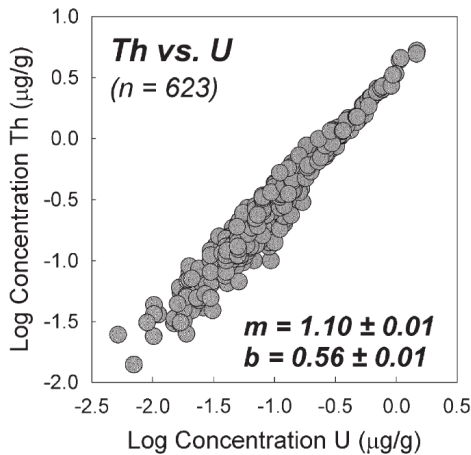
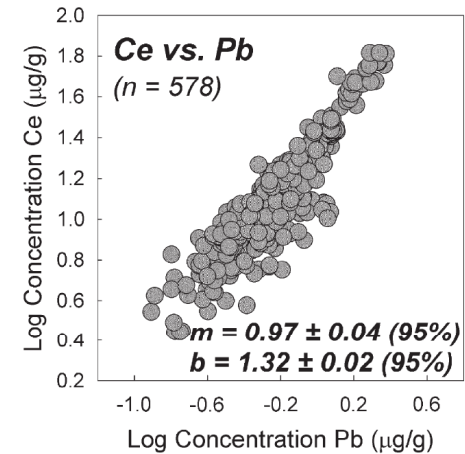
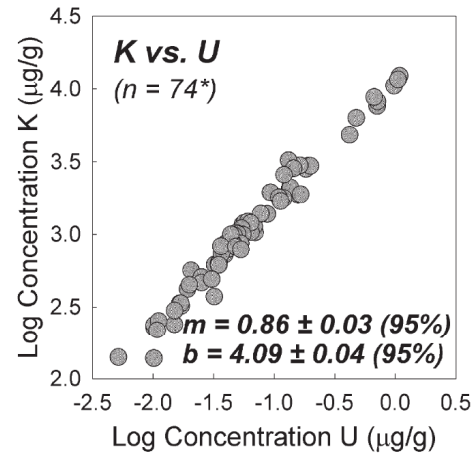
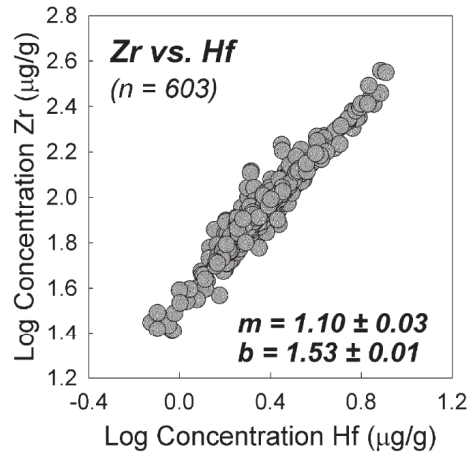
Slopes of **unity** in bivariate concentration plots indicate identical D-values (i.e. identical behaviour) ►

„Constancy“ of distinct ratios



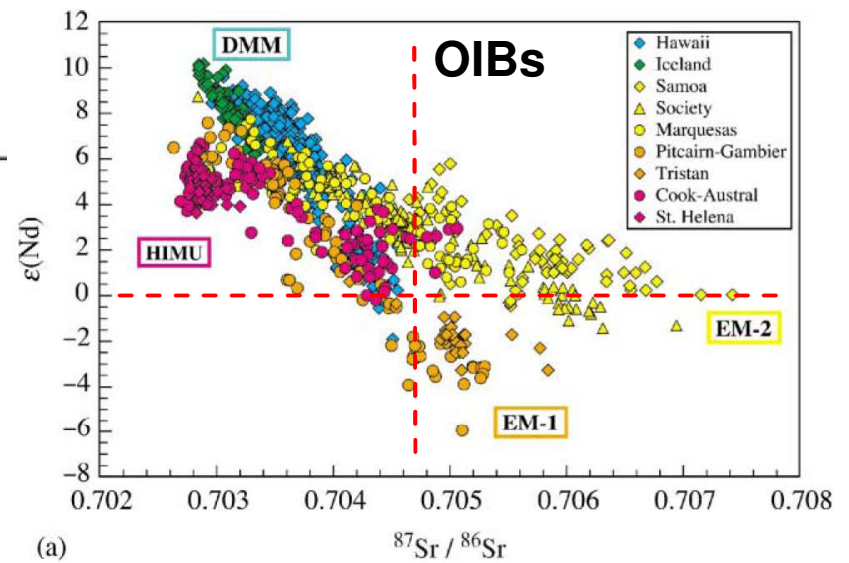
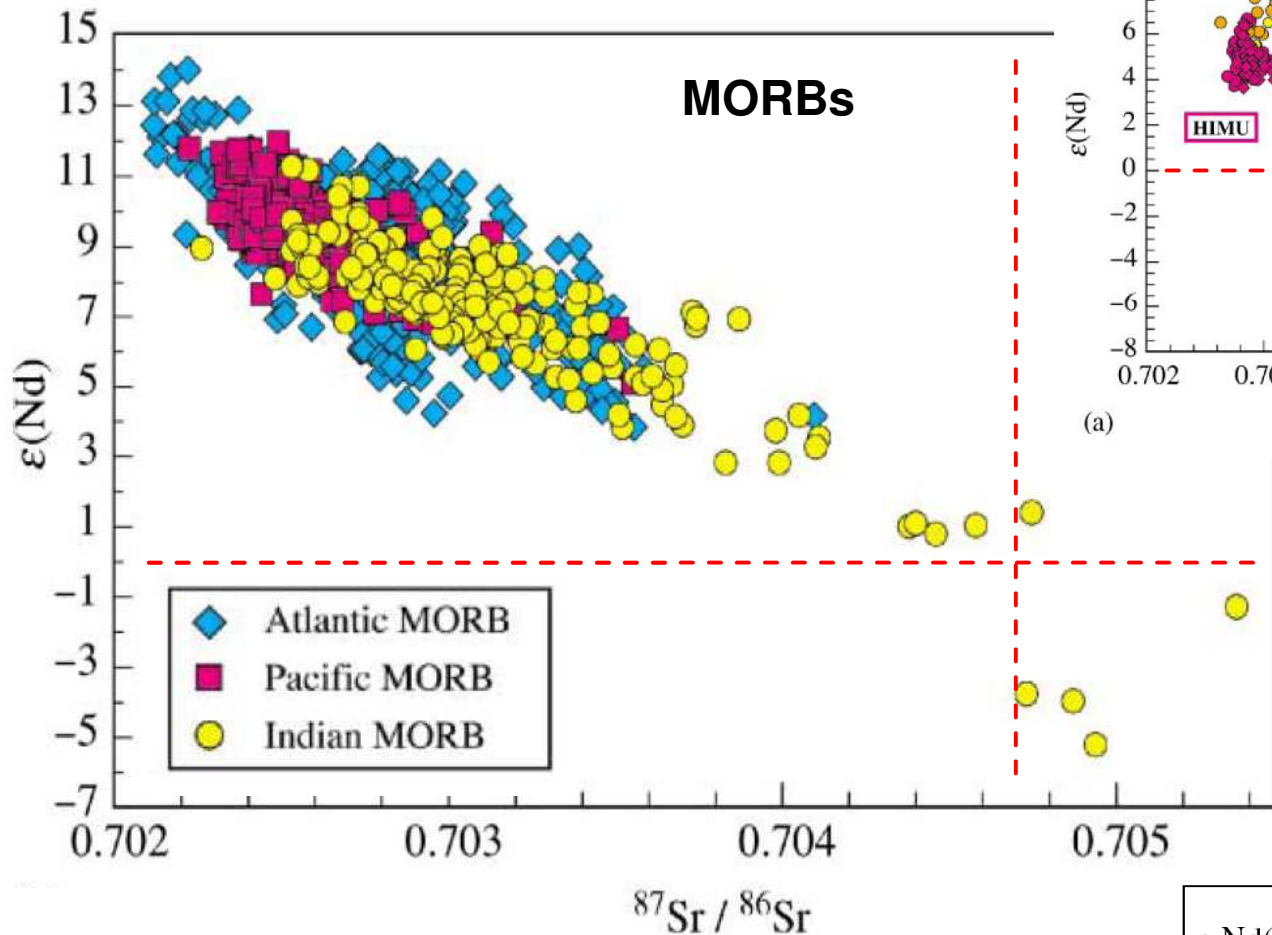
# Melt compositions at mid ocean ridges

## Trace element compositions



# Melt compositions at mid ocean ridges

Isotope compositions

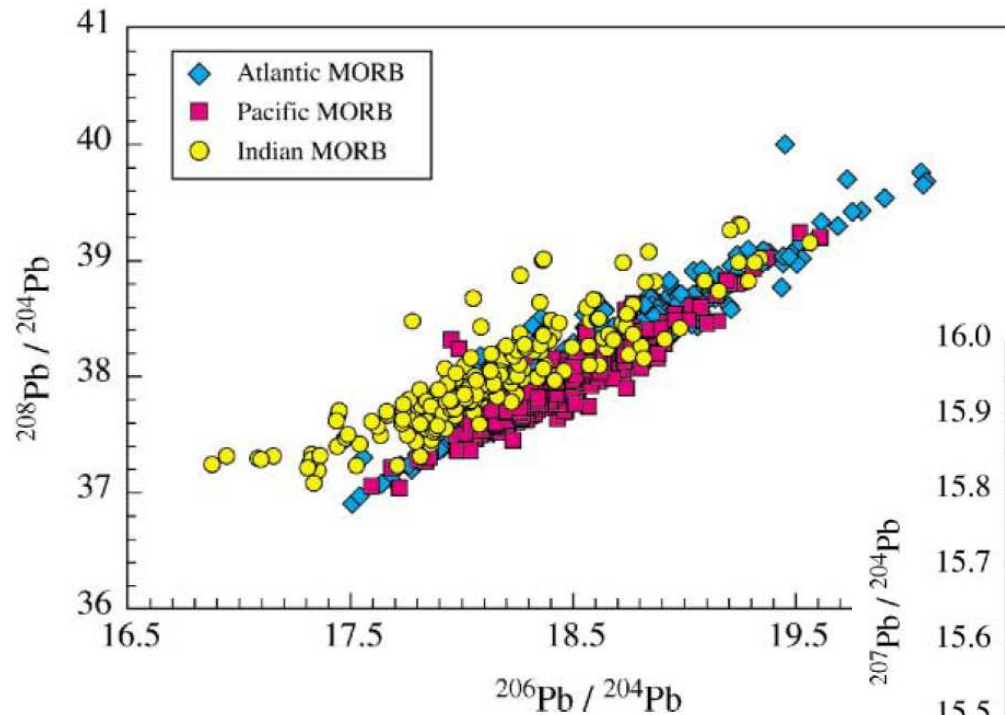


$^{87}\text{Rb}$  (48.8 Ga)  $^{87}\text{Sr}$   
 $^{147}\text{Sm}$  (106 Ga)  $^{143}\text{Nd}$

$$\epsilon \text{ Nd}(t) = \left( \frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}}(t)}{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}(t)} - 1 \right) \times 10^4$$

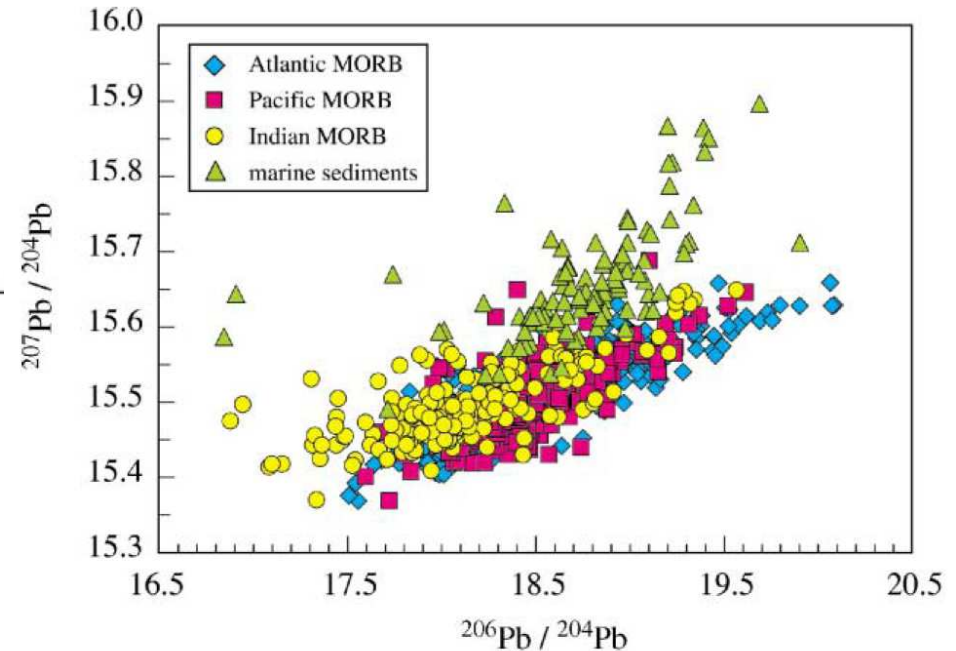
# Melt compositions at mid ocean ridges

## Isotope compositions



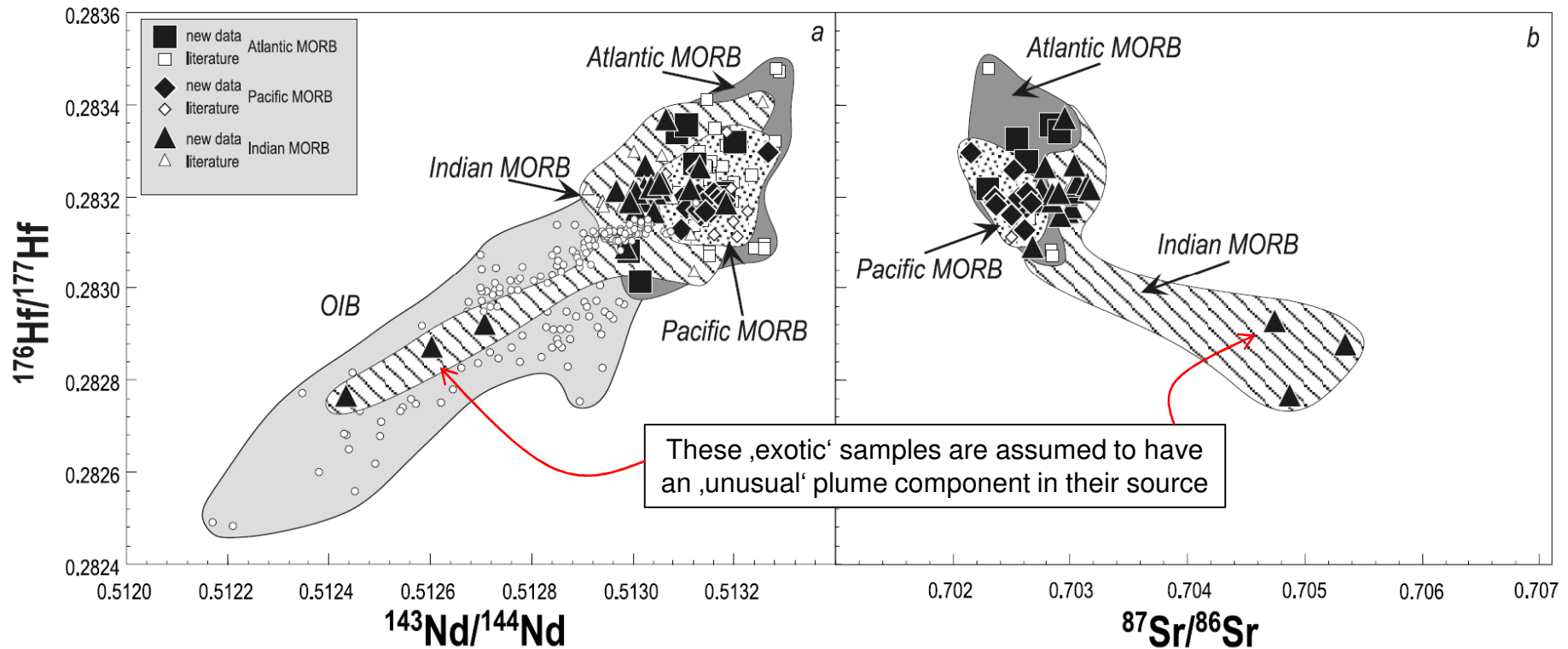
No distinct differences exist, but Indian MORB seems to be slightly enriched in  $^{208}\text{Pb}$

$^{238}\text{U}$  (4.47 Ga)  $^{206}\text{Pb}$   
 $^{235}\text{U}$  (0.738 Ga)  $^{207}\text{Pb}$   
 $^{232}\text{Th}$  (14.0 Ga)  $^{208}\text{Pb}$



# Melt compositions at mid ocean ridges

## Hf-Nd isotope compositions

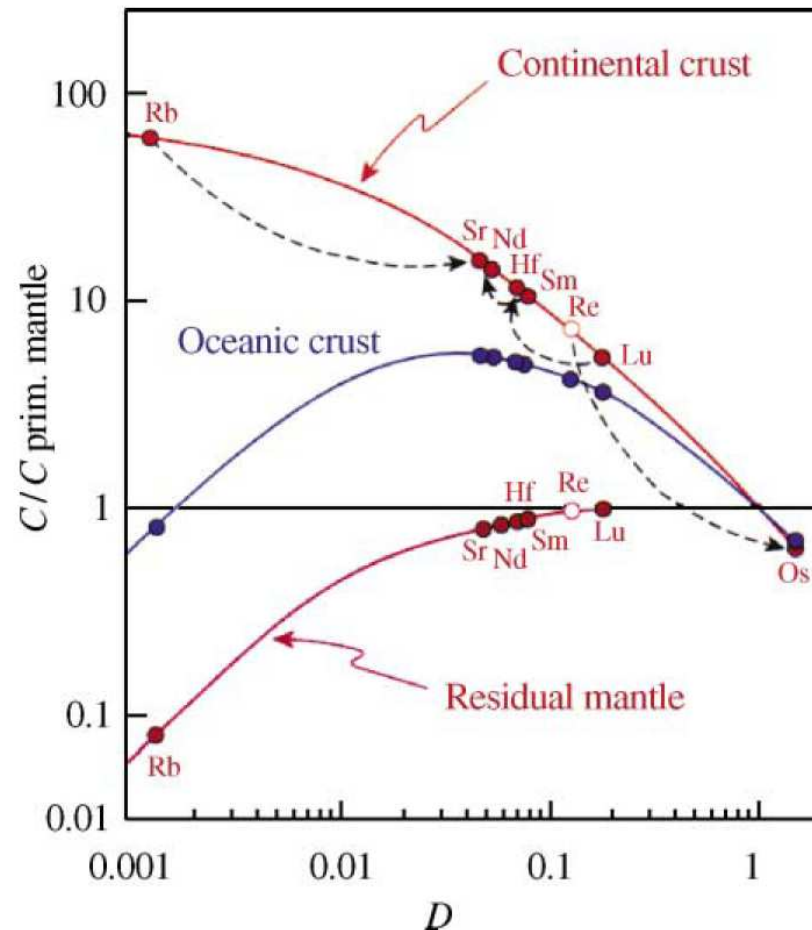


$^{87}\text{Rb}$  (48.8 Ga)  $^{87}\text{Sr}$   
 $^{176}\text{Lu}$  (35.7 Ga)  $^{176}\text{Hf}$   
 $^{147}\text{Sm}$  (106 Ga)  $^{143}\text{Nd}$

*Chauvel & Blichert-Toft (2001)*

# Melt compositions at mid ocean ridges

Causes of observed isotope compositions



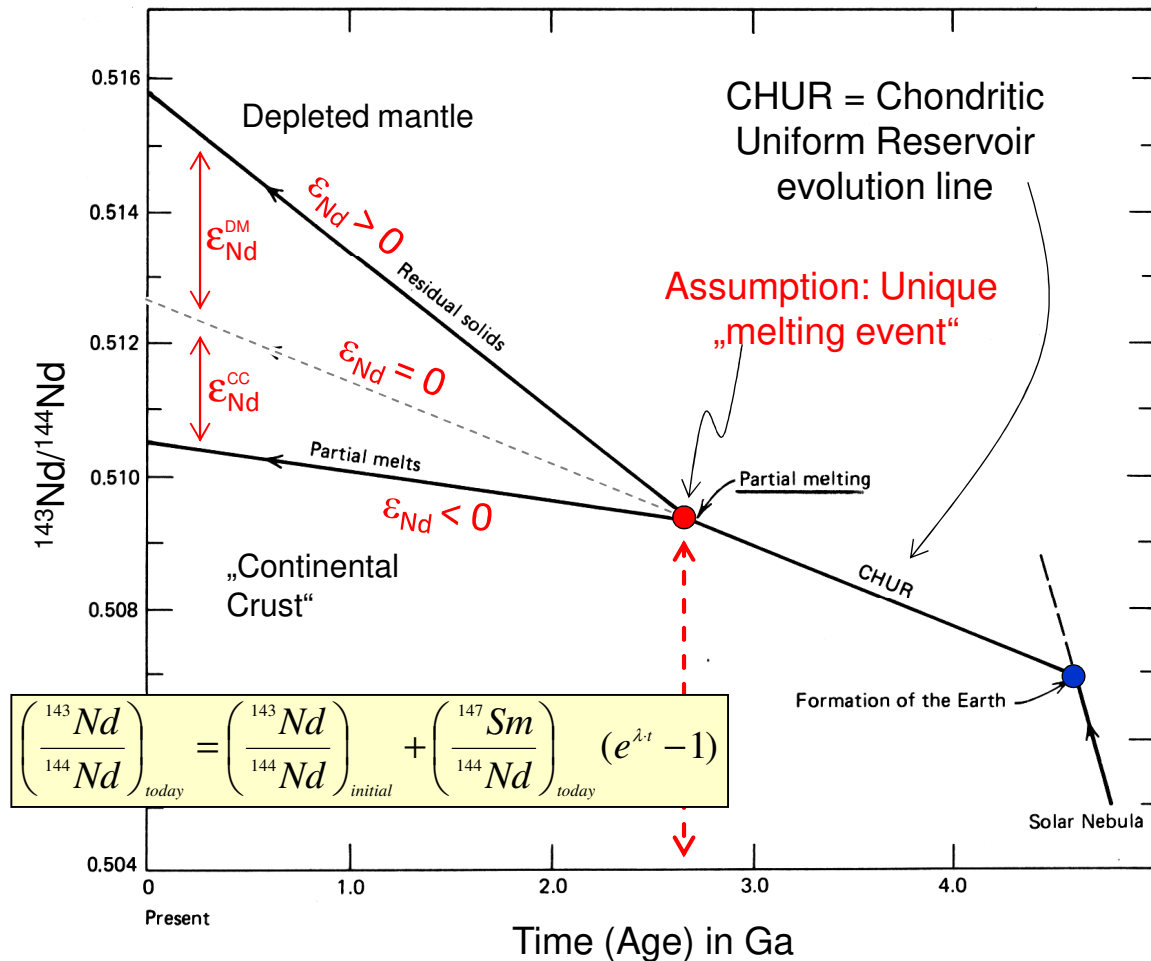
Ratio	DM	CC
Rb/Sr	low	high
Re/Os	low	high
U/Pb	low	high
Th/Pb	low	high
Sm/Nd	high	low
Lu/Hf	high	low

Due to this, and given enough **time**, different geochemical **reservoirs** develop **different radiogenic isotope signatures**

# Melt compositions at mid ocean ridges

How to explain the observed isotope compositions

(here  $^{143}\text{Nd}/^{144}\text{Nd}$ ) – valids basically also for other radiogenic isotope ratios



## Data base:

$$(^{147}\text{Sm}/^{144}\text{Nd})_{\text{DM}} = 0.2395$$

$$(^{143}\text{Nd}/^{144}\text{Nd})_{\text{DM}} = 0.513110$$

$$(^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} = 0.1960$$

$$(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512630$$

DM estimate from Salters & Stracke (2004),  $^{147}\text{Sm}/^{144}\text{Nd}$  calculated from concentrations in DM and using natural isotope abundances. Values for CHUR from Bouvier et al. (2008).

Decay const.  $^{147}\text{Sm} = 6.54 \times 10^{-12} \text{ 1/a}$

## Melt compositions at mid ocean ridges

How to explain the observed isotope compositions

(here  $^{143}\text{Nd}/^{144}\text{Nd}$ ) – valid basically also for other radiogenic isotope ratios

Isotope evolution of CHUR over time:

$$\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_0^{\text{CHUR}} = \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_i^{\text{CHUR}} + \left(\frac{^{147}\text{Sm}}{^{144}\text{Nd}}\right)_0^{\text{CHUR}} (e^{\lambda \cdot t} - 1)$$

Isotope evolution of DM over time:

$$\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_0^{\text{DM}} = \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_i^{\text{DM}} + \left(\frac{^{147}\text{Sm}}{^{144}\text{Nd}}\right)_0^{\text{DM}} (e^{\lambda \cdot t} - 1)$$

At the time of the **single melting event** (red point in the previous diagram), the **initial values** of DM and CHUR are equal! This allows to calculate the „age“ of this event.

$^{143}\text{Nd}/^{144}\text{Nd}$  values can be expressed as **deviation** relative to CHUR for any time **t** in the past:

$$\epsilon_{\text{Nd}}^{\text{sample}}(t) = \left[ \frac{(^{143}\text{Nd}/^{144}\text{Nd})_t^{\text{sample}}}{(^{143}\text{Nd}/^{144}\text{Nd})_t^{\text{CHUR}}} - 1 \right] \times 10^4$$

„sample“ means any reservoir (DM, CC, ...) or any individual sample, t is time

**Data:**

$$(^{147}\text{Sm}/^{144}\text{Nd})_{\text{DM}} = 0.2395$$

$$(^{143}\text{Nd}/^{144}\text{Nd})_{\text{DM}} = 0.513110$$

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*DM estimate from Salters & Stracke (2004),  $^{147}\text{Sm}/^{144}\text{Nd}$  calculated from concentrations in DM and using natural isotope abundances. Values for CHUR from Bouvier et al. (2008).*

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## Melt compositions at mid ocean ridges

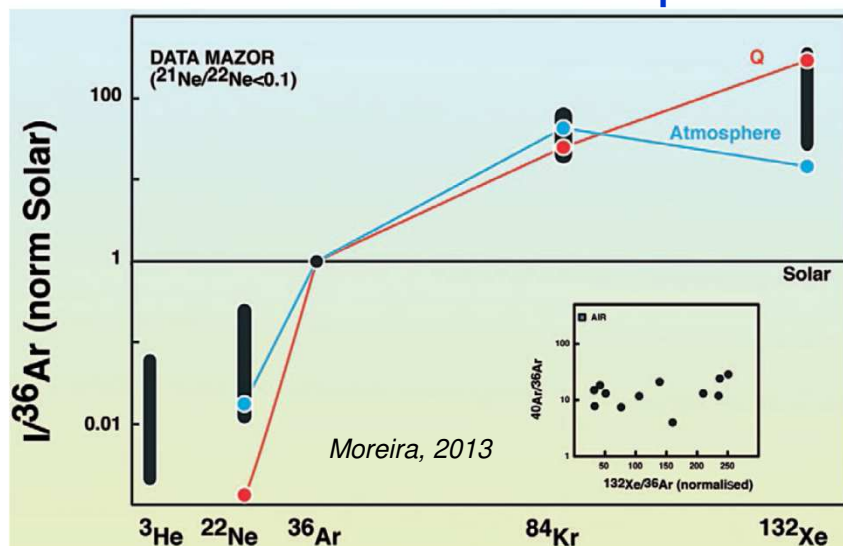
Noble gas systematics – mantle degassing

### Sources of terrestrial noble gases

- **Inherited** (from the solar nebula)
- **Nucleogenic** (products of nuclear reactions)
- **Cosmogenic**



### Noble gas elemental ratios in chondrites and the terrestrial atmosphere



**Table 3** Some nuclear reactions producing noble gases.

Reaction	Upper crust production ratios <sup>a</sup>
${}^6\text{Li}(n,\alpha){}^3\text{H}(\beta-){}^3\text{He}$	${}^3\text{He}/{}^4\text{He} = 1 \times 10^{-8}$
${}^{17}\text{O}(\alpha,n){}^{20}\text{Ne}$	${}^{20}\text{Ne}/{}^4\text{He} = 4.4 \times 10^{-9}$
${}^{18}\text{O}(\alpha,n){}^{21}\text{Ne}$	${}^{21}\text{Ne}/{}^4\text{He} = 4.5 \times 10^{-8}$
${}^{24}\text{Mg}(n,\alpha){}^{21}\text{Ne}$	${}^{21}\text{Ne}/{}^4\text{He} = 1 \times 10^{-10}$
${}^{25}\text{Mg}(n,\alpha){}^{22}\text{Ne}$	Combined crustal production:
${}^{19}\text{F}(\alpha,n){}^{22}\text{Na}(\beta+){}^{22}\text{Ne}$	${}^{22}\text{Ne}/{}^4\text{He} = 9.1 \times 10^{-8}$
${}^{35}\text{Cl}(n,\gamma){}^{36}\text{Cl}(\beta-){}^{36}\text{Ar}$	${}^{40}\text{Ar}/{}^{36}\text{Ar} = 1.5 \times 10^7$

Source: Porcelli *et al.* (2002).

<sup>a</sup> See Ballentine and Burnard (2002) for details and production rates in different compositions.

## Melt compositions at mid ocean ridges

### Noble gas systematics – mantle degassing

**Table 1** Solar system isotopic compositions of He, Ne, and Ar.

<i>Reservoir</i>	$^3\text{He}/^4\text{He}$ ( $\times 10^{-6}$ )	$R/R_A^a$	$^{20}\text{Ne}/^{22}\text{Ne}$	$^{21}\text{Ne}/^{22}\text{Ne}$	$^{38}\text{Ar}/^{36}\text{Ar}$	$^{40}\text{Ar}/^{36}\text{Ar}$
Solar	457	326	13.8	0.0328	0.1825	$\sim 3 \times 10^{-4}$
Planetary	143	102	8.2	0.024	0.188	$\sim 3 \times 10^{-4}$
Earth atmosphere	1.4	1	9.8	0.0290	0.1880	295.5

After [McDougall and Honda \(1998\)](#).

<sup>a</sup>  $R/R_A = (^3\text{He}/^4\text{He})_{\text{observed}} / (^3\text{He}/^4\text{He})_{\text{air}}$ , where air  $^3\text{He}/^4\text{He} = 1.4 \times 10^{-6}$ .

**Table 2** Half-lives of parent nuclides for noble gases.

<i>Nuclide</i>	<i>Half-life</i>	<i>Daughter</i>	<i>Yield</i> (atoms/decay)	<i>Comments</i>
$^3\text{H}$	12.26 yr	$^3\text{He}$	1	Continuously produced in atm
$^{238}\text{U}$	4.468 Gyr	$^4\text{He}$	8	
$^{235}\text{U}$	0.7038 Gyr	$^{136}\text{Xe}$	$3.6 \times 10^{-8}$ ( $4.4 \pm 0.1$ ) $\times 10^{-8}$	Spontaneous fission $^{238}\text{U}/^{235}\text{U} = 137.88$
$^{232}\text{Th}$	14.01 Gyr	$^4\text{He}$	7	
$^{40}\text{K}$	1.251 Gyr	$^4\text{He}$	6	Th/U = 3.8 in bulk Earth
$^{244}\text{Pu}$	80.0 Myr	$^{136}\text{Xe}$	$< 4.2 \times 10^{-11}$	No significant production in Earth
$^{129}\text{I}$	15.7 Myr	$^{129}\text{Xe}$	1	$^{40}\text{K} = 0.01167\%$ total K $^{244}\text{Pu}/^{238}\text{U} = 6.8 \times 10^{-3}$ at 4.56 Ga $^{129}\text{I}/^{127}\text{I} = 1.1 \times 10^{-4}$ at 4.56 Ga

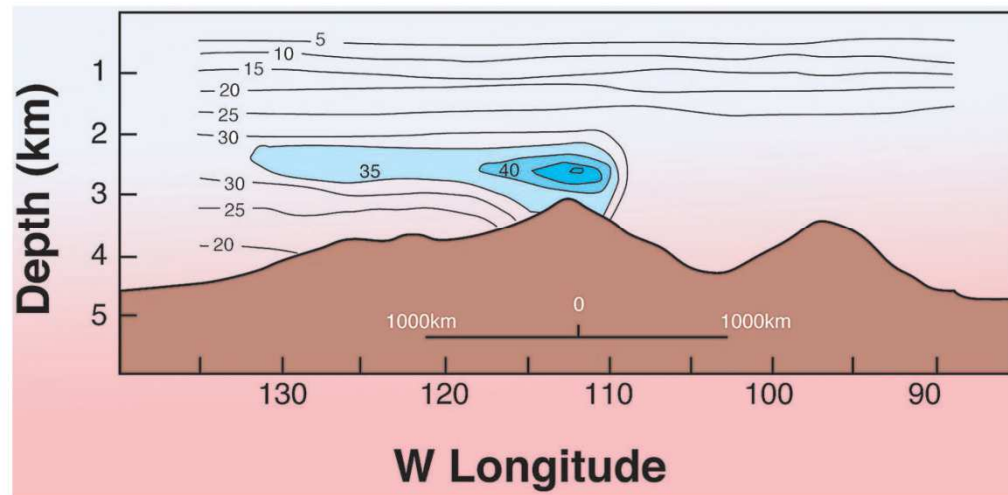
Source: [Porcelli et al. \(2002\)](#).

From [Hilton & Porcelli \(2003\)](#), *Treatise on Geochemistry*

# Melt compositions at mid ocean ridges

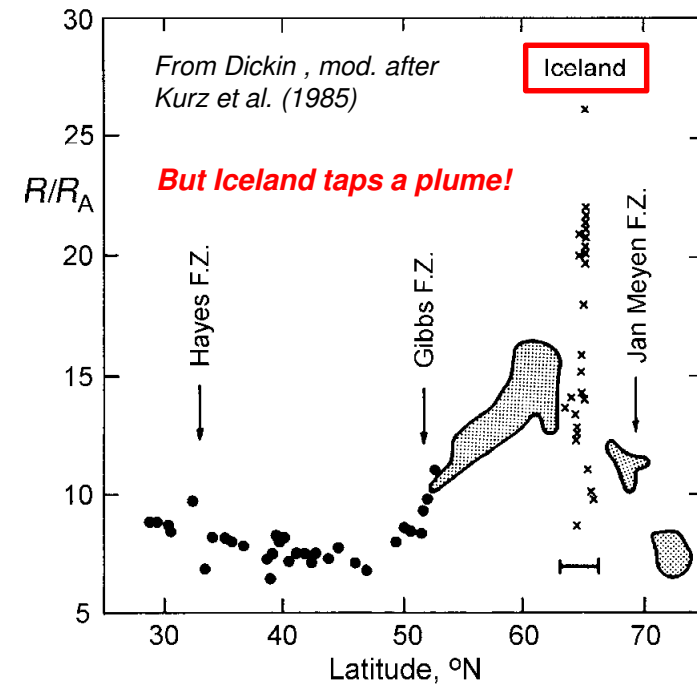
Noble gas systematics – is there primordial He in the mantle?

$^3\text{He}$  excess in seawater above the East Pacific Rise



Excluding nucleogenic sources for  $^3\text{He}$ , the **elevated  $^3\text{He}/^4\text{He}$  ratios** relative to air (i.e.,  $R/R_A > 1$ ) in **MORB** and **submarine hydrothermal fluids** suggest that the Earth's mantle is **not** completely degassed (i.e., comprises primordial Helium)

$^3\text{He}/^4\text{He}$  relative to air along the Mid Atlantic Ridge

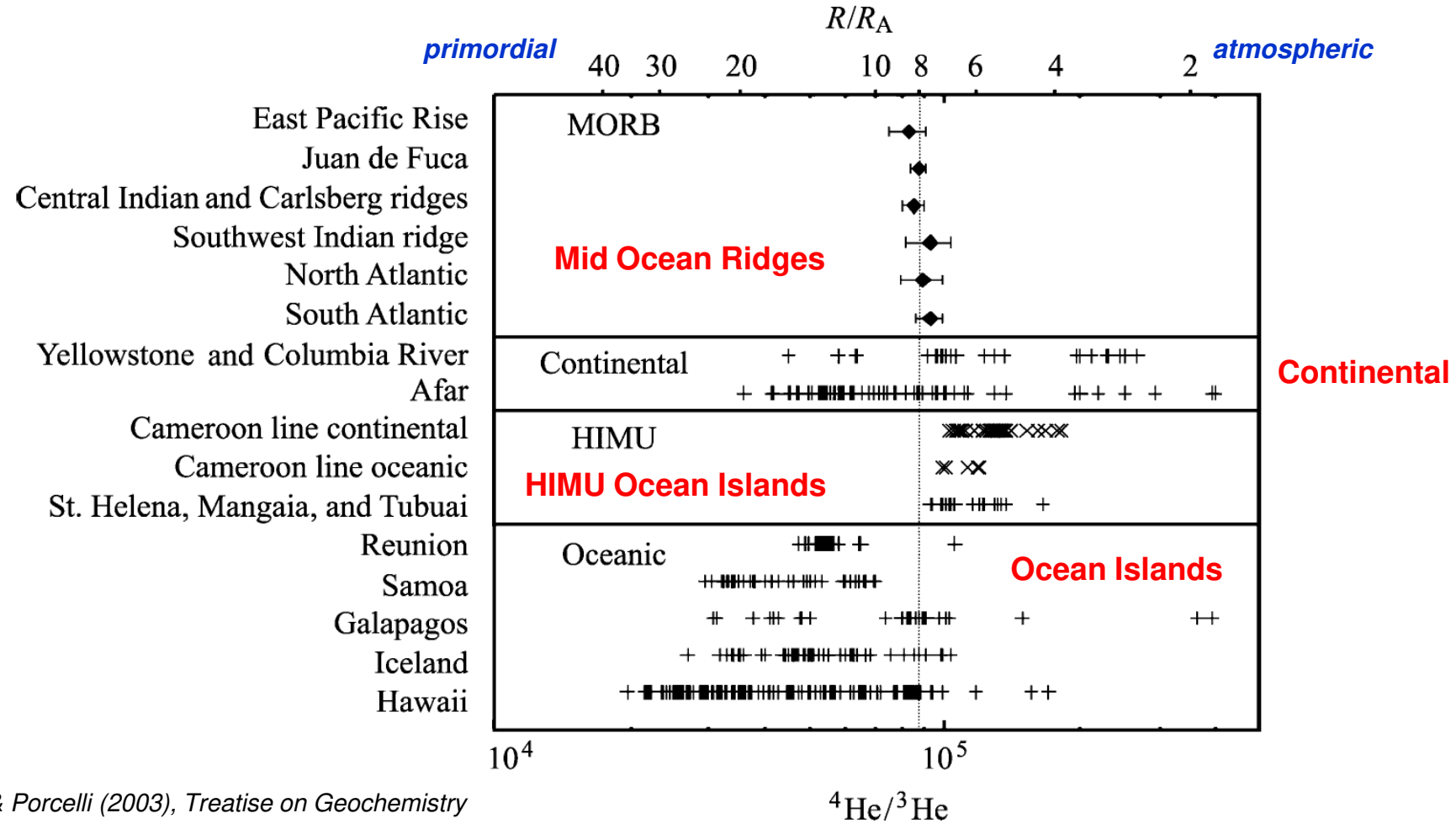


# Mid ocean ridges – MELT COMPOSITIONS

## Melt compositions at mid ocean ridges ... and elsewhere ;)

Noble gas systematics – is there primordial He in the mantle?

He isotope composition of basalts from different settings



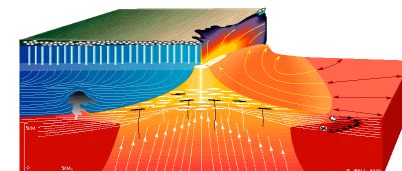
Hilton & Porcelli (2003), Treatise on Geochemistry

## Melt compositions at mid ocean ridges ... and elsewhere ;)

Partial melting in the mantle is a fundamental process that determines the **primary** composition of basalts, but the final composition of MORB is controlled by **fractional crystallisation!**

During this process, *primary basalt* differentiates to form **cumulate rocks** (e.g. wehrlites, gabbros) on the one hand, and **differentiated basalts** on the other.

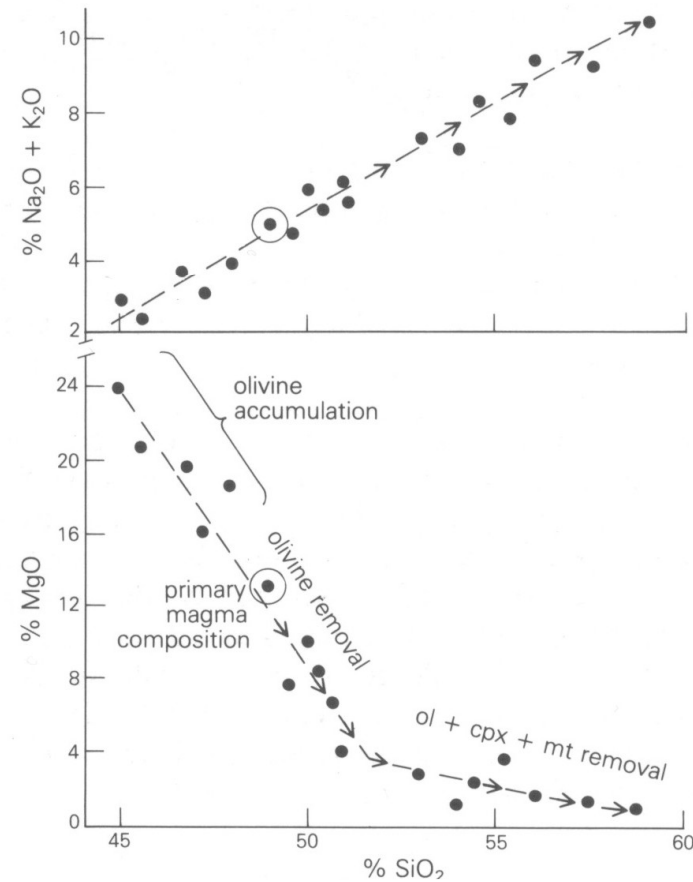
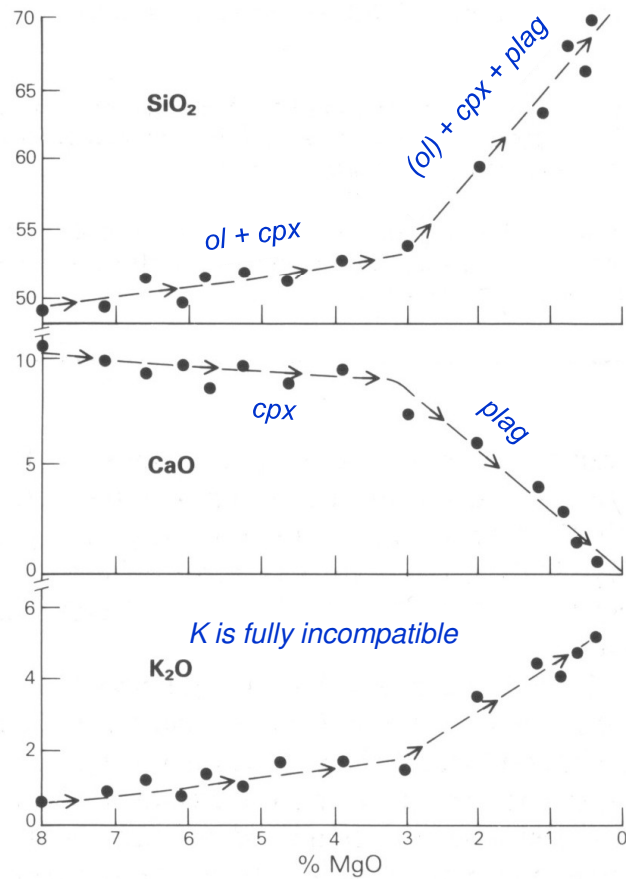
In the oceanic crust, mostly differentiated basalt forms the upper gabbros along with the extrusives (sheeted dikes, pillow lavas, lava flows & sills)





# Melt compositions at mid ocean ridges

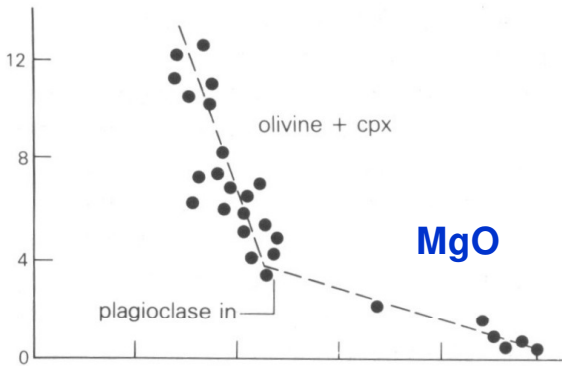
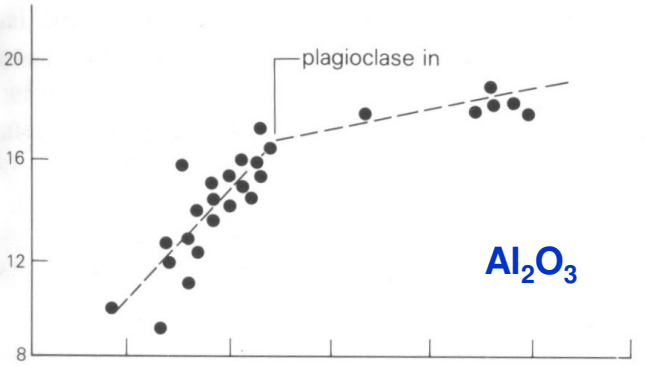
Low-pressure fractional crystallisation of MORB (very briefly!)



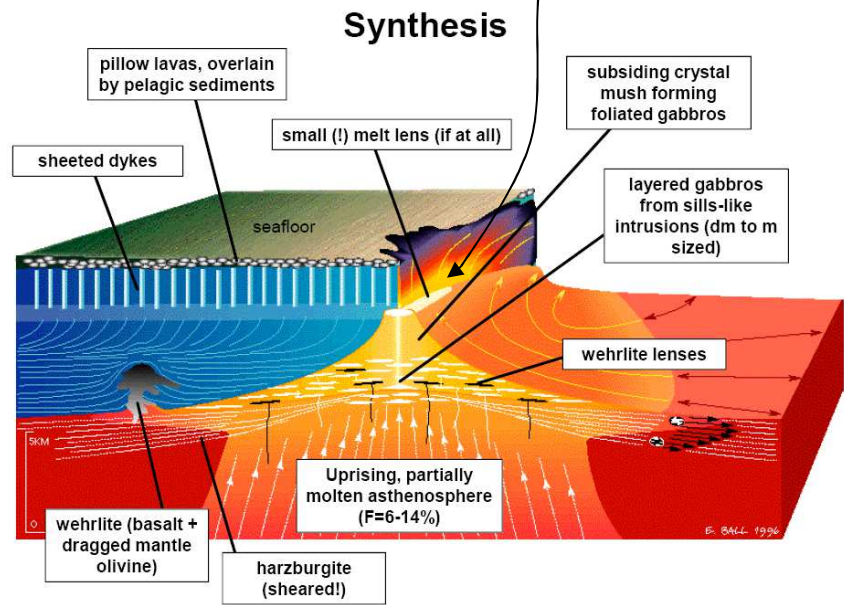
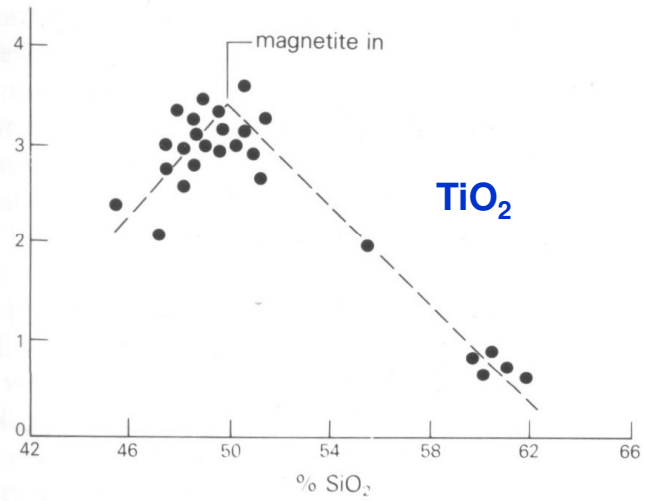
# Mid ocean ridges – MELT COMPOSITIONS

## Melt compositions at mid ocean ridges

Low-pressure fractional crystallisation of MORB (very briefly!)

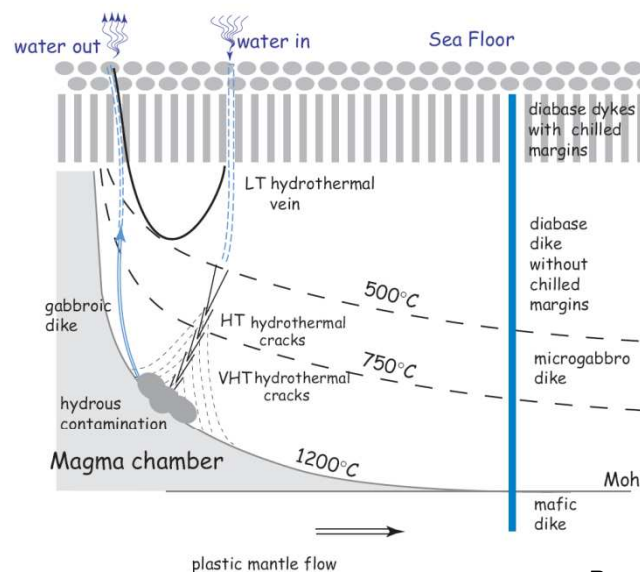
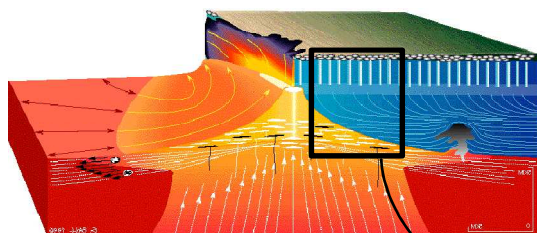


These fractional crystallisation processes predominantly occur during magma ascend and in the melt lenses located at the base of the sheeted dike complex



## Hydrothermal processes at mid ocean ridges

Seawater infiltrates the oceanic crust along fractures



Bosch et al. (2004)

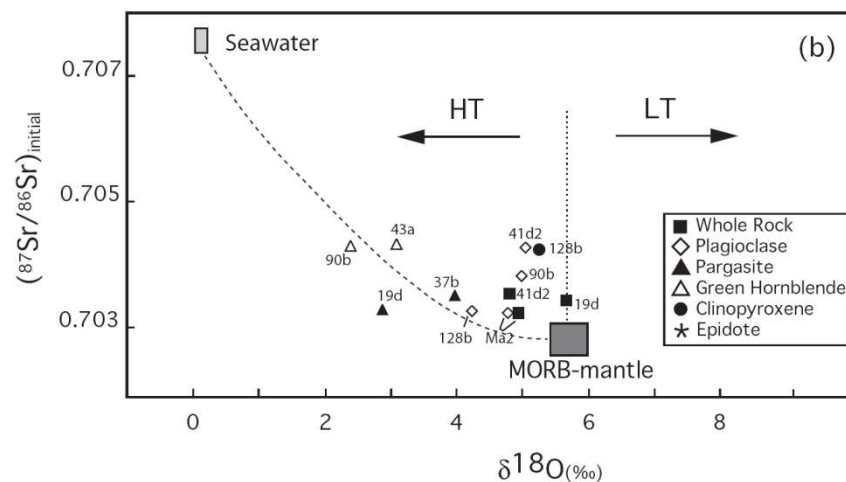
Seawater infiltration causes **hydratisation** reactions, leading to the formation of **water bearing mineral phases**. These are in **basalt/gabbro** (*oceanic crust*) for example:

**lawsonite, amphibole, chlorite, epidote**

and in **harzburgite/peridotite** (*oceanic lithospheric mantle*):

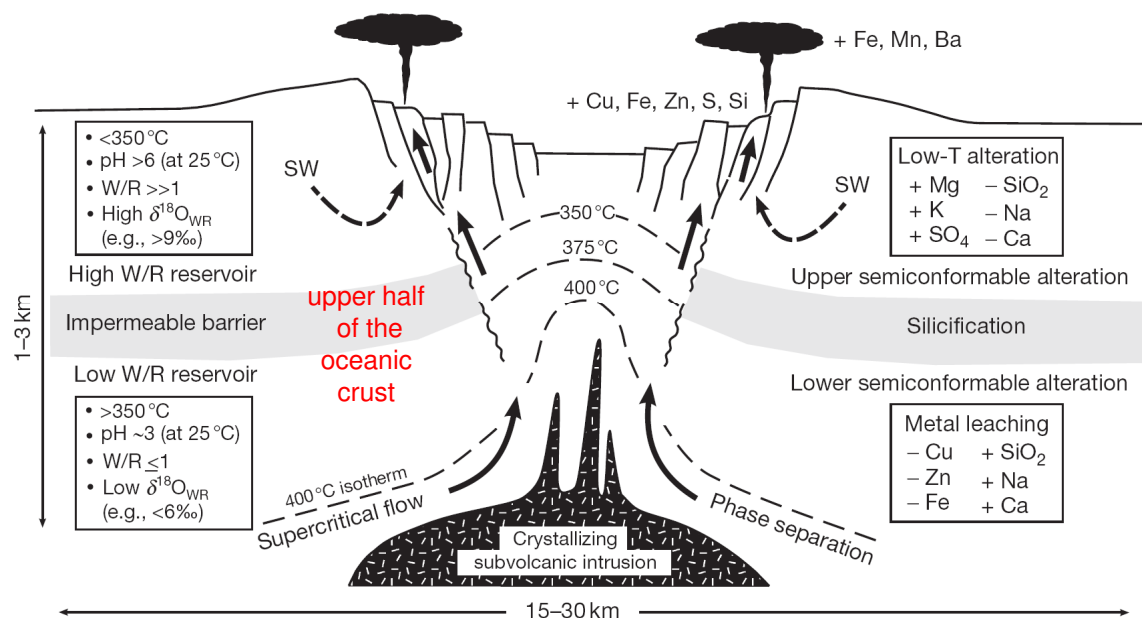
**serpentine, talc, chlorite, amphibole**

→ oxygen isotope composition of **hydrous phases** in gabbros from the Oman ophiolite reveal **exchange with seawater** at temperatures up to **~975°C (!)**



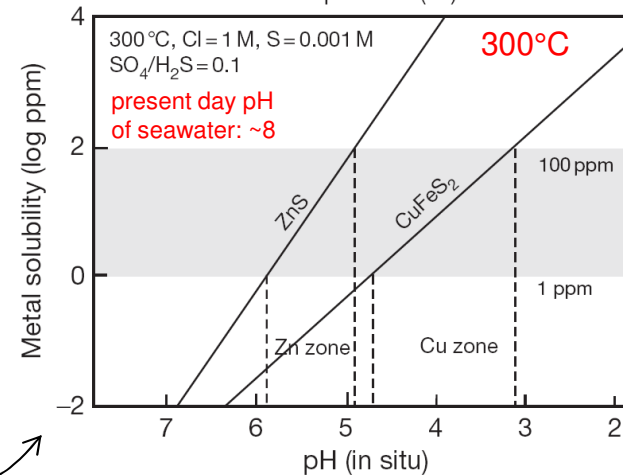
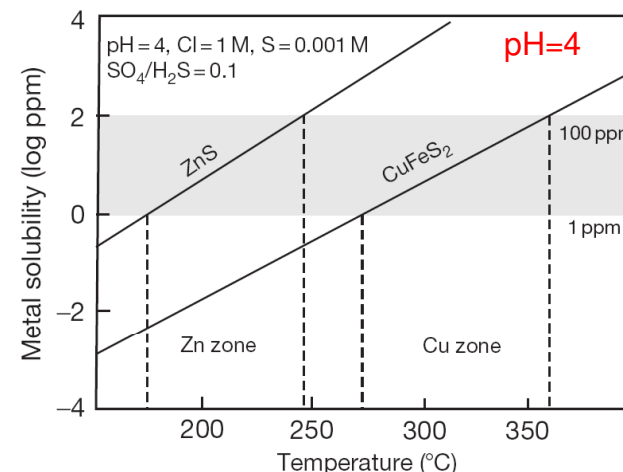
## Hydrothermal processes at mid ocean ridges

Convective hydrothermal systems and VMS deposits



LT and HT **hydrothermal circulation** systems and **mass fluxes** to and out of the **oceanic crust (lithosphere)**. Metals such as **Cu, Zn, and Fe**, but also Ca, Si and S are **leached out**, whereas Mg, K and SO<sub>4</sub><sup>2-</sup> were **added** to the oceanic (crust) lithosphere.

Solubility of **chalcopyrite** and **sphalerite** (in ppm) in **hydrothermal fluids** dependent on **temperature** and **pH value**.



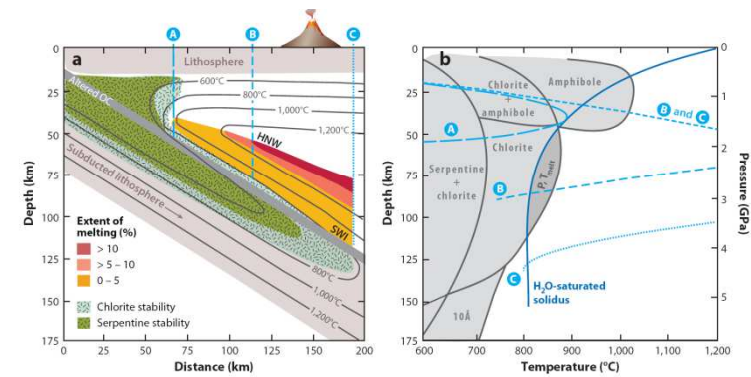
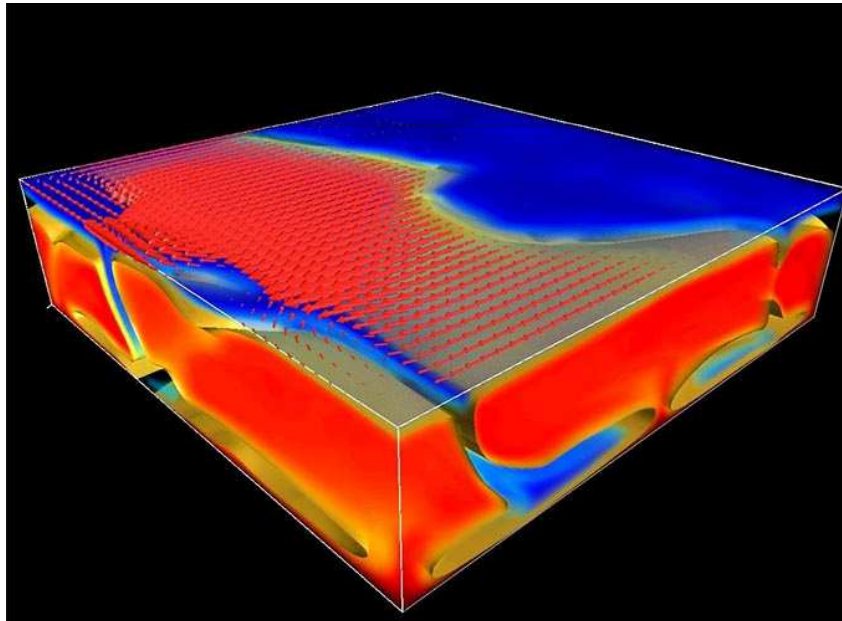
## Mid ocean ridges – magmatic processes

### SYNTHESIS - ABSTRACT

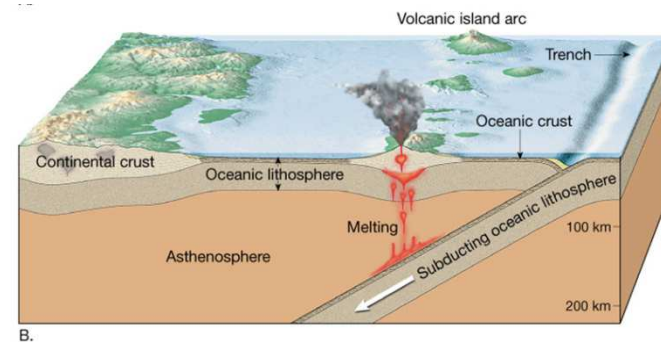
- MORB's provide information about the **(upper) mantle** of the Earth
- They form by **decompression partial melting** of predominantly **spinel-peridotite (lherzolite)**, melting degrees (F) are in the order of **8-10%**
- MORB's have overall **depleted trace-element patterns**, and SiO<sub>2</sub> and MgO contents of undifferentiated magmas are **comparatively uniform** (~50 wt% and ~8-10 wt%, respectively)
- **Ni** and **Cr** contents of **undifferentiated (primary) magmas** in equilibrium with mantle peridotite are **~400 ppm Ni** and **~500-600 ppm Cr**
- Low-pressure **fractional crystallisation** of **Ol + Plag + Cpx** lowers MgO, CaO, Al<sub>2</sub>O<sub>3</sub>, Ni and Cr contents, but increases SiO<sub>2</sub>
- **Radiogenic Nd - Hf** and **unradiogenic Sr** isotope compositions of MORB indicate that the **depletion of the upper portion of the mantle** started **early in Earth's history** (evidence for this also comes from short-lived <sup>146</sup>Sm (103 Ma) <sup>142</sup>Nd)

Chapter 3

# (Oceanic) Island Arcs



Grove et al., 2012



B.

## Arc magmatism: Introduction – Terms - Rocks

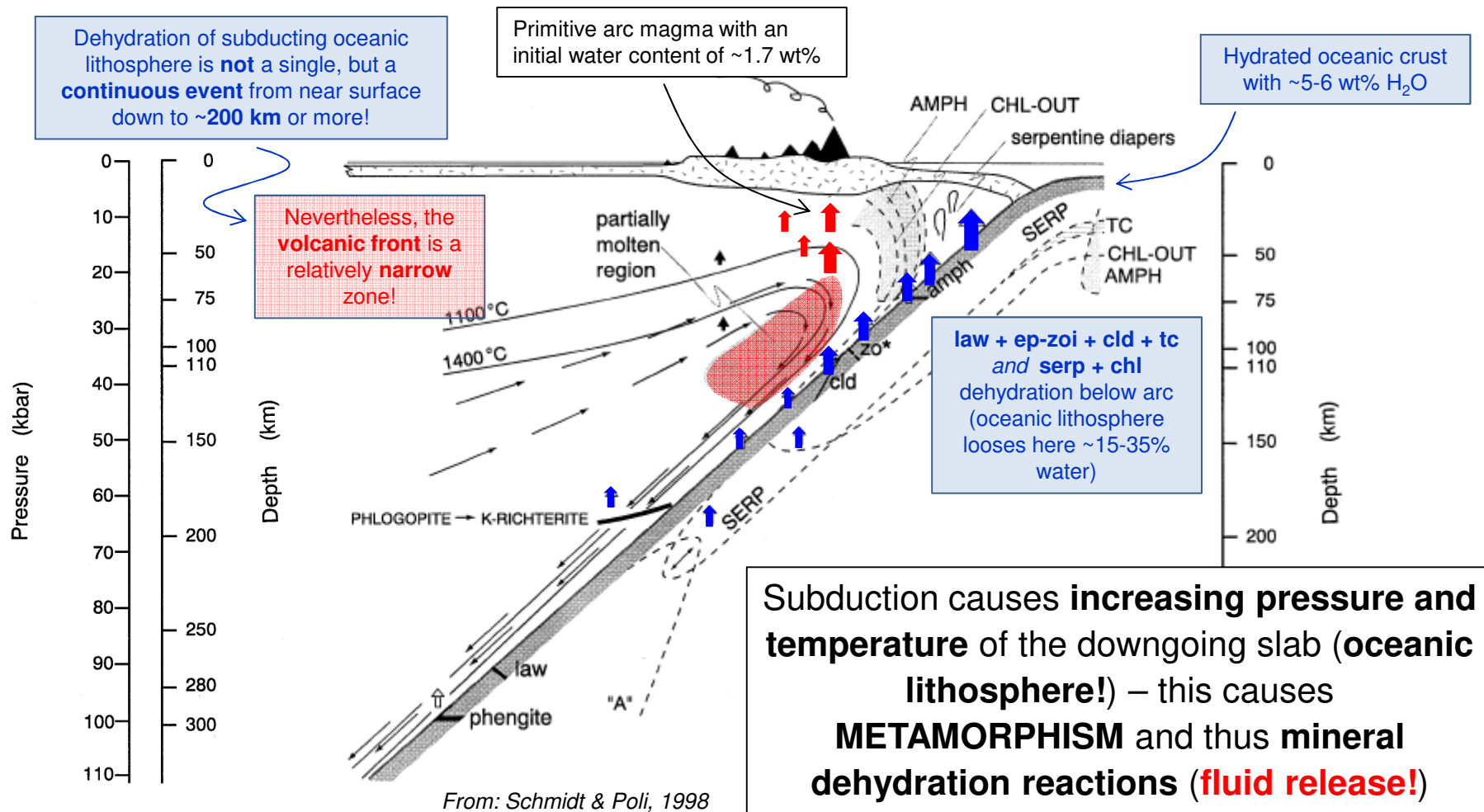
### Three selected terms relevant in the context of arc magmatic rocks

**Adakite:** An andesite or dacite with very high LRRE at low HREE ( $\text{La/Yb} > 9$ , very high Sr/Y, e.g.  $> 50$ ) (from Adak Island)

**Boninite:** Andesites with very low  $\text{TiO}_2$  ( $< 0.5 \text{ wt}\%$ ) (from Bonin Island)

**Calc-alkaline:** Magmas with higher  $\text{Na} + \text{K}$  and  $\text{SiO}_2$  at given Mg# than in tholeiites

# Arc magmatism is caused by subduction zone processes: What happens during subduction?

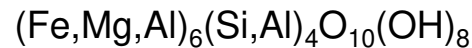


# What happens during subduction?

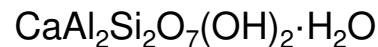
Water bearing phases in hydrous basalt and peridotite

## Phases stable in hydrous basalt (numbers in parentheses give H<sub>2</sub>O content in wt%)

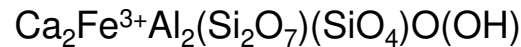
Chlorite (~12%)



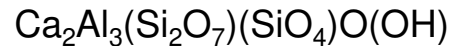
Lawsonite (~11.3%)



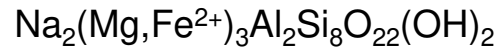
Epidote (~1.9%)



Zoisite (~1.9%)



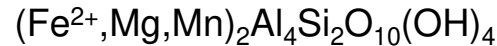
Amph (e.g. Glaukophane, ~2.1%)



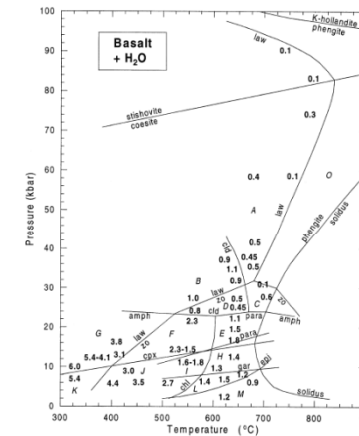
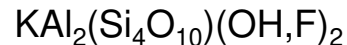
Paragonite (~6.1%)



Chloritoid (~7.4%)

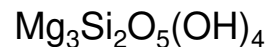


Phengite (~2.5%)

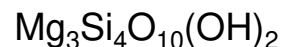


## Phases stable in hydrous peridotite (numbers in parentheses give H<sub>2</sub>O content in wt%)

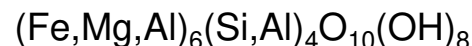
Serpentine (e.g. Chrysotil, ~13%)



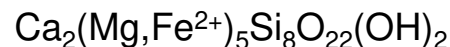
Talc (~4.7%)



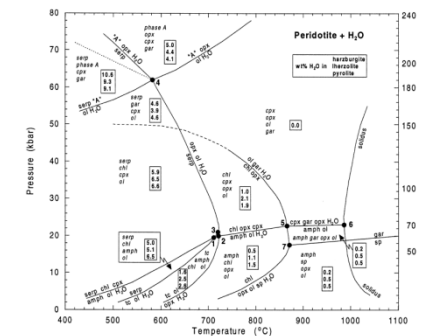
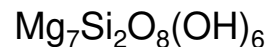
Chlorite (~10.8%)



Amphibole (e.g. Aktinolite, ~2.3%)

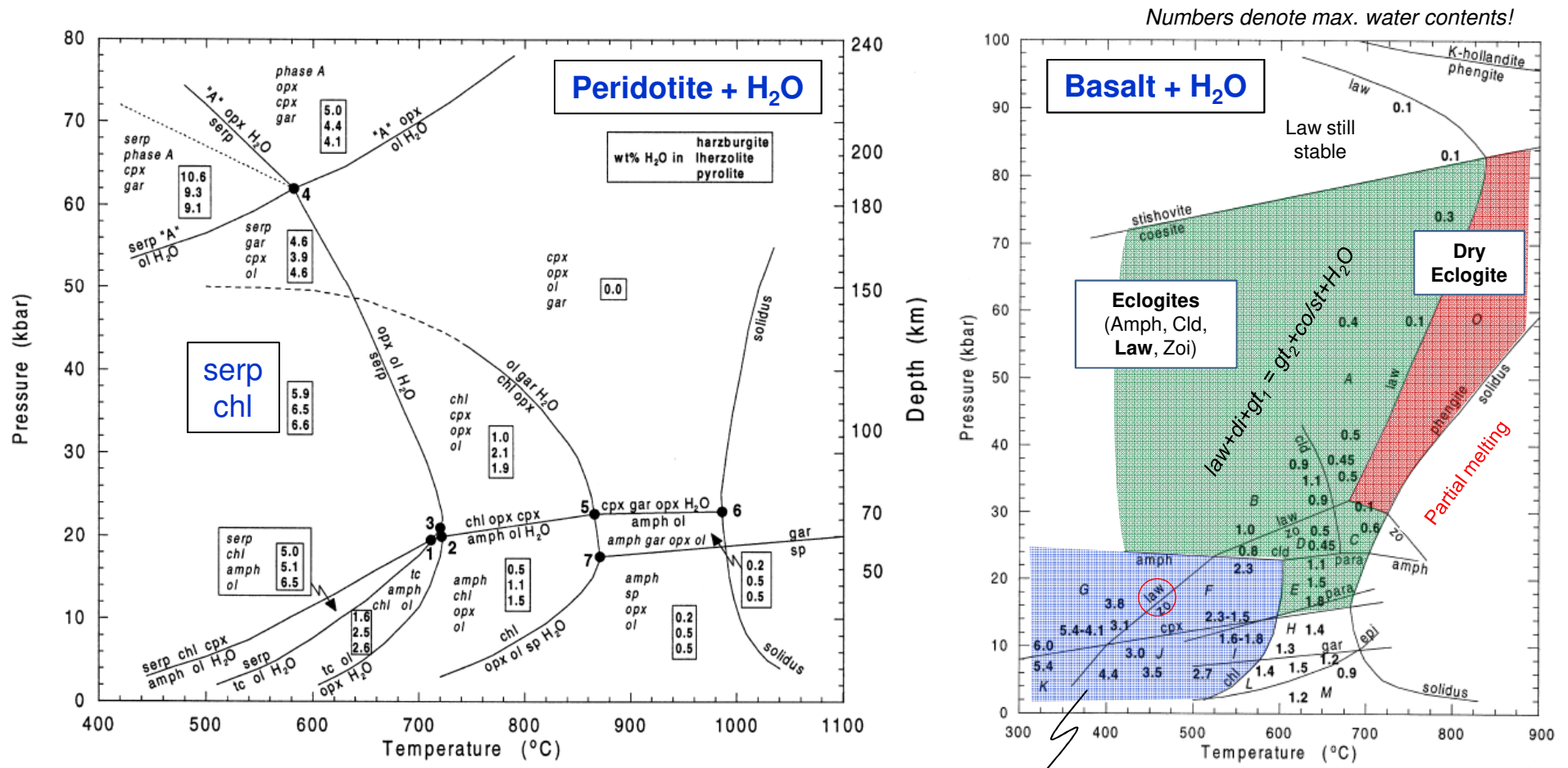


Phase A (~11.8%)



# What happens during subduction?

Dehydration phase relations in hydrous basalt and peridotite



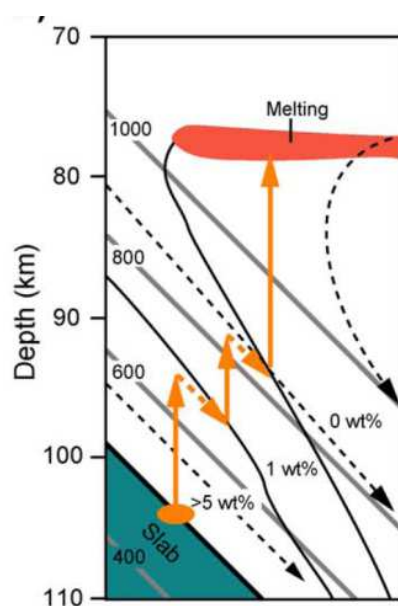
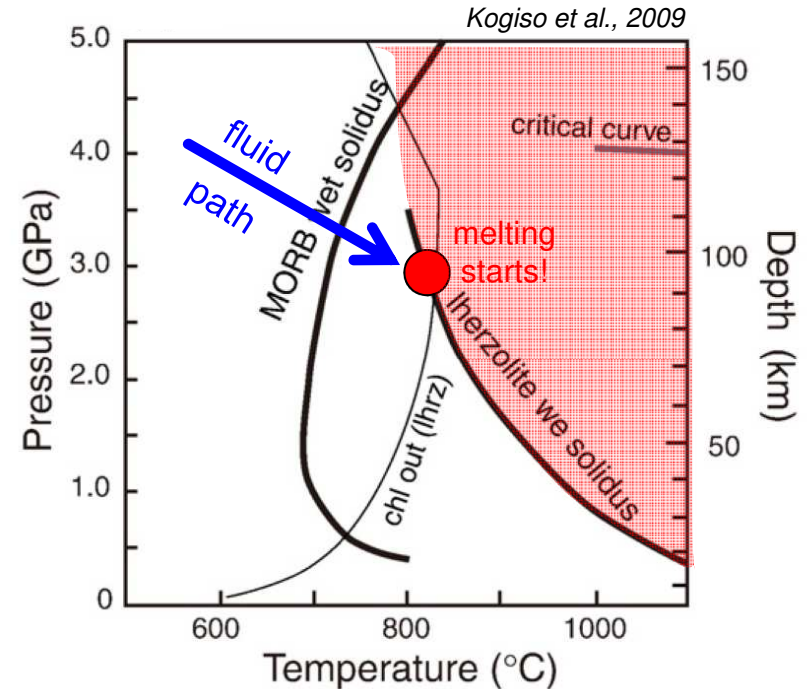
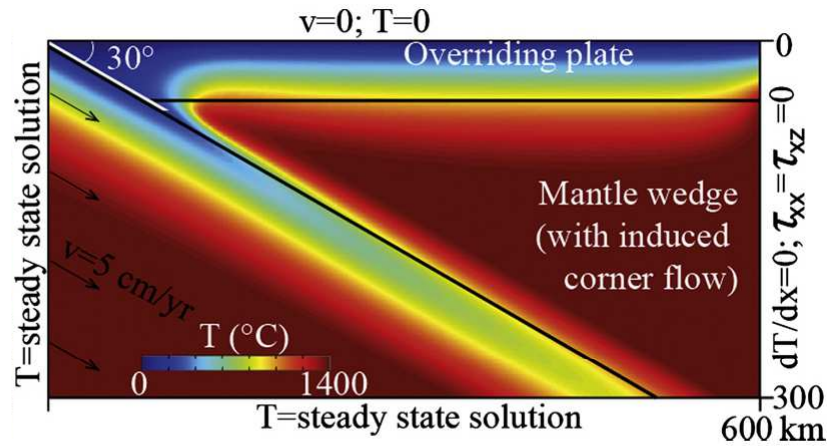
1 GPa = 10 kbar ~ 30 km depth

Blueschist: Law + Gln + Chl + Alb/Jad

Schmidt & Poli, 1998

# What happens during subduction?

Thermal structure within a subduction zone



Builhol et al., 2015

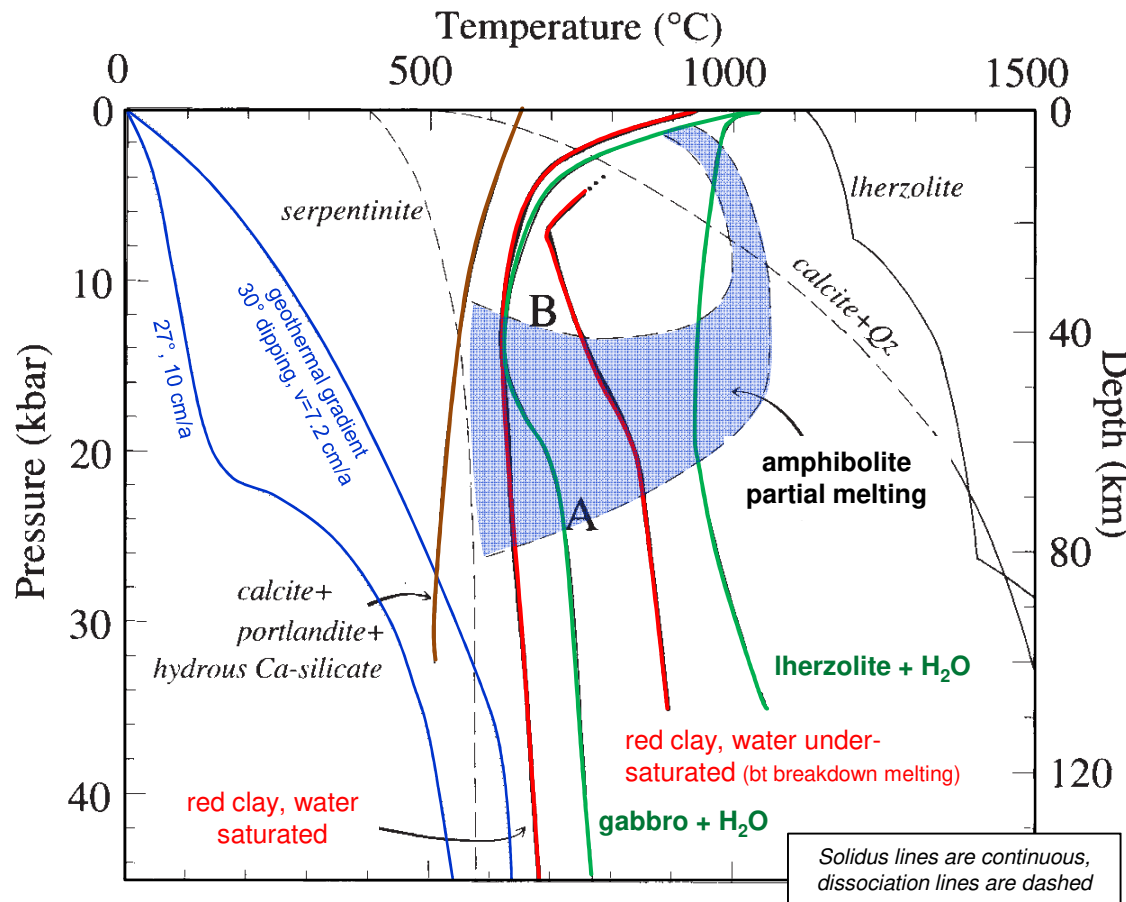
Thermal structure within a subduction zone and flow path of fluids

Manning, 2004

Fluid rises **upwards** towards **higher** temperatures until the solidus of **wet Iherzolite** is reached !!!!

# What happens during subduction?

Solidi and dissociation curves in subduction zone lithologies



Nichols et al., 1994, Nature

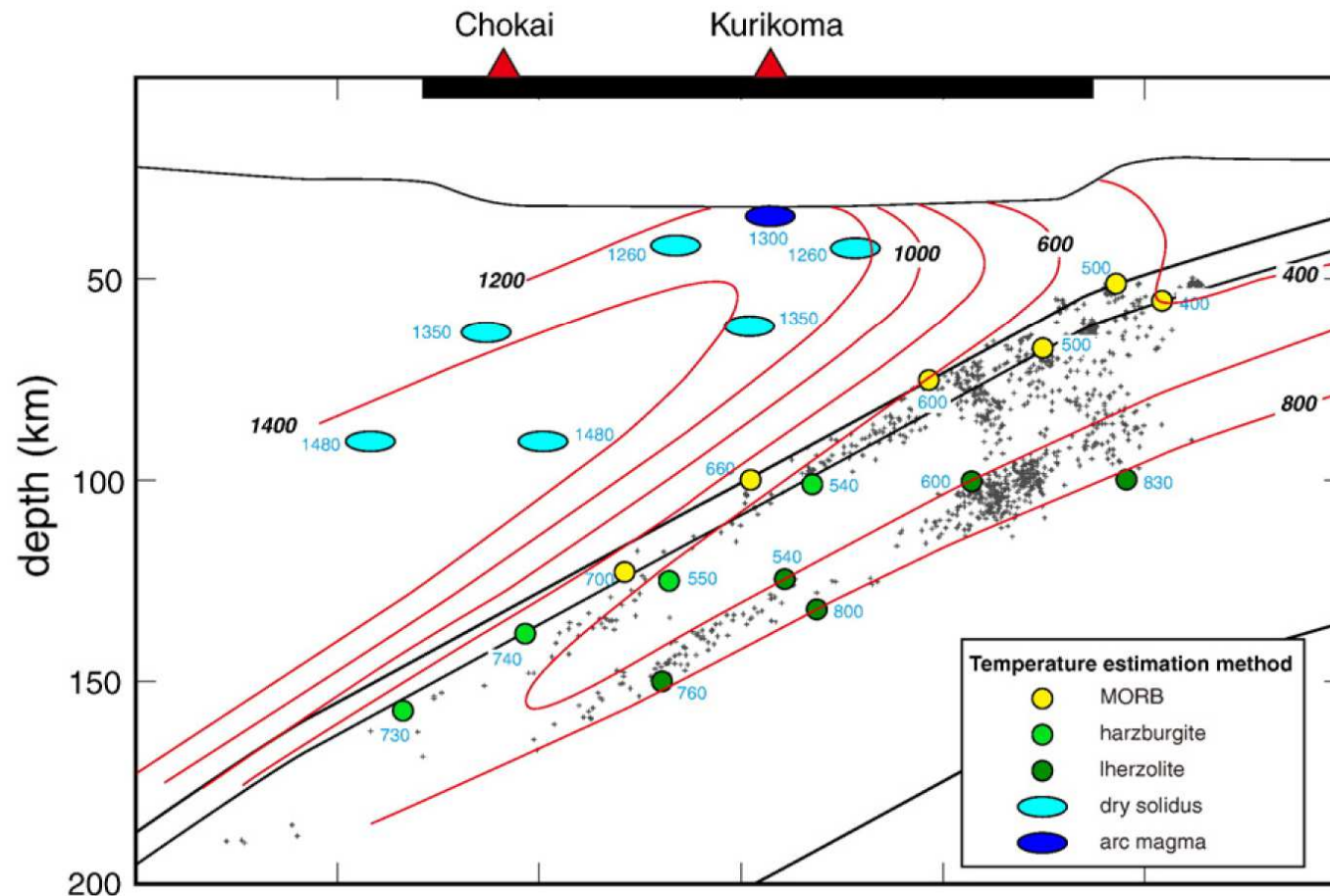
1 GPa = 10 kbar ~ 30 km depth

**Melting** of subduction zone lithologies (oceanic crust, oceanic mantle lithosphere & subducted sediments) requires **slow subduction** (<5-6 cm/a)

But, if melting does occur, the melting order might be as follows:  
 Carbonates > clay + H<sub>2</sub>O > gabbro + H<sub>2</sub>O > clay (us) > lherz. + H<sub>2</sub>O

# What happens during subduction?

## Temperature distribution – real estimates



Estimated **temperature distribution** beneath NE Japan.

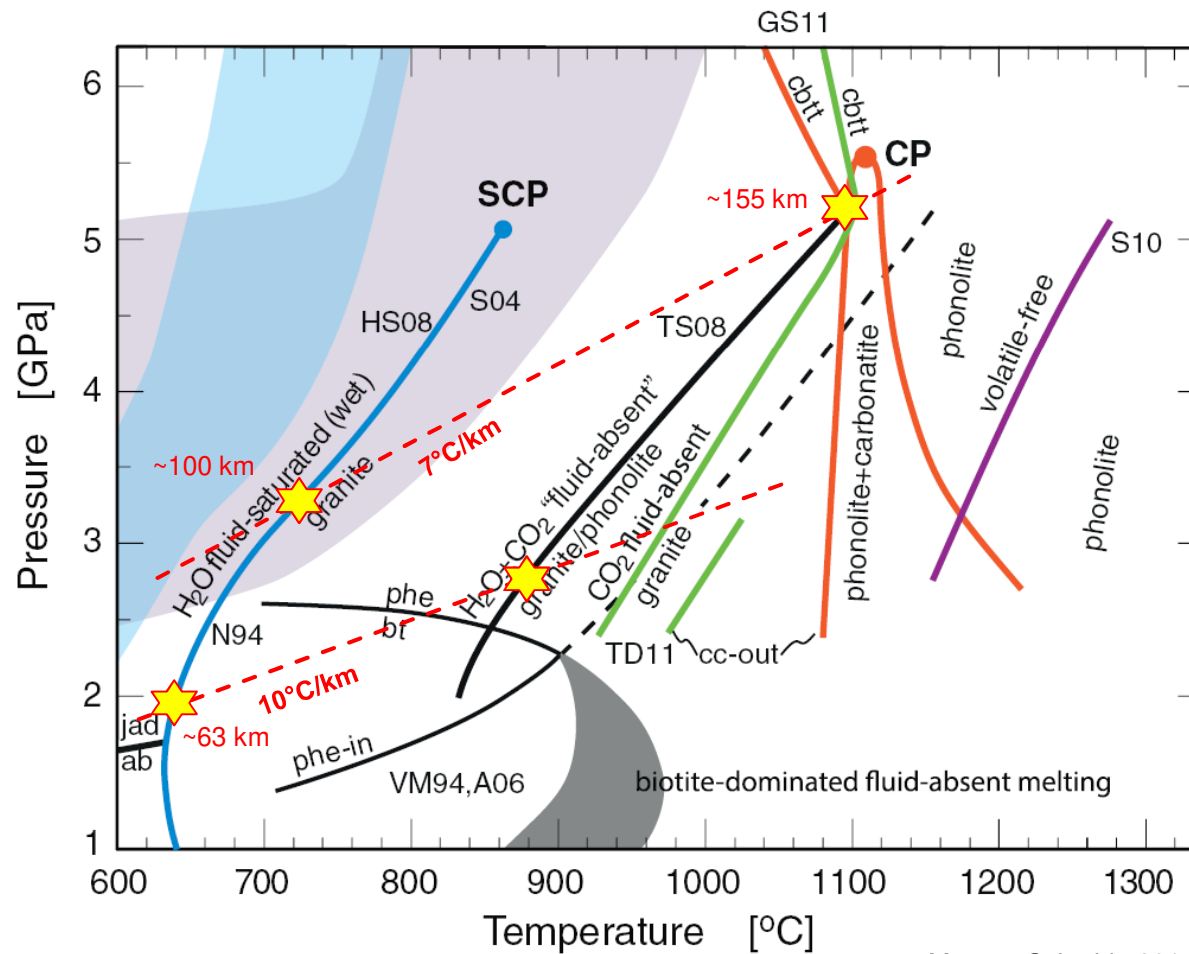
Note the distribution of **intraslab earthquakes**, which are assumed to be caused by **dehydration reactions** (,dehydration embrittlement')

1 GPa = 10 kbar ~ 30 km depth

*Kogiso et al., 2009*

# What happens during subduction?

Phase relations and **solidi** in (hydrated) pelagic sediments



Mann & Schmidt, 2015

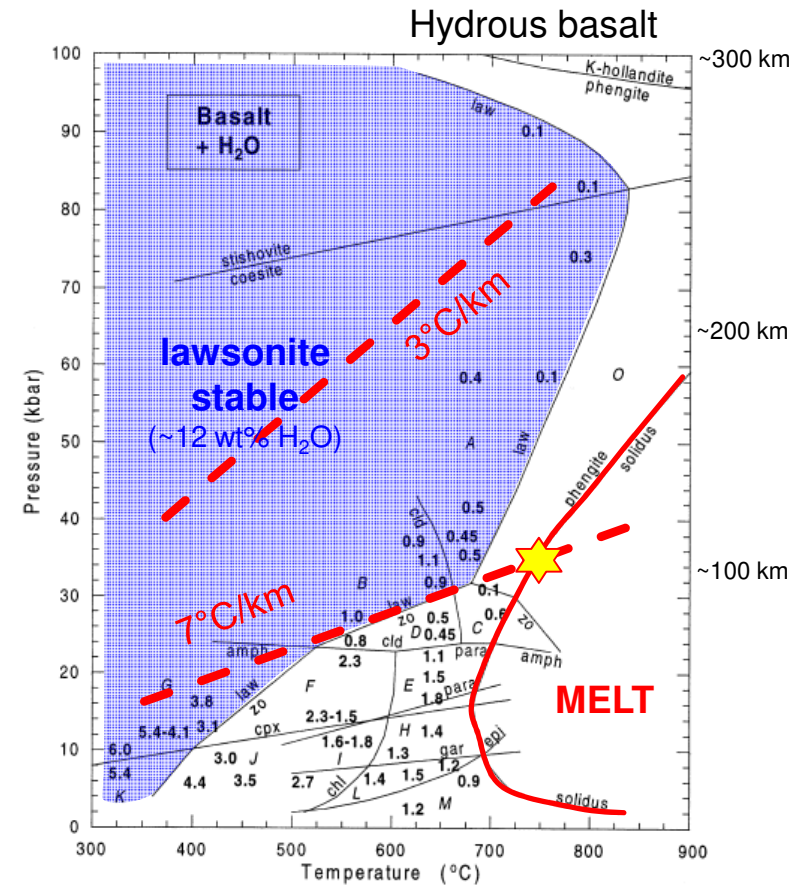
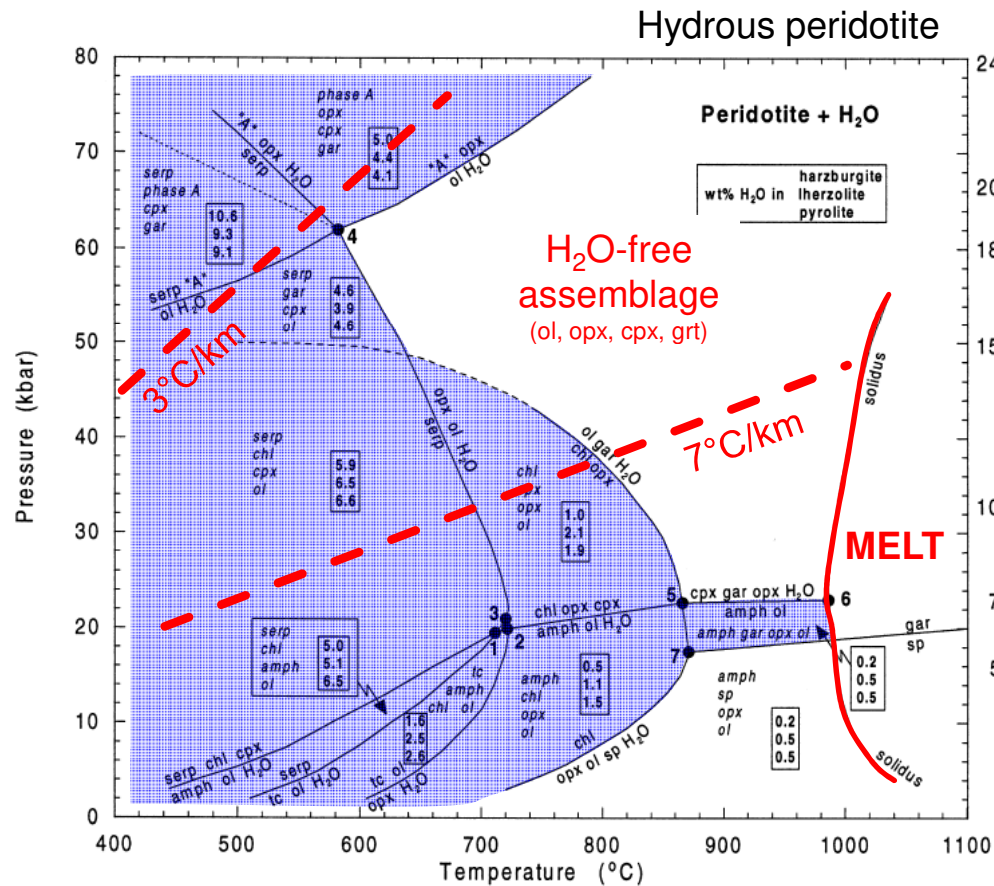
1 GPa = 10 kbar ~ 30 km depth

**Pelites** and **greywackes** make up to ~10% of the uppermost oceanic crust. Their **water contribution** is small, but they host high contents of **LILE** (Cs, Rb, Ba, K, Sr) and <sup>10</sup>Be (and <sup>207</sup>Pb)

At higher **temperature gradients**, dragged **sediments** will start **melting!** Melts will have a hydrous granodioritic to granitic composition.

# What happens during subduction?

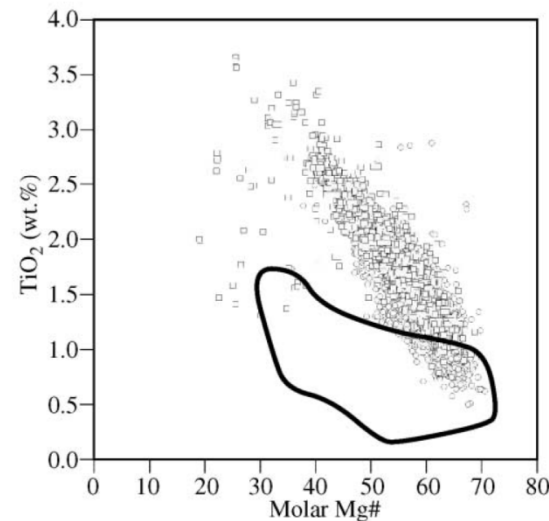
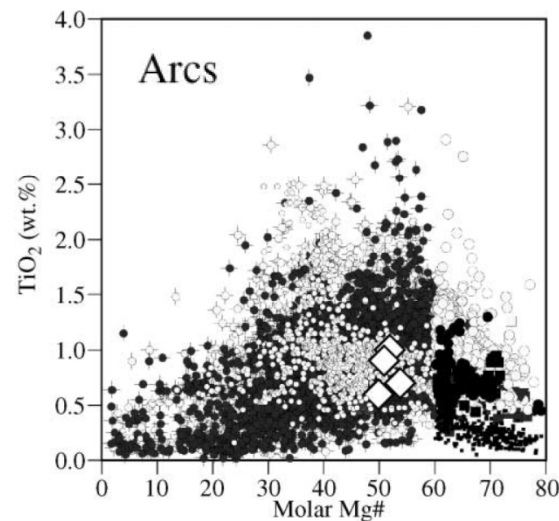
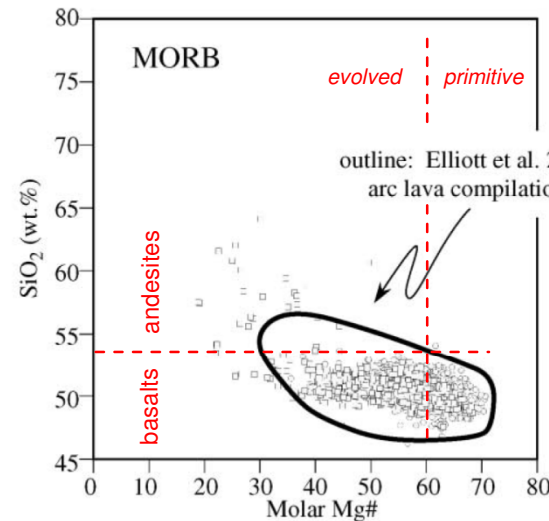
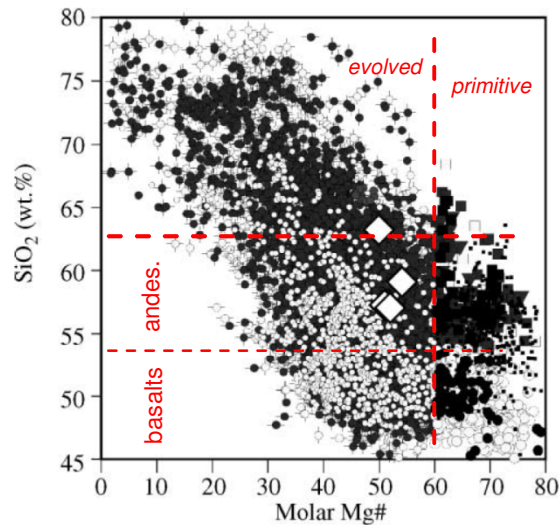
Will the downgoing magmatic lithosphere melt?



1 GPa = 10 kbar ~ 30 km depth

# Chemical signatures in island arc magmas

## Major element characteristics

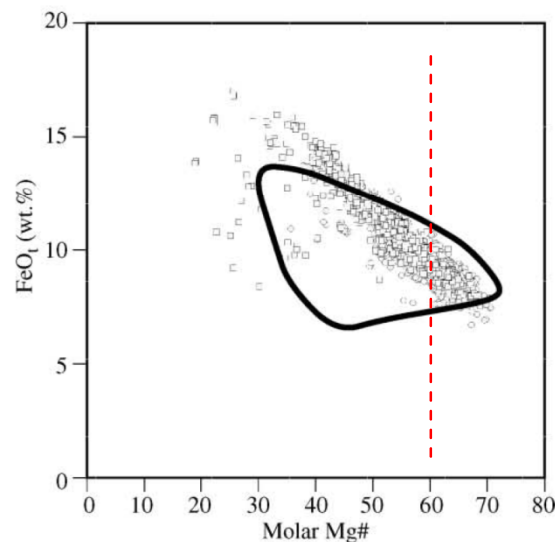
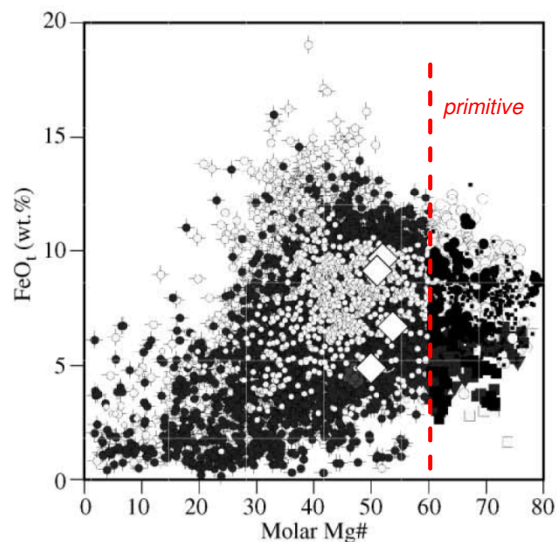
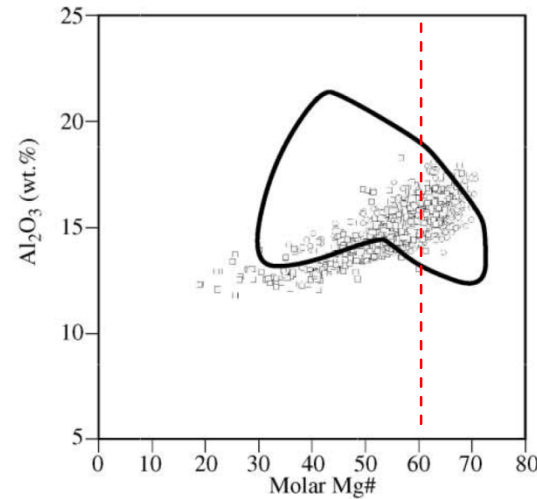
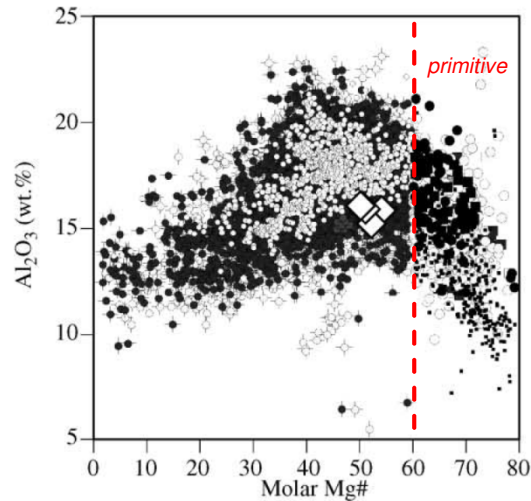


**Primitive** island arc magmas have SiO<sub>2</sub> up to ~**63 wt%**, TiO<sub>2</sub> is **lower** than in MORB. These are **primary features!** Decrease of TiO<sub>2</sub> from Mg#~50 marks onset of Fe-Ti-oxide crystallisation.

$$Mg\# = 100 \times \text{molar} \left[ \frac{MgO}{MgO + FeO} \right]$$

# Chemical signatures in island arc magmas

## Major element characteristics



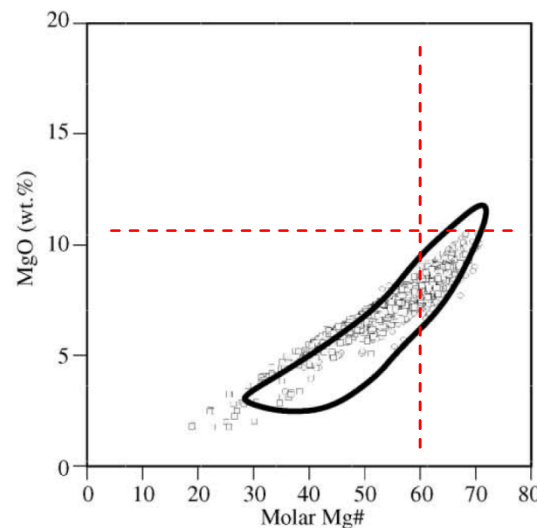
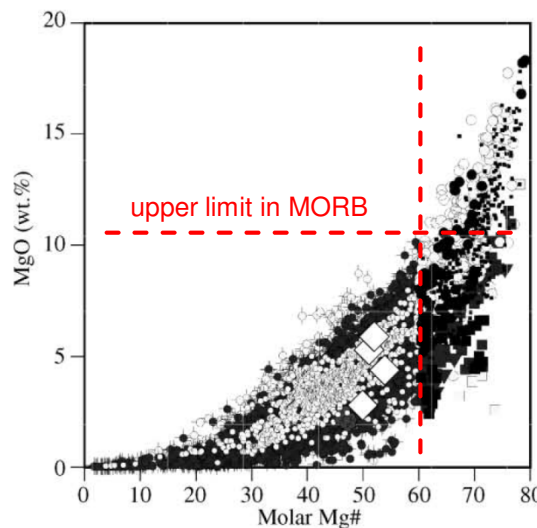
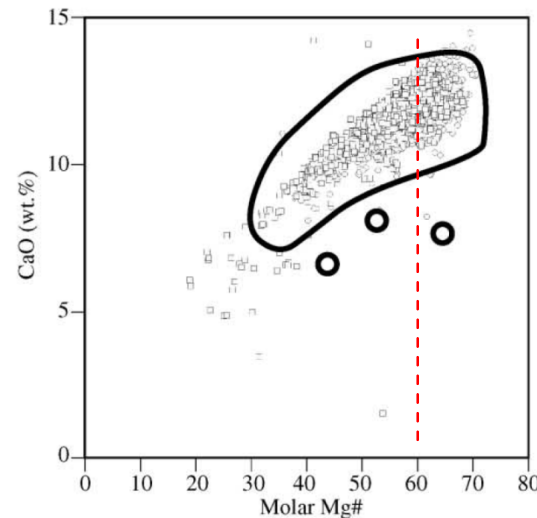
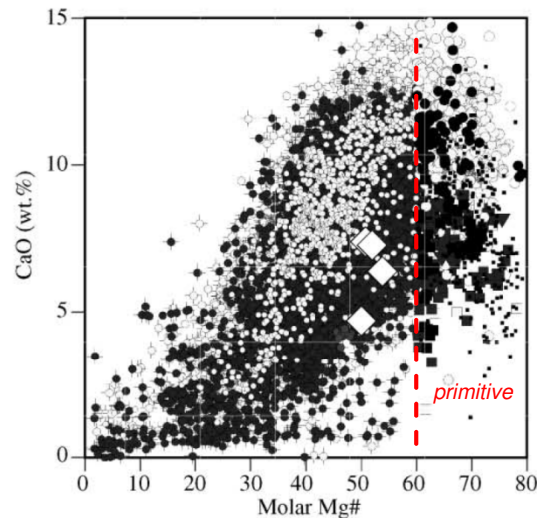
$\text{Al}_2\text{O}_3$  tends to be **similar** or higher as in MORB, with firstly increasing, then decreasing values from **Mg#~45** (onset of **plag** fractionation).

Primary arc magmas predominantly crystallise **ol + cpx** ( $\pm$ opx) to form wehrlite, (ol)-clinopyroxenite, websterite.

**Hornblende** is unstable if liquidus temperatures are  $>1100^\circ\text{C}$

# Chemical signatures in island arc magmas

## Major element characteristics

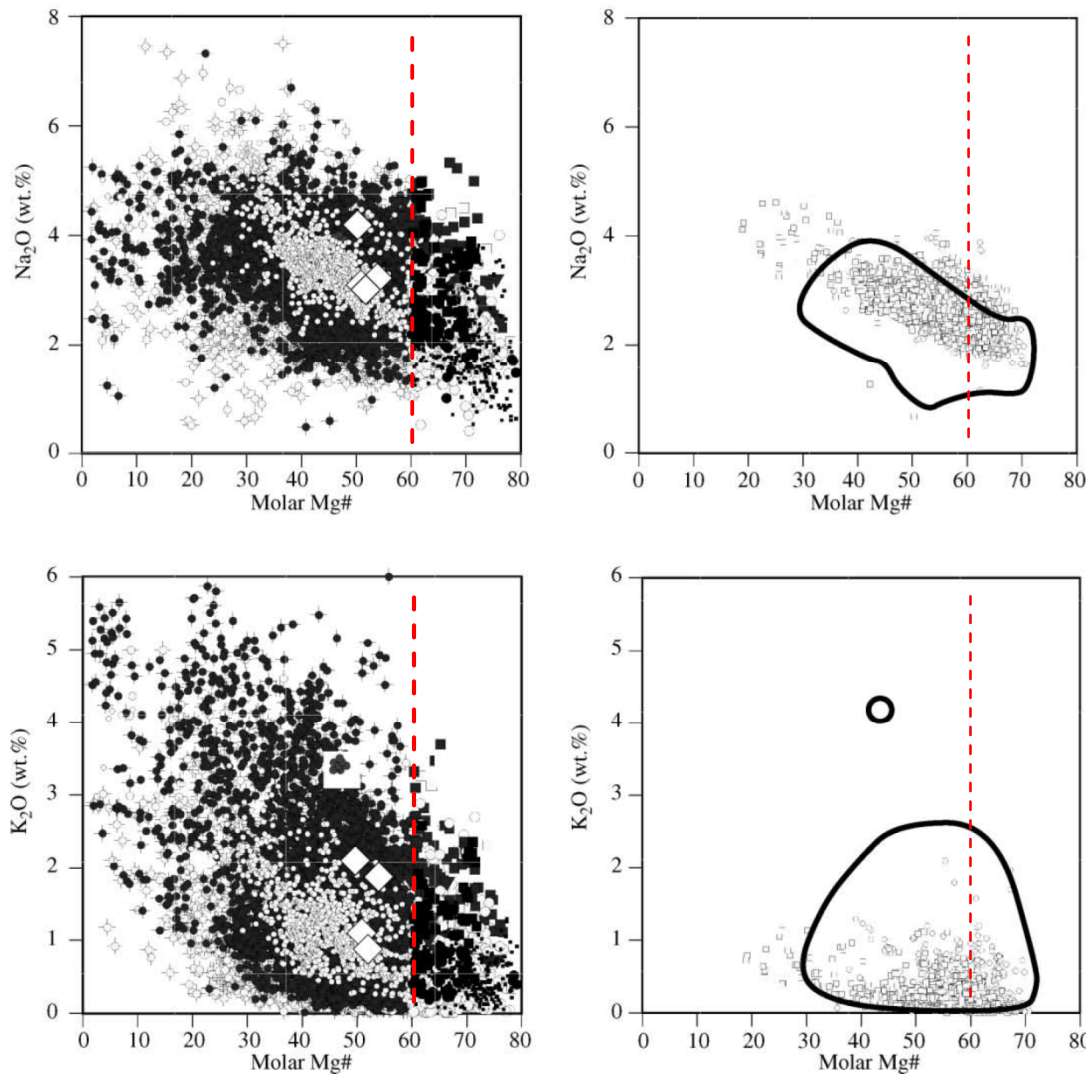


**CaO** in primitive arc lavas is similar or mostly **lower** than in MORB. Evolved series have much lower CaO than evolved MORB. *The term calc-alkaline series for arc magmas is therefore **misleading!!***

**MgO** values **exceed** those observed in the most **primitive MORB** lavas, but also decrease to much **lower** values

# Chemical signatures in island arc magmas

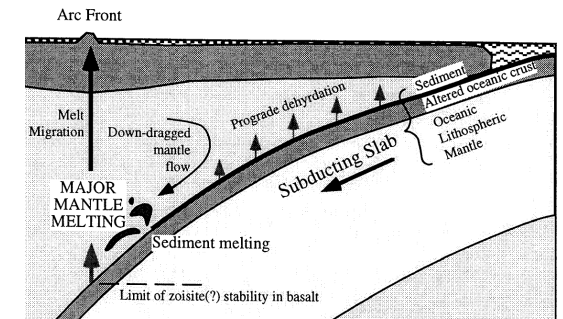
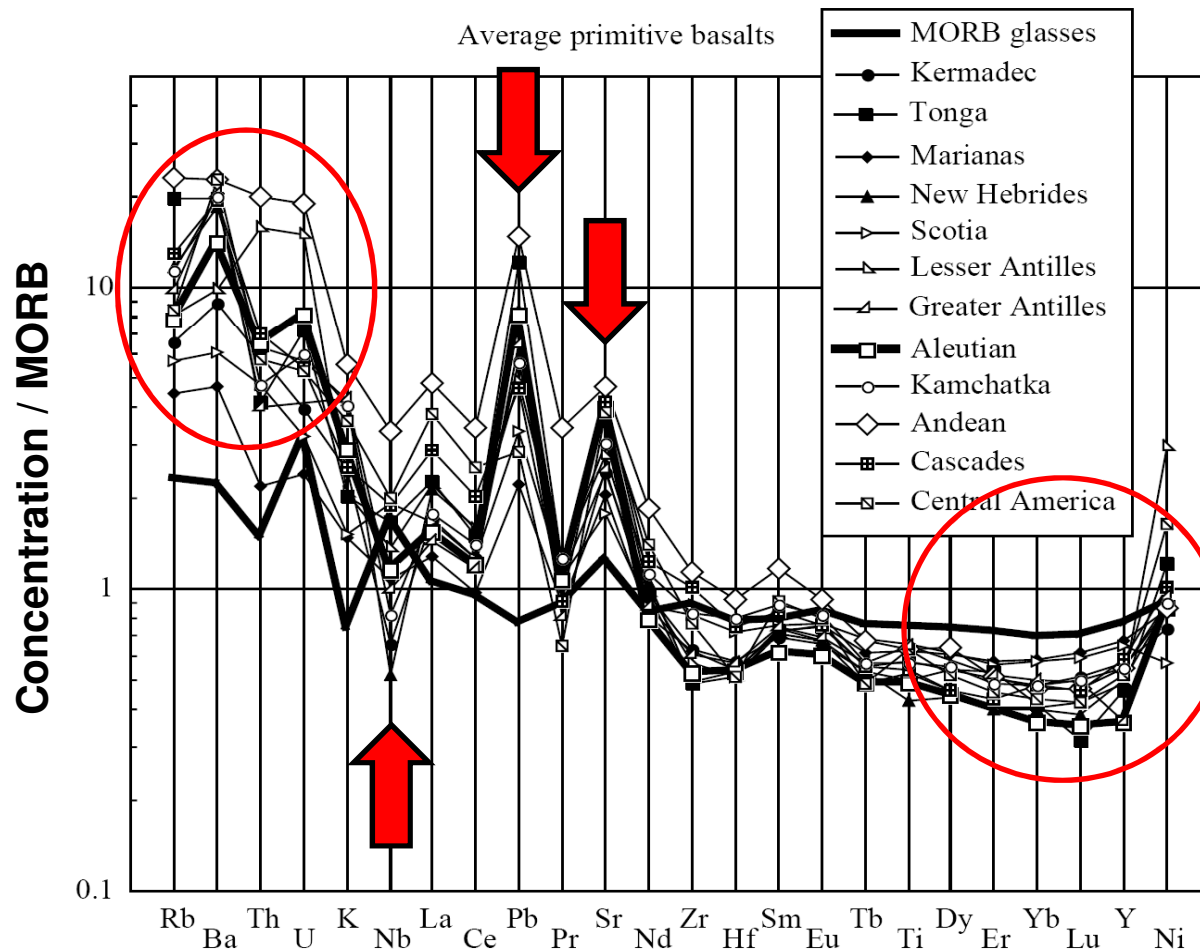
## Major element characteristics



**Na<sub>2</sub>O** and **K<sub>2</sub>O** in primitive arc magmas can be both, **lower and higher** than in primitive MORB. This variation is a **primary feature**, reflecting the involvement of more **different chemical endmembers** (sub-arc mantle, subducted lithosphere, fluids & sediments) during arc magma generation than during MORB formation (i.e., not simply fractionation!)

# Chemical signatures in island arc magmas

Trace element characteristics



Elliott et al., 1997

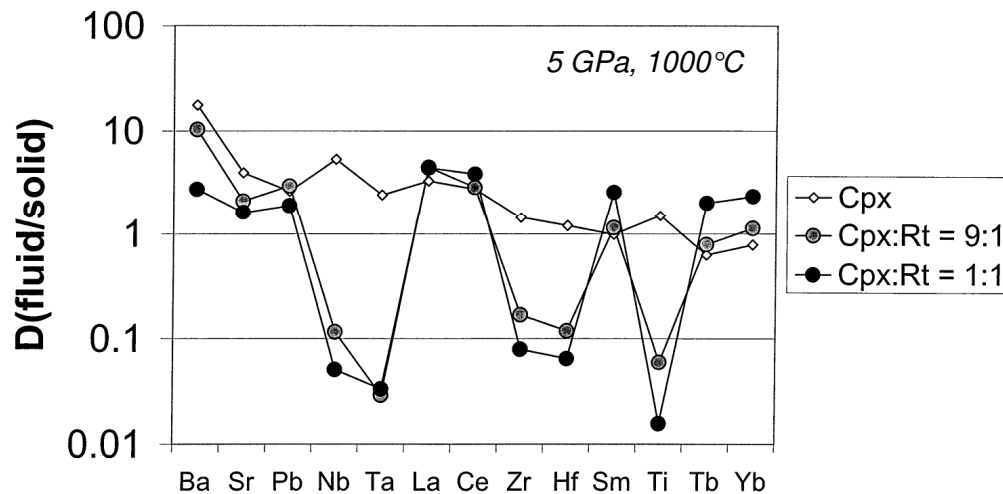
**Trace element**  
characteristics of **primitive**  
island arc magmas relative  
to MORB:

- Enrichment in LILE
- Depletion in Nb and Ta
- Pos. Pb + Sr anomalies
- Depletion in HREE

# Chemical signatures in island arc magmas

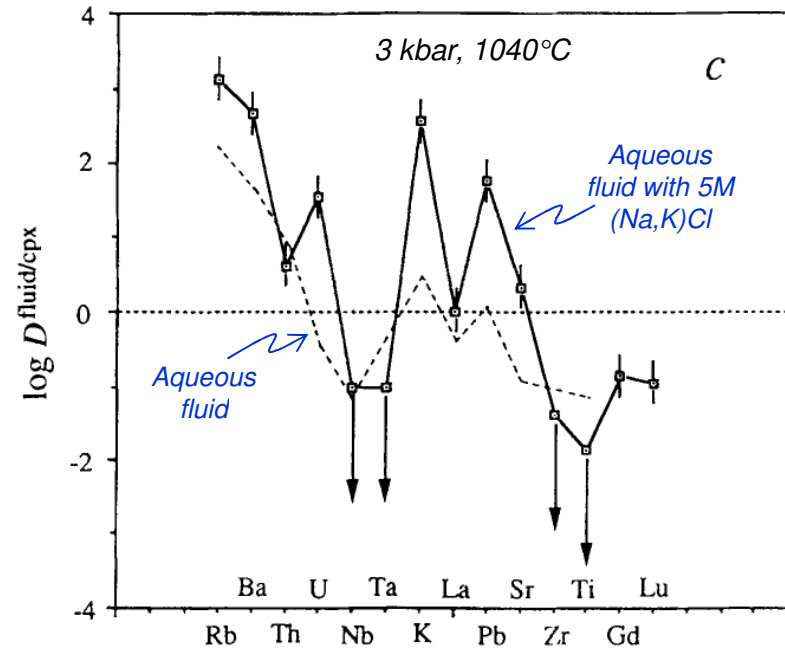
Causes of specific anomalies

## Fluid/Rutile partitioning



Partitioning of trace-elements between **fluid** and **cpx/rutile** at **5 GPa** (~150 km) and 1000°C. Note that all **HFSE** (Nb, Ta, Zr, Hf, Ti) **retain** in the residue as long as **rutile** is present

From: Stalder et al., 1998



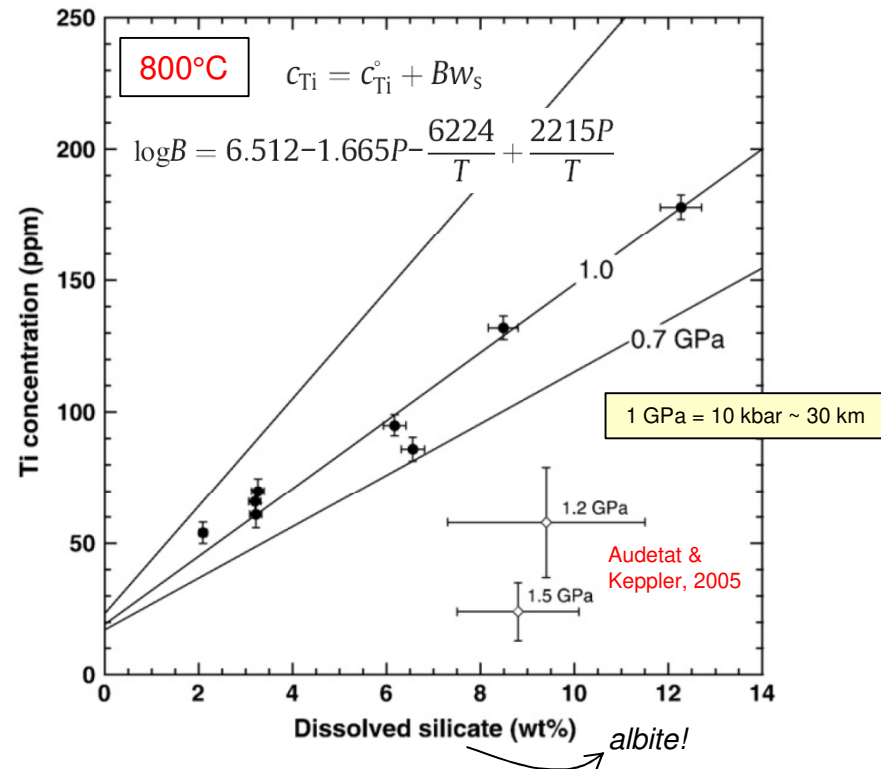
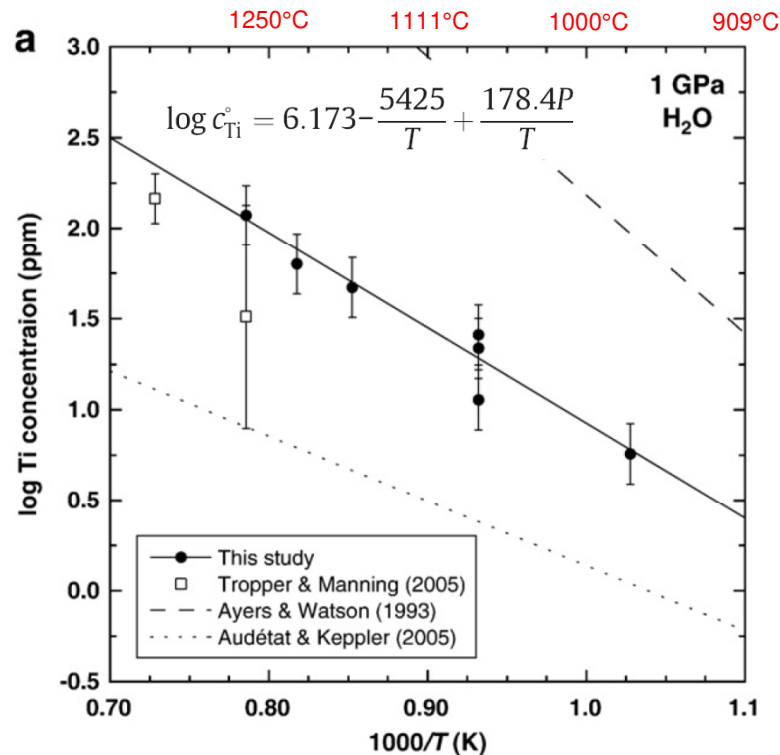
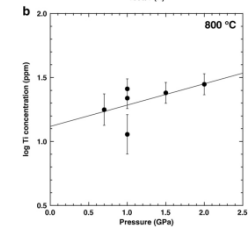
Partitioning of trace-elements between **fluid** and **clinopyroxene** at 3 kbar (~10 km) and 1040°C. **LILE, U, K, Pb** and  $\pm$ Th but not Sr were **enriched in the fluid**, **HFSE** (and **HREE**) remain in the **solid phase**

From Keppler, 1996, Nature

# Chemical signatures in island arc magmas

Causes of specific anomalies –  
rutile solubility in aqueous fluids

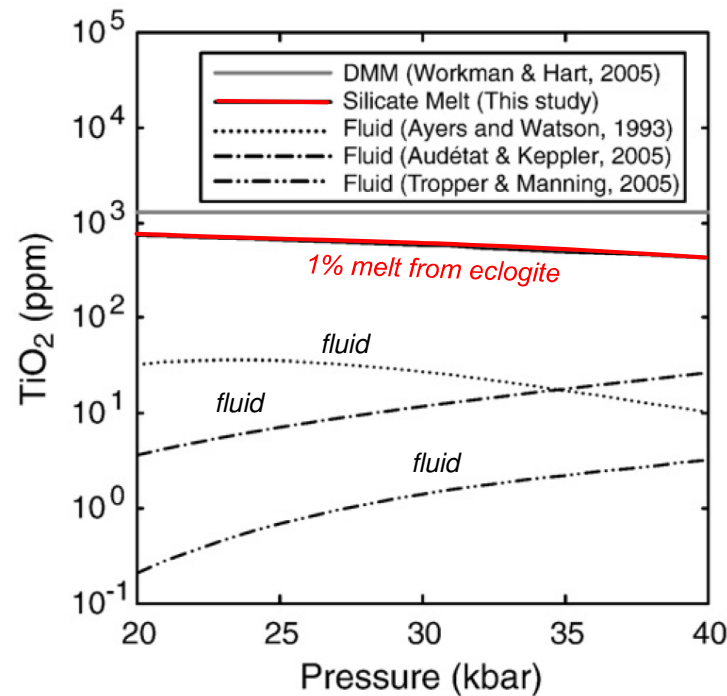
Rutile solubility only  
little depends on pressure!



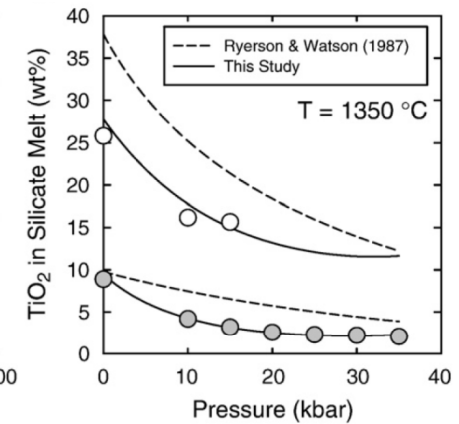
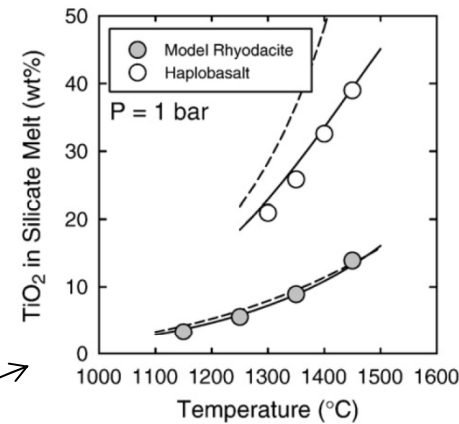
Solubility of **rutile** in **aqueous fluids** dependent on temperature, pressure and amount of **dissolved silicate** – **SOLUBILITY IS LOW!**

# Chemical signatures in island arc magmas

Causes of specific anomalies – rutile saturation in silicate melts (during eclogite melting)



TiO<sub>2</sub> concentration in a melt produced by **1% partial melting** of a MORB-type **eclogite** at H<sub>2</sub>O saturated conditions



Rutile saturation in model **rhyodacite** and **basalt** as a function of pressure and temperature

Gaetani et al., 2008, EPSL

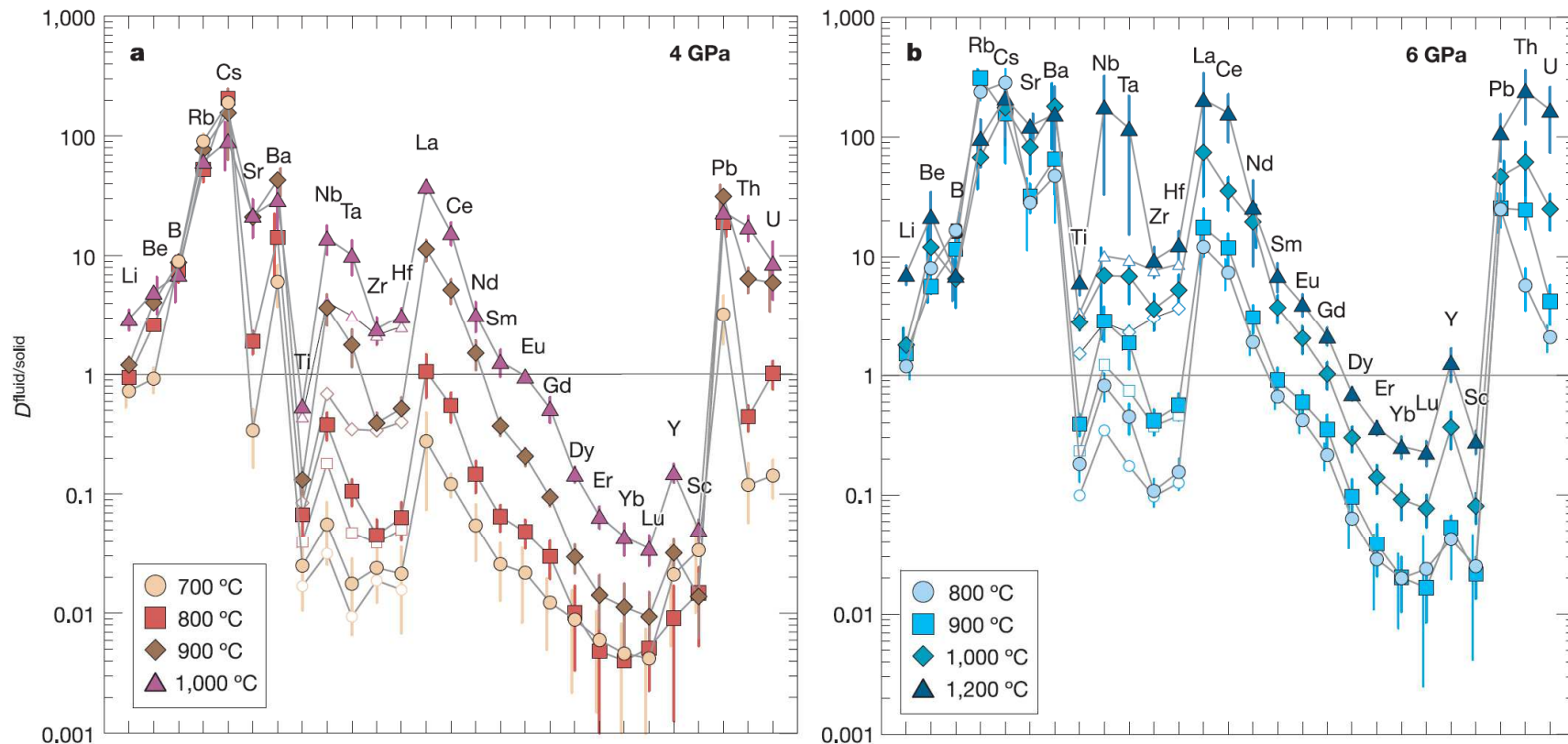
1 GPa = 10 kbar ~ 30 km

These experiments lead to this result and to this conclusion:

The **low** TiO<sub>2</sub> saturation concentration in melts formed potentially during **eclogite melting** during **slow subduction** suggests that **rutile will remain stable in the downgoing slab** !

# Chemical signatures in island arc magmas

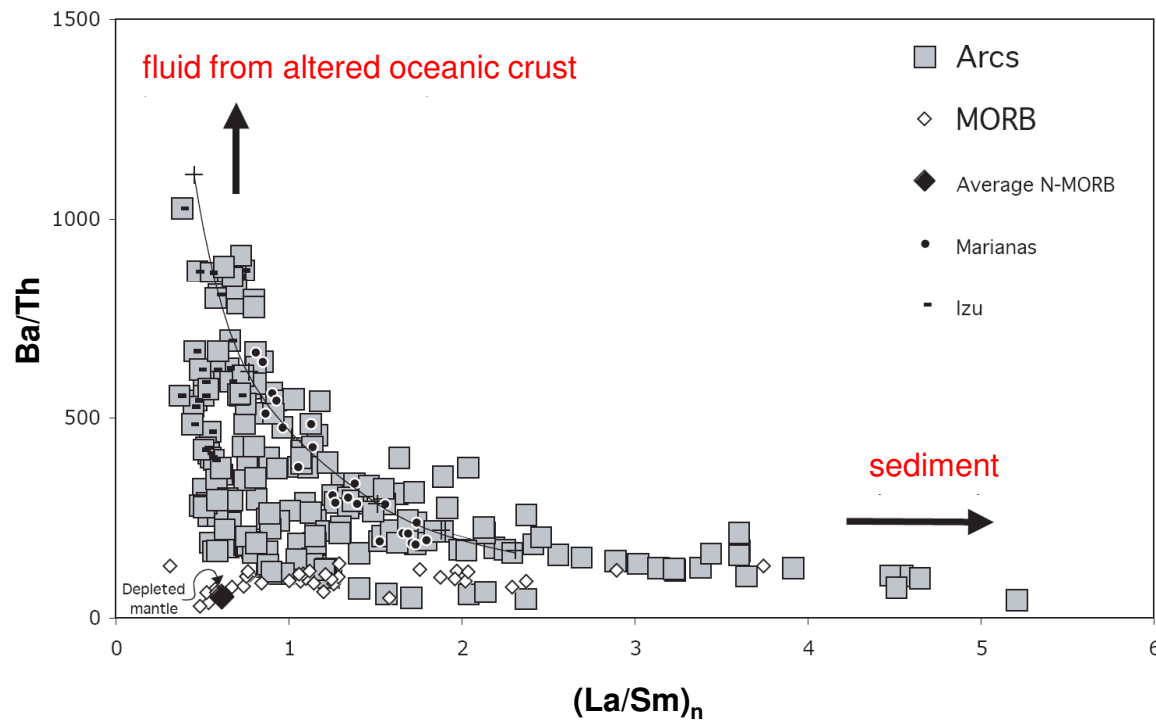
More experiments.... at ~ 5-6 GPa the solubility of most elements increases drastically



Partitioning of trace-elements between **fluids/supercritical liquids** and **residual eclogite** at 4 and 6 GPa (~120 and ~180 km)

# Chemical signatures in island arc magmas

Sediment vs. fluid signature in arc magmas

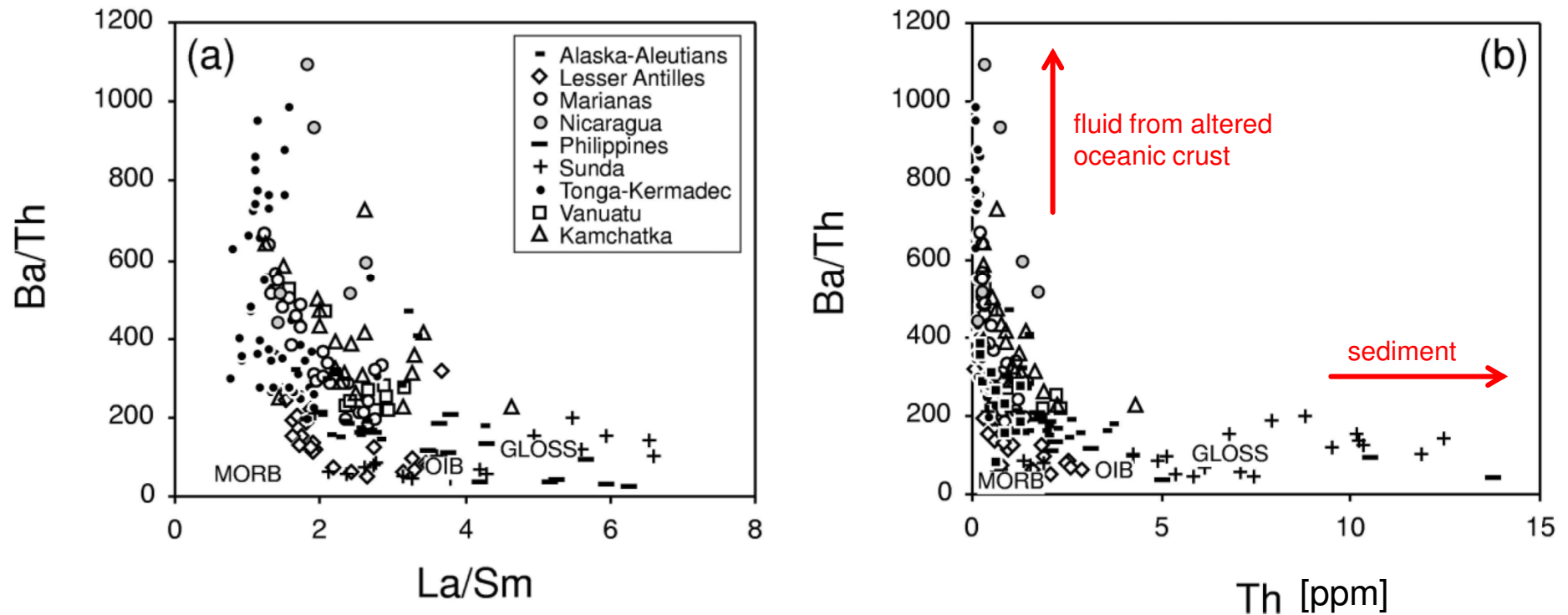


Elliott, 2003, *Tracers of the slab*

**Barium is fluid mobile,**  
 Th is not, but **Th is enriched in sediments**  
 over Ba. **La/Sm in fluids**  
 from altered mafic crust is **low** (i.e., mostly unfractionated relative to MORB), but **high** in **continent derived pelagic sediments (clay)**

# Chemical signatures in island arc magmas

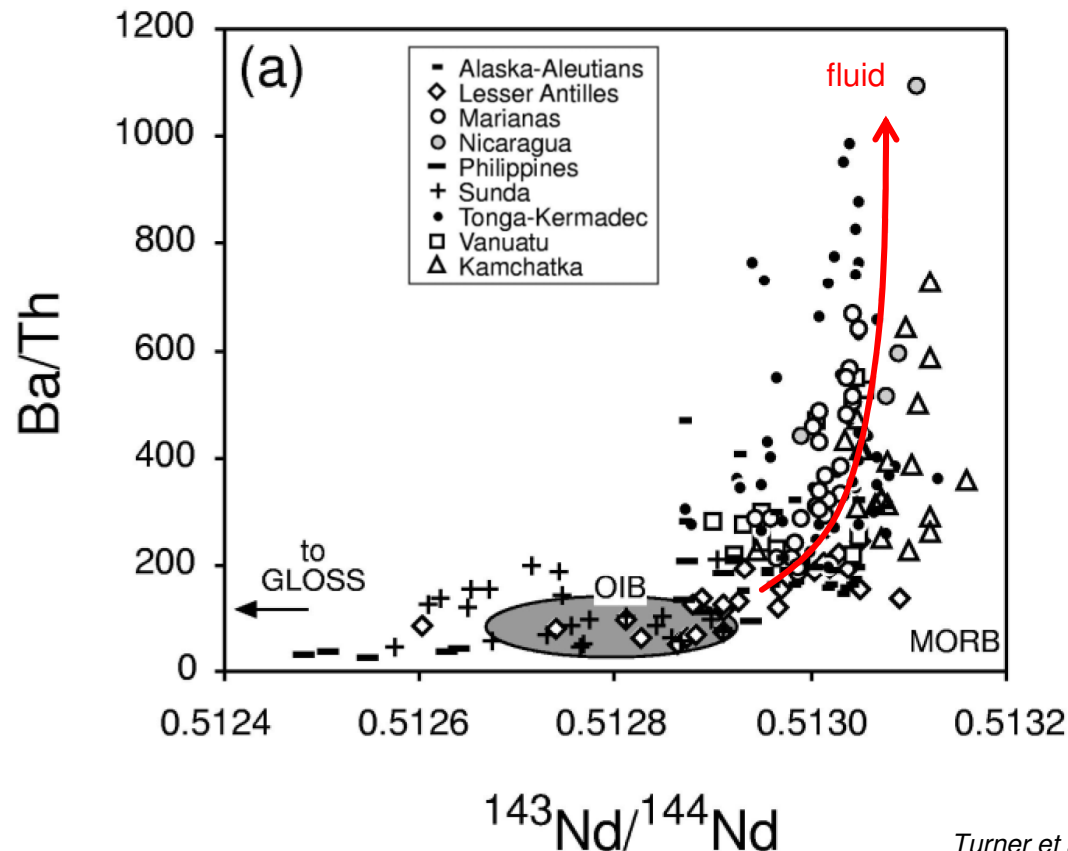
Sediment vs. fluid signature in arc magmas



Ba/Th vs. La/Sm and Ba/Th vs. Th concentration in a number of island arcs. As La/Sm, **Th concentrations in pelagic sediments are high!**

# Chemical signatures in island arc magmas

Sediment vs. fluid signature in arc magmas ... where does the fluid come from?

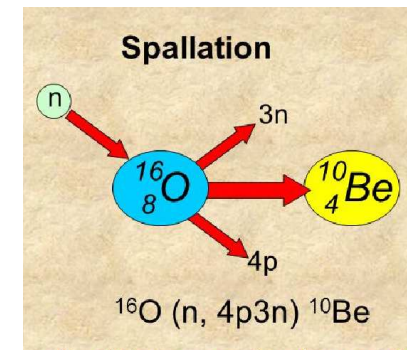


Radiogenic Nd-isotope ratios, i.e. **high**  $^{143}\text{Nd}/^{144}\text{Nd}$  in samples with **high Ba/Th** indicate that the fluid component is predominantly derived from the **altered oceanic lithosphere**, not from the subducted sediment. It also implies that **Nd is moderately mobile** during slab dehydration.

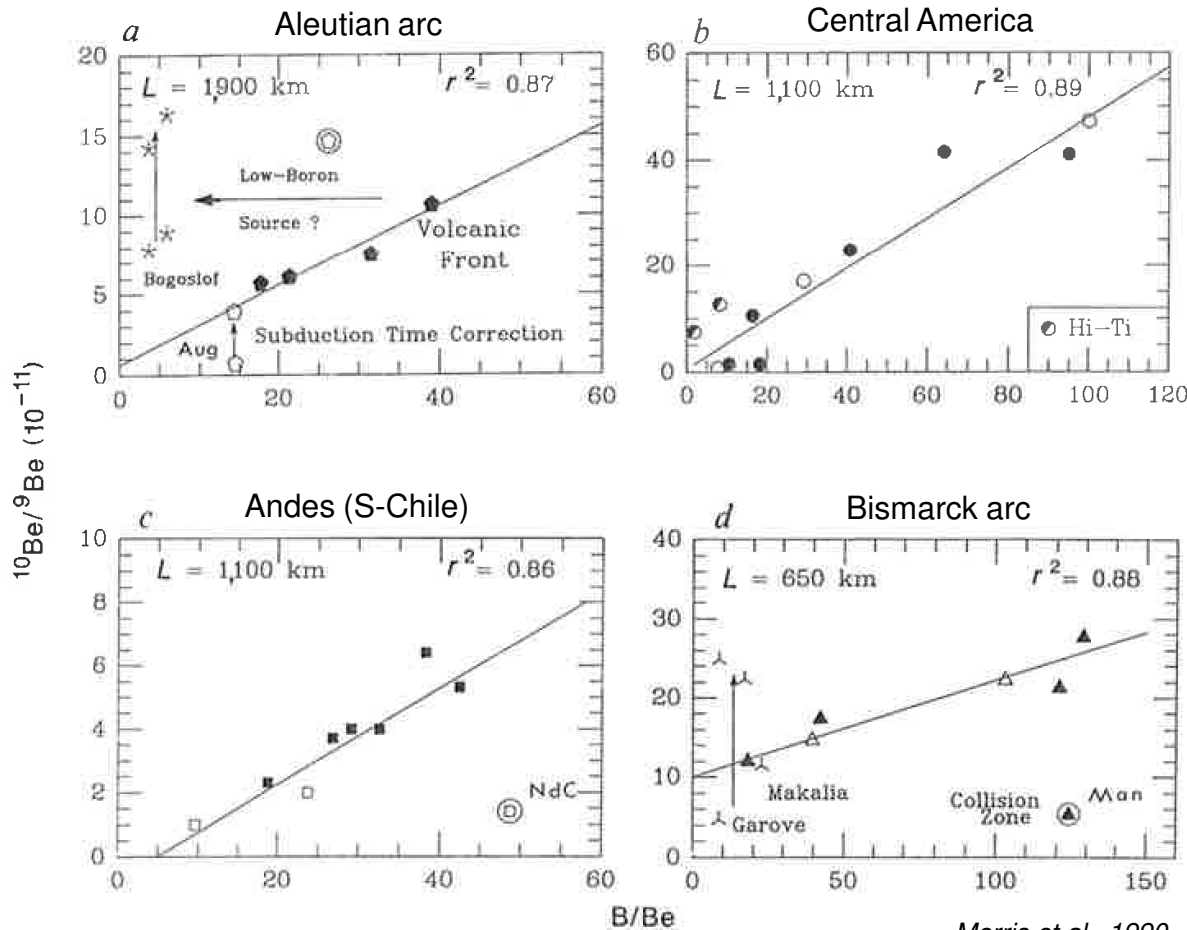
# Chemical signatures in island arc magmas

(Time) constraints from B,  $^{10}\text{Be}$  and U-Th disequilibrium

Formation of  $^{10}\text{Be}$  in the atmosphere caused by cosmic rays



Half life of  $^{10}\text{Be}$ : ~1.5 Ma



Morris et al., 1990

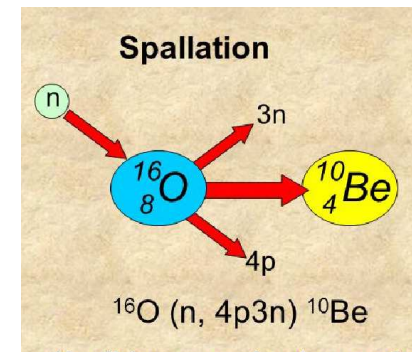
$^{10}\text{Be}/^9\text{Be}$  and B/Be ratios in **arc lavas** from four different settings.

Note the overall linear correlations at **different slopes** and the **variable ranges in B/Be**

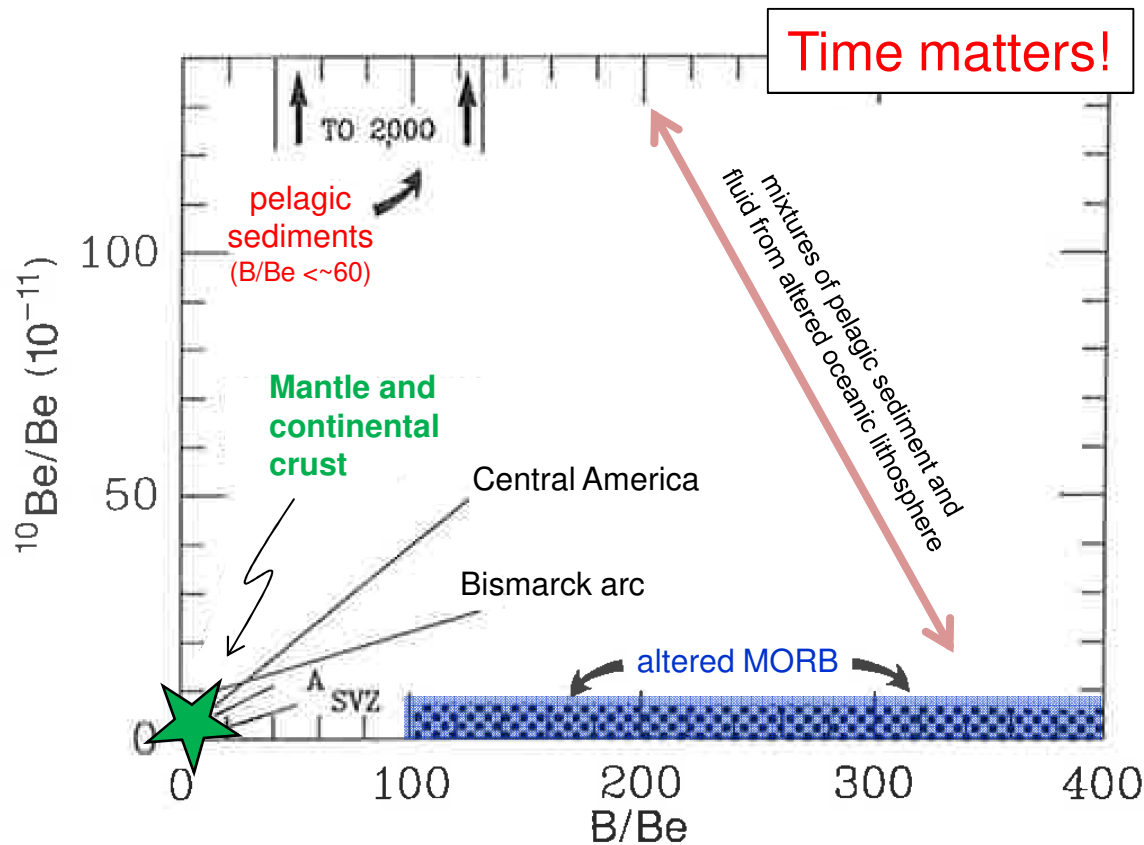
# Chemical signatures in island arc magmas

(Time) constraints from B,  $^{10}\text{Be}$  and U-Th disequilibrium

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Half life of  $^{10}\text{Be}$ : ~1.5 Ma



Morris et al., 1990

$^{10}\text{Be}/^9\text{Be}$  and B/Be ratios of **different components** (altered mafic crust, sediments & sub-arc mantle) contributing to the formation of **island arc magmas.**

Boron in sea water: ~5 ppm

Boron in sediments: ~50-150 ppm

B in MORB and OIB: ~1-3 ppm

B in altered oceanic crust: ~10-300 ppm

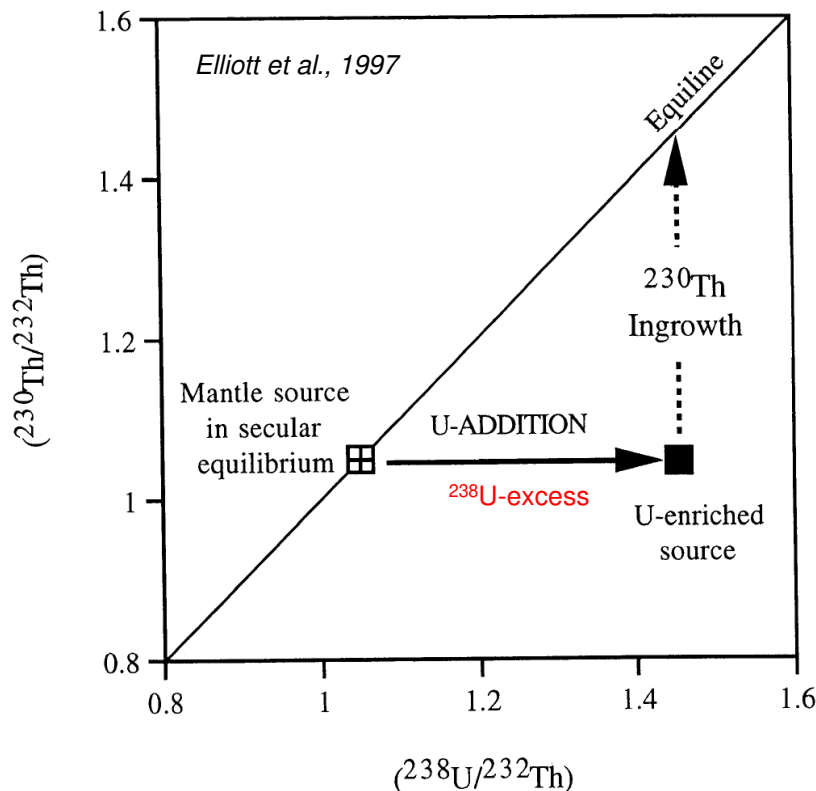
$^9\text{Be}$  is quite uniform in all reservoirs!

# Chemical signatures in island arc magmas

More time constraints from U-Th disequilibrium

$$\left(\frac{^{230}\text{Th}}{^{232}\text{Th}}\right)_P = \left(\frac{^{230}\text{Th}}{^{232}\text{Th}}\right)_I e^{-\lambda_{230}t} + \frac{^{238}\text{U}}{^{232}\text{Th}}(1 - e^{-\lambda_{230}t})$$

*measured*
*„initial“=f(t)*
*„ingrowth“=f(t)*



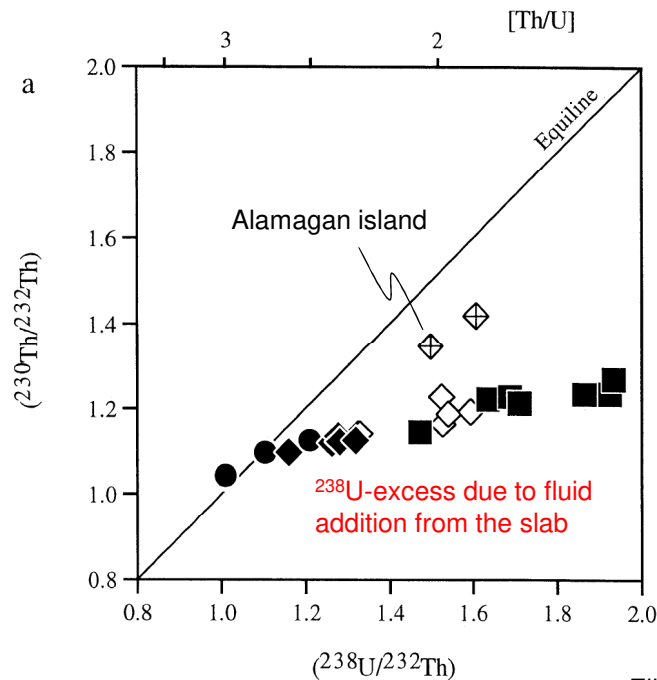
Equiline diagram. Samples in **secular equilibrium** with respect to  $^{238}\text{U}$  and  $^{230}\text{Th}$  [i.e.  $A(^{238}\text{U}) = A(^{230}\text{Th})$ ] plot along the **equiline**.

Any elemental fractionation between U and Th moves a point (sample) **away from the equiline**, either to the right (**U-addition**), or to the left (**Th-addition**)

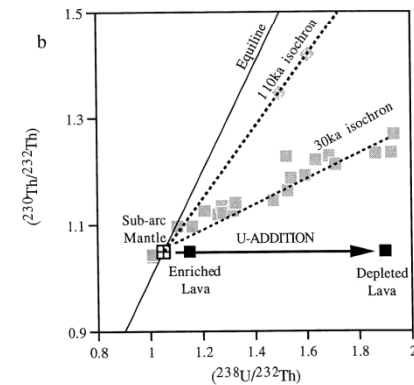
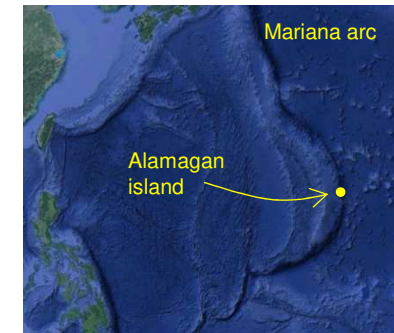
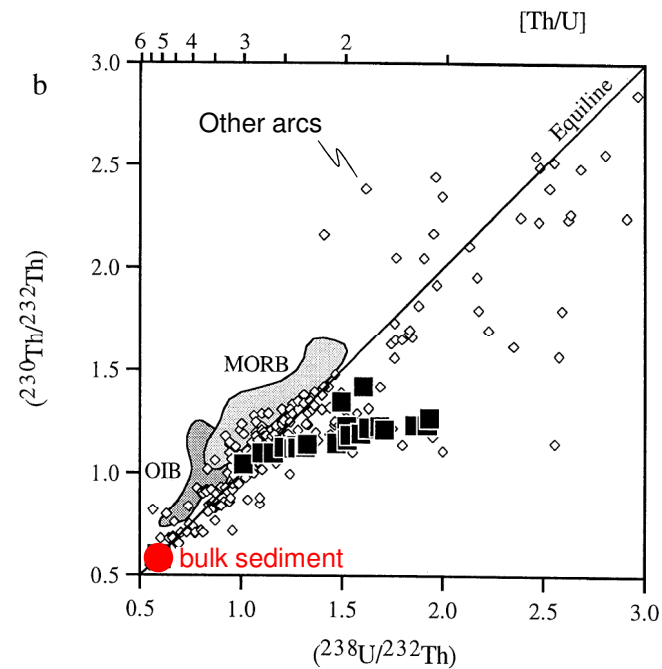
*Half life of  $^{230}\text{Th}$ : ~75 ka*

# Chemical signatures in island arc magmas

## U-Th disequilibrium in Mariana arc lavas



Elliott et al., 1997



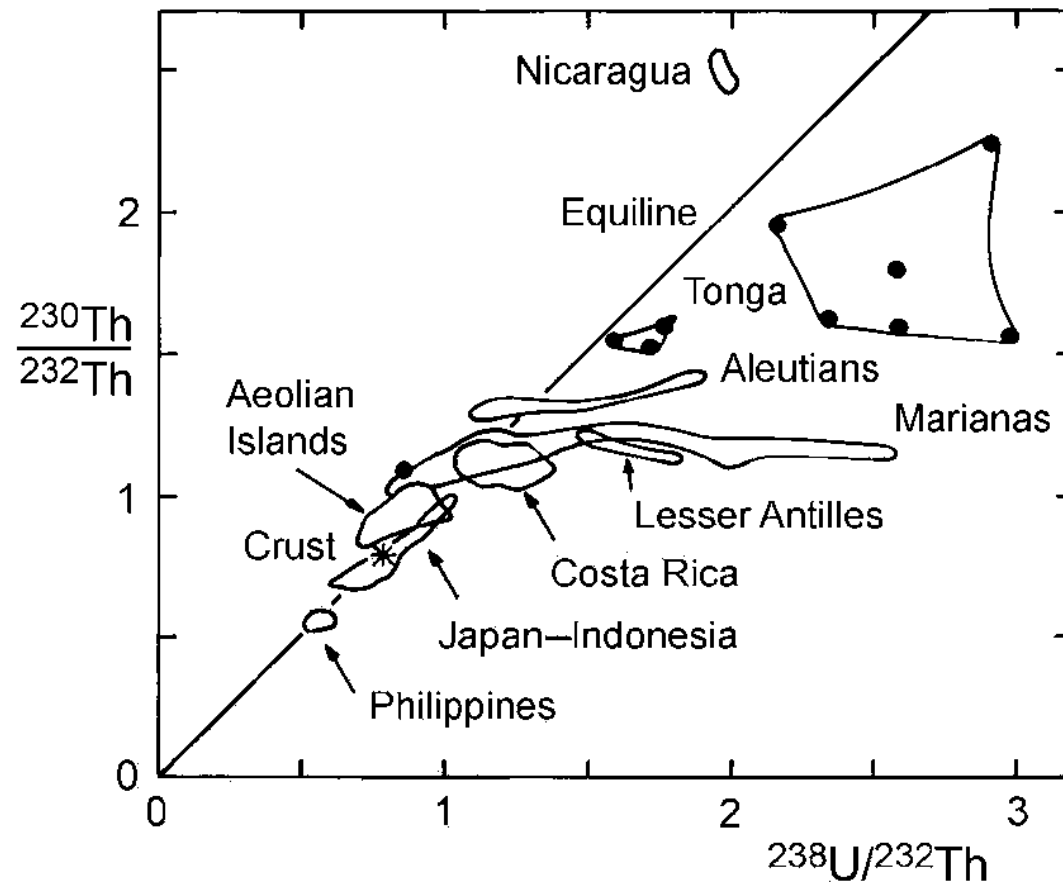
**Mariana arc** samples plot distinctly off the equiline, indicating that **fluid-controlled U-transfer, melting, melt migration** and **eruption** of Mariana arc lavas are „fast“ processes!

Secular disequilibrium is a **unique feature** observed in **island arc lavas** (note the slight offset of MORB and OIB due to Th/U fractionation during partial melting)

Assuming variably enriched and depleted sources with constant Th/U (or variable amounts of U-addition), secular disequilibrium was imprinted ~30 ka ago

# Chemical signatures in island arc magmas

U-Th disequilibrium in arc lavas



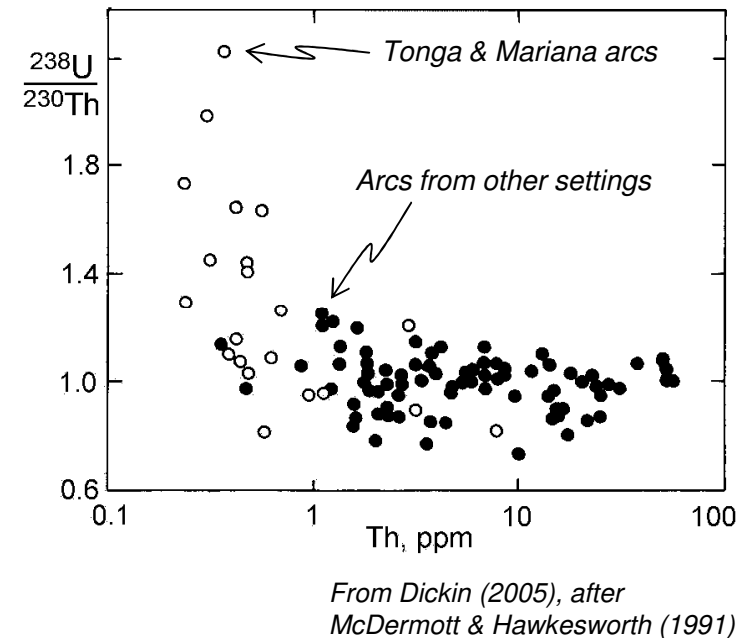
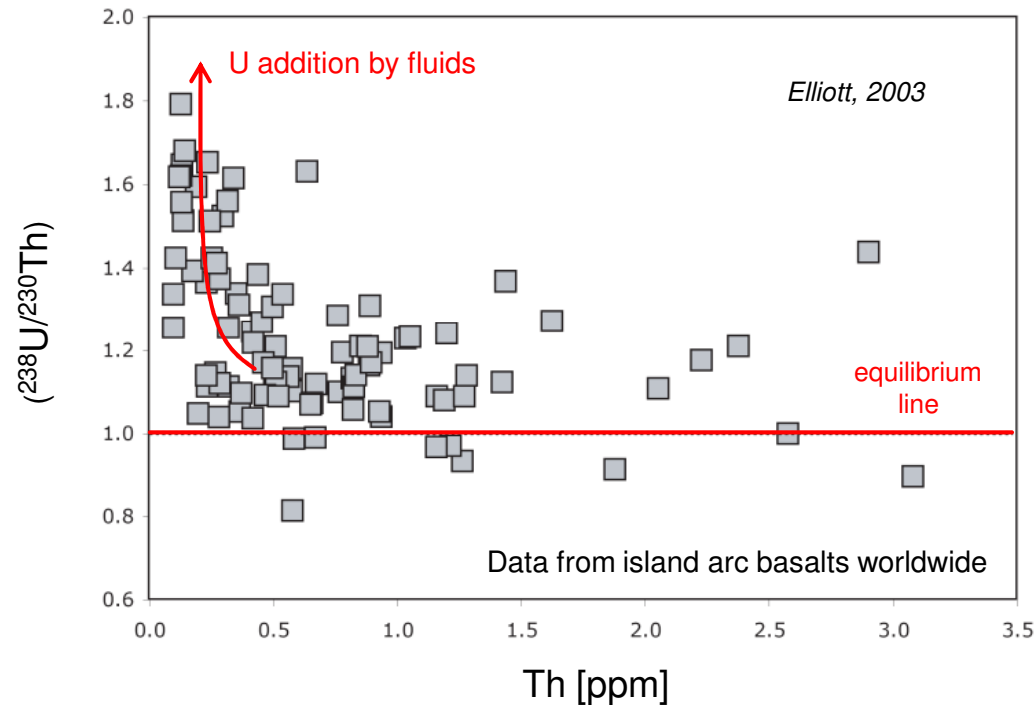
From Dickin (2005), after  
Hawkesworth et al. (1991)

**U-Th disequilibrium in island arc magmas from different locations**

The **degree of disequilibrium** is caused by different amounts of U added from the slab, or by **differing times** between mantle source metasomatism and lava eruption at the surface

# Chemical signatures in island arc magmas

## U-Th disequilibrium in island arc magmas

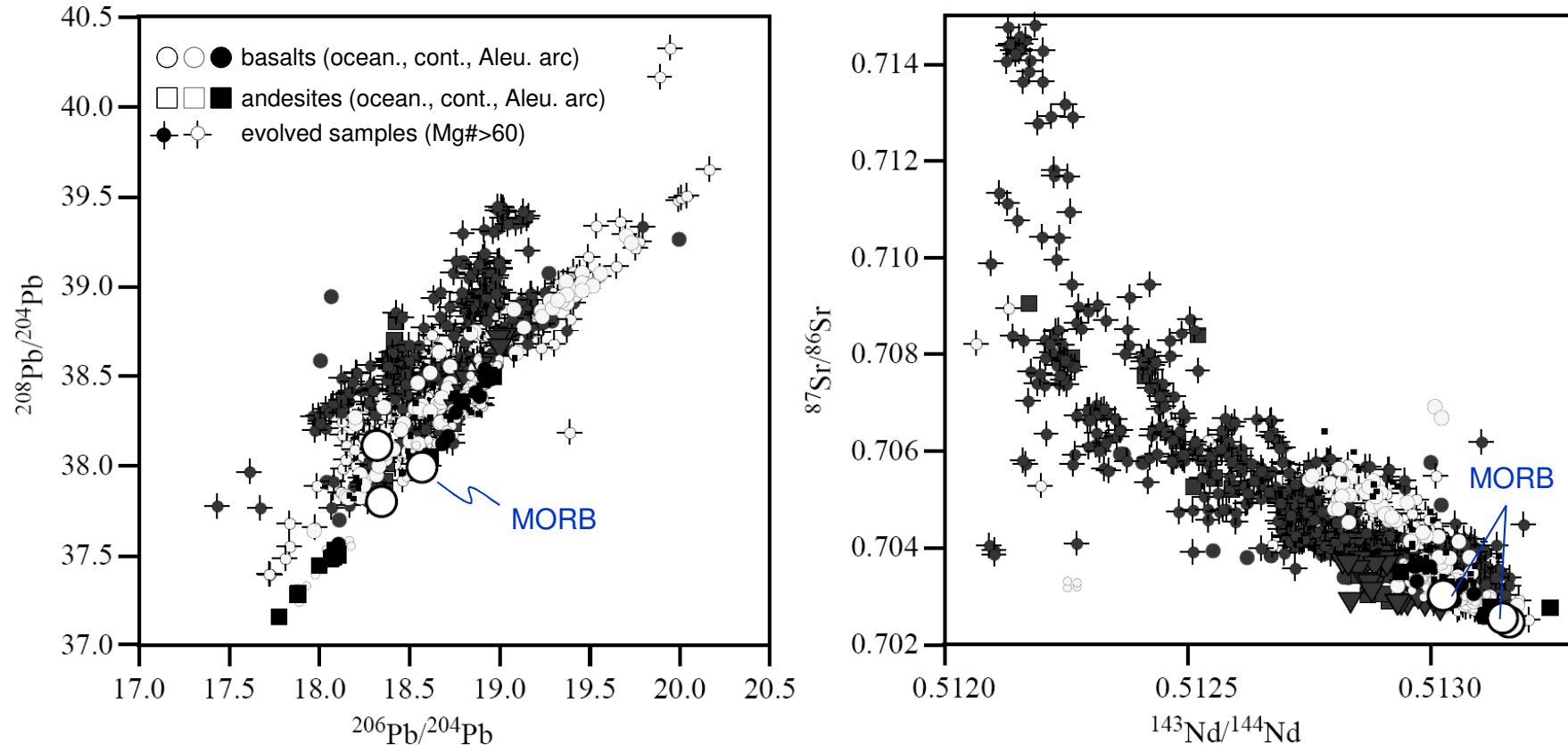


**Secular disequilibrium** between  $^{238}\text{U}$  and  $^{230}\text{Th}$  is **strongest** in island arc lavas with **low Th contents**. This confirms that **U addition** is predominantly by **fluids** released from the downgoing oceanic lithosphere

# Chemical signatures in island arc magmas

## Overall radiogenic isotope signatures

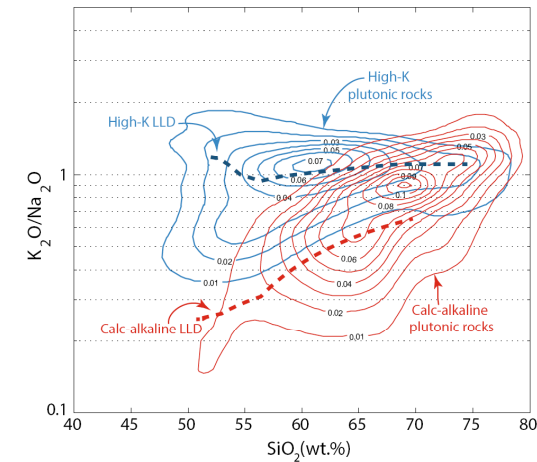
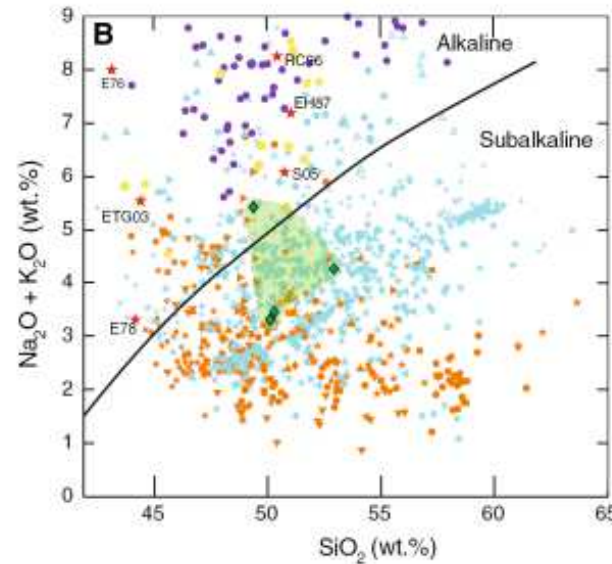
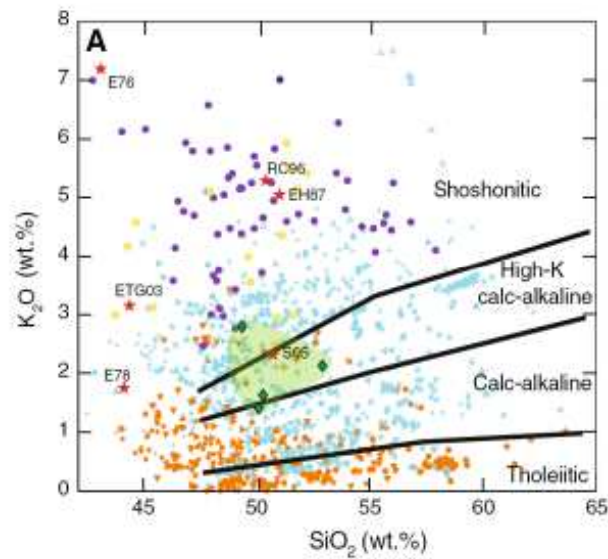
*Kelemen et al., 2003, Treatise on Geochemistry*



Pb-isotopes are mostly enriched relative to MORB, with distinctly higher  $^{208}\text{Pb}$  at given  $^{206}\text{Pb}$ , indicating a long-term **high Th/U** in the source of most arc lavas. The **Nd-Sr isotope array** reflects the aforementioned involvement of subducted **continental sediments** during magma genesis

# Chemical signatures in island arc magmas

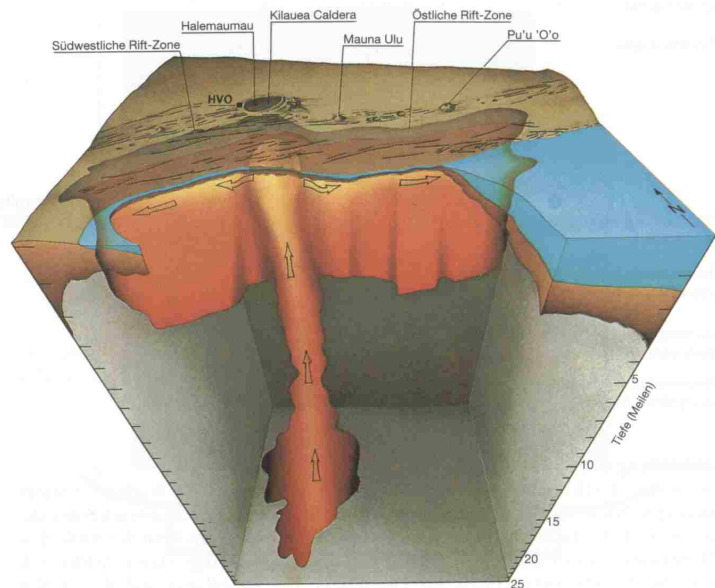
Liquid lines of descent in calc-alkaline magmas



- |  |                         |                          |
|--|-------------------------|--------------------------|
| ◆ Dariv Primitive Melts                                    | <b>Oceanic Arcs</b>     |                          |
| ★ Biotite-Saturated High-K Experimental Starting Materials | ▲ Aeolian               | ▲ Kurile                 |
| <b>Biotite-bearing High Mg# lavas</b>                      | ▲ Aleutian              | ▲ Tonga & Kermadec       |
| ● Sierra Nevadas   | ▲ Bismarck (Solomon)    | ▲ Lesser Antilles        |
| ● Mexican Volcanic Belt                                    | ● Izu-Bonin             | ▲ Marianas               |
|  |                         | ▲ Vanuatu (New-Hebrides) |
|  |                         | ▲ Scotia                 |
|  |                         | ★ Yap                    |
|  | <b>Continental Arcs</b> |                          |
|  | ● Aegean                | ● Honshu                 |
|  | ● Andean                | ● Ryuku                  |
|  | ● Sunda-Banda           | ● Kamchatka              |
|  | ● Cascades              | ● Mexican Volcanic Belt  |
|  | ● Central America       | ● New Zealand            |
|  |                         | ● Papua New Guinea       |
|  |                         | ● Italy                  |
|  |                         | ● Luzon                  |
|  |                         | ● Sulawesi               |

Chapter 4

# Ocean islands in brief...



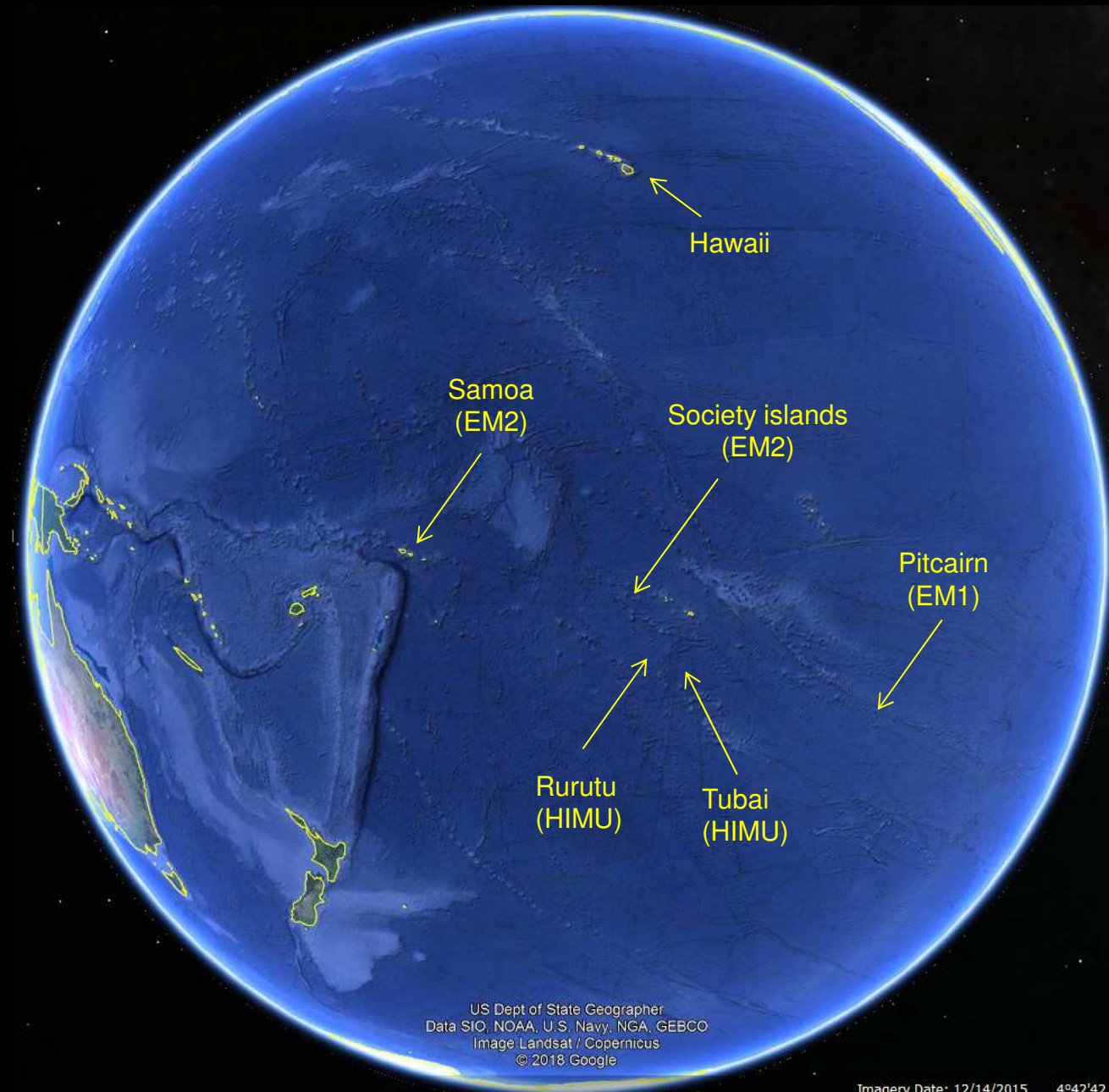
*Hawaii*

Ocean Island Basalts = OIB

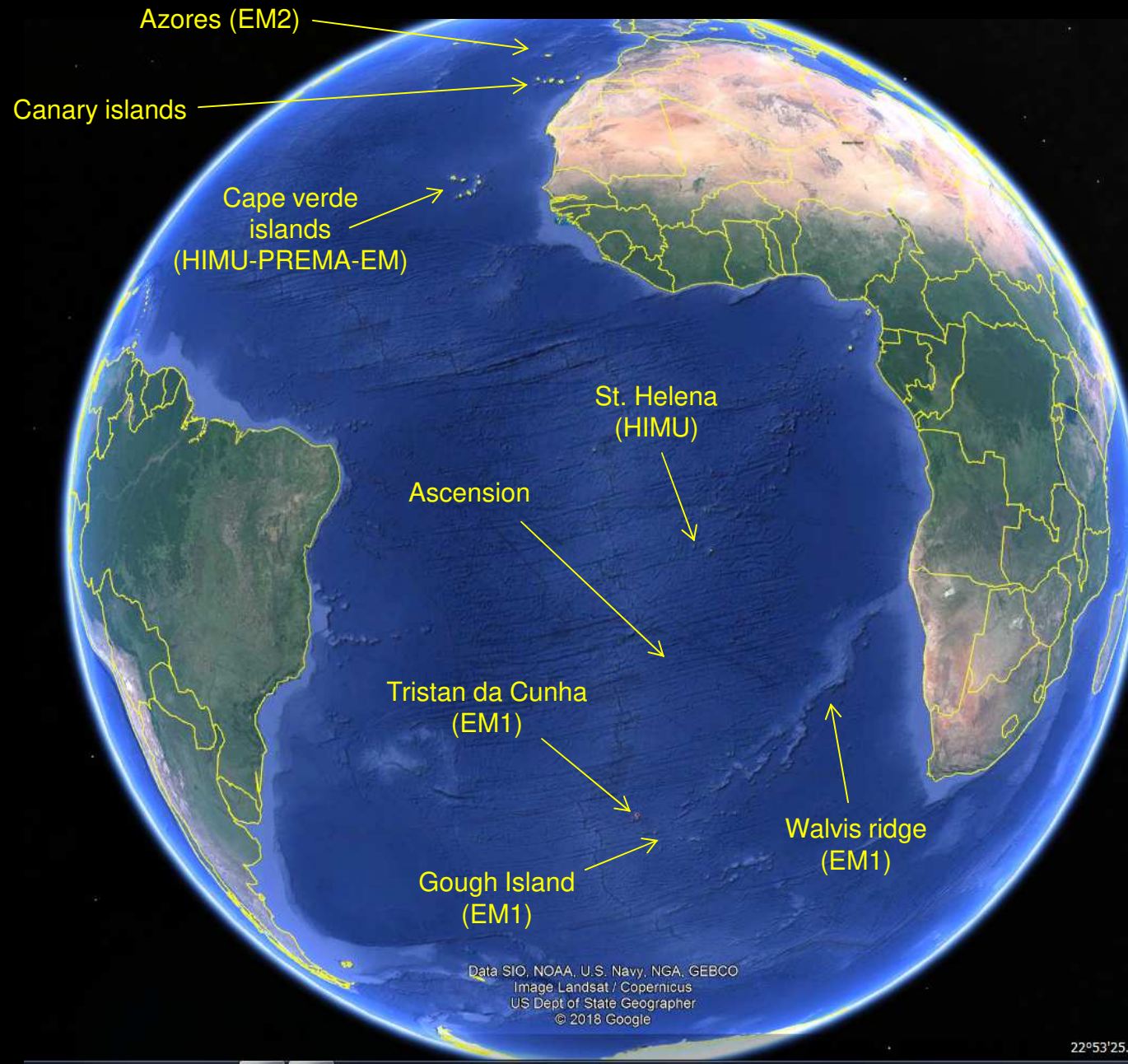
Ocean island basalts = oceanic intraplate basalts

Continental basalts = continental intraplate basalts

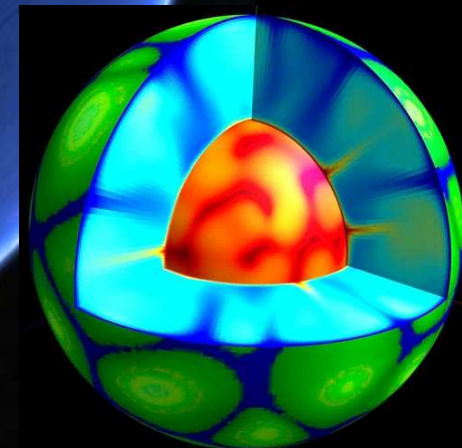
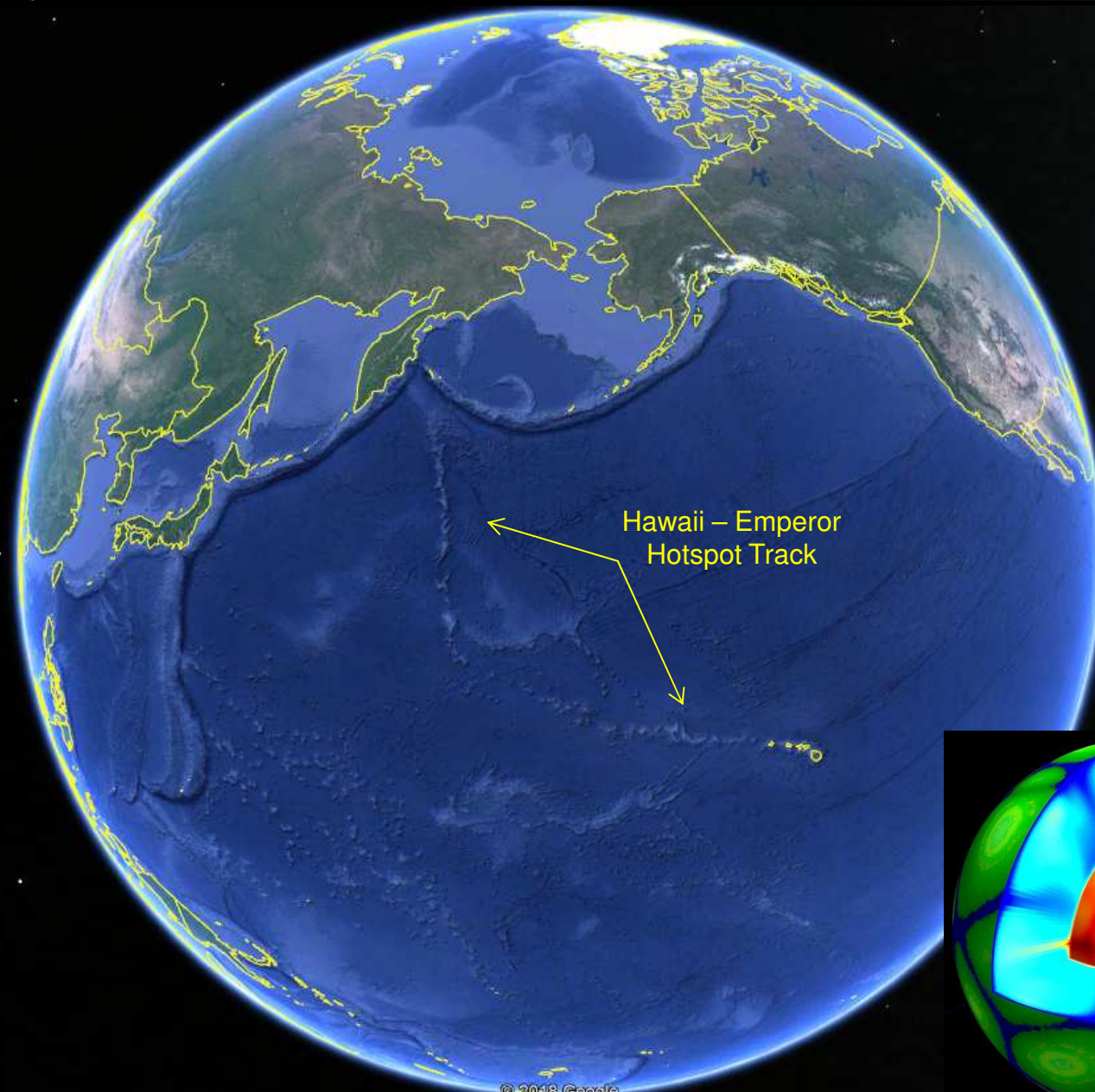
# Ocean islands



# Ocean islands



# Ocean islands



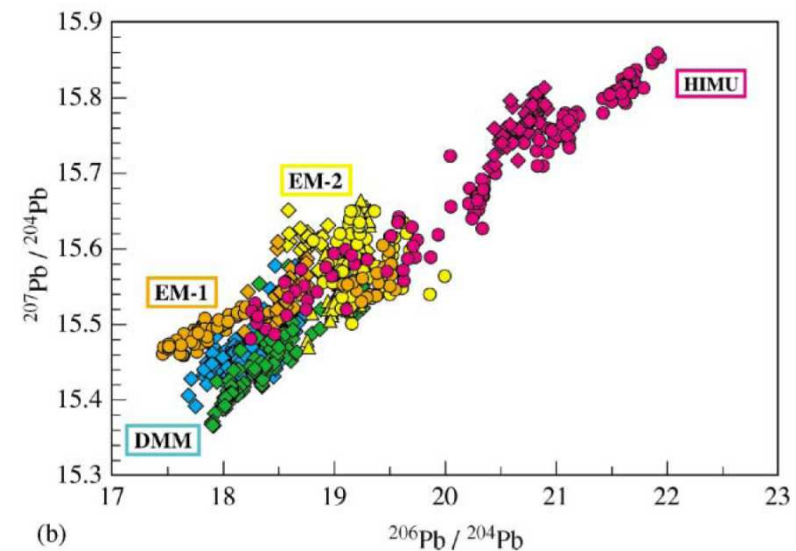
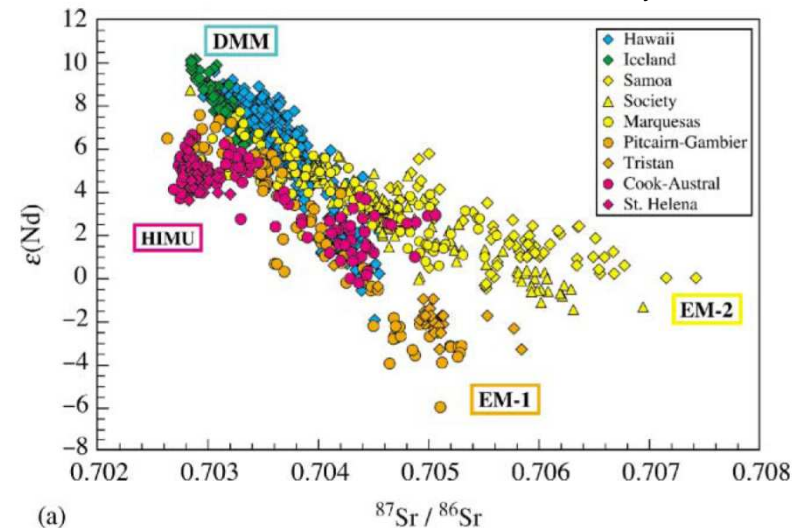
# Composition of ocean island basalts

## Radiogenic isotope features

As firstly outlined in 1964 by **Gast, Tilton & Hedge**, the strongly different isotope signatures in OIB if compared to MORB cannot be explained by different melting depths.

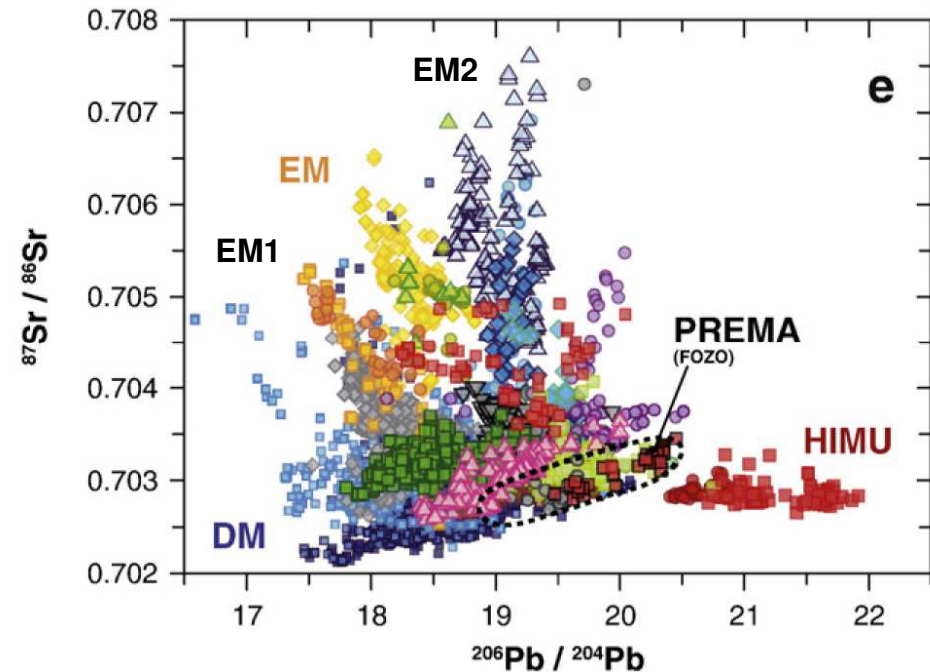
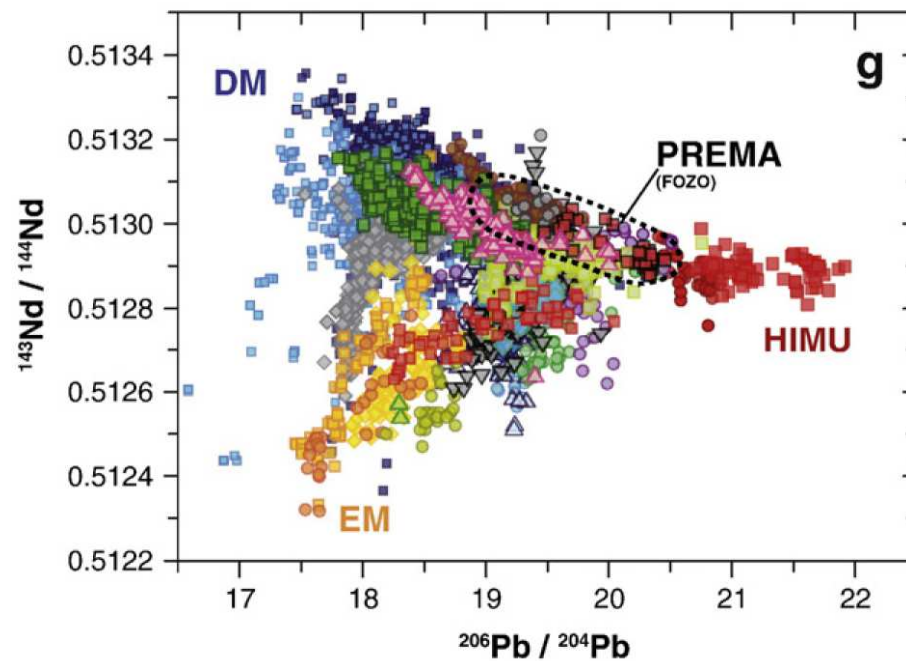
Instead the OIB mantle source must be chemically different, and in some respect been isolated from the MORB mantle source for millions to billions of years!

*Hofmann, 2003, Treatise on Geochemistry*



# Composition of ocean island basalts

## Radiogenic isotope features

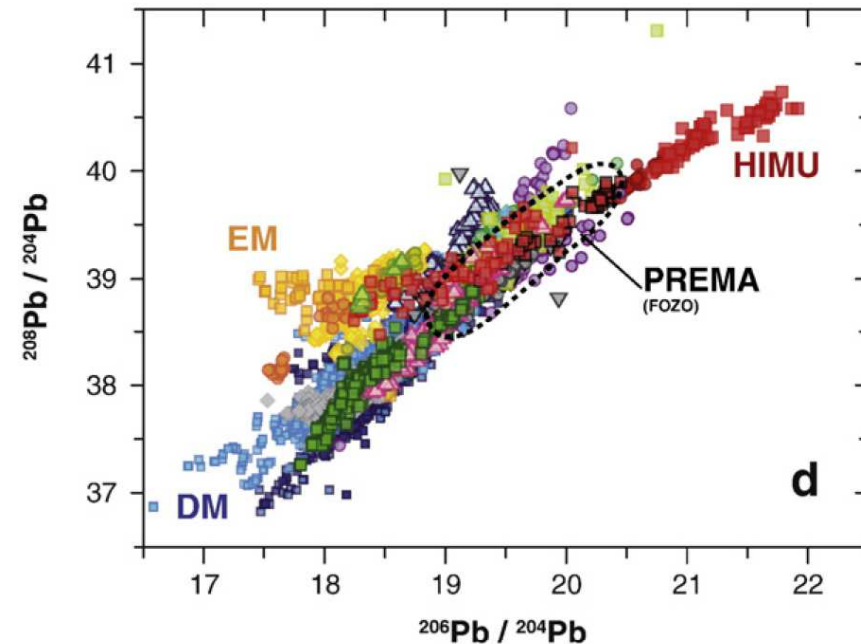
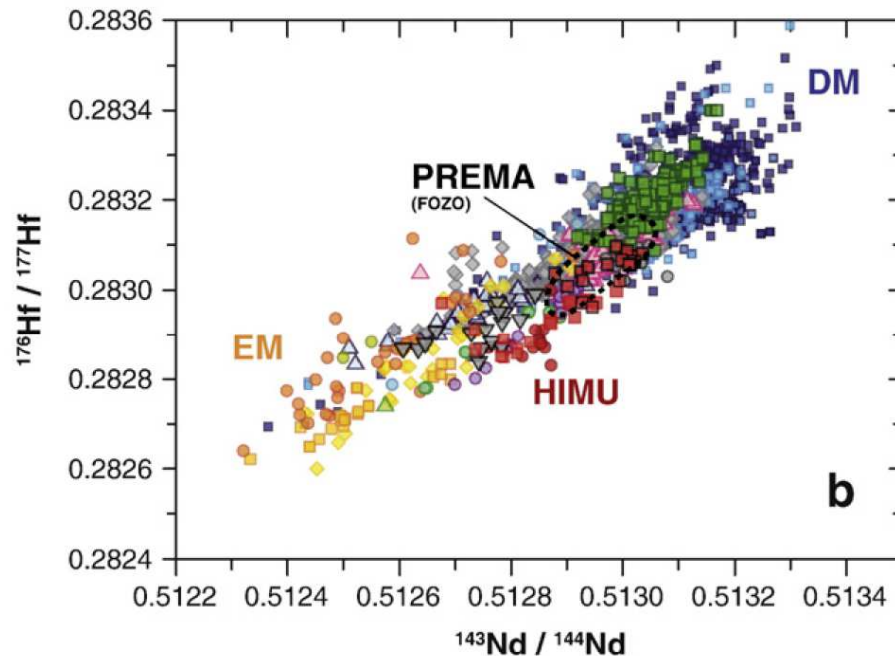


- |                   |                    |                       |
|-------------------|--------------------|-----------------------|
| ■ Atlantic MORB   | ● St. Helena       | ▼ Cape Verdes         |
| ■ Pacific MORB    | ● Kerguelen Heard  | ◆ Fernando de Noronha |
| ■ Indian MORB     | ■ Pitcairn         | ● Madeira             |
| ◆ Hawaii          | ● Walvis Ridge     | ■ Canary Islands      |
| ▲ Galapagos       | ● Tristan da Cunha | ● Ascension           |
| ● Society Islands | ▲ Gough            | ■ Iceland             |
| ▲ Samoa           | ● Comores          | ■ Austral Cook: HIMU  |
| ◆ Marquesas       | ● Azores           | ■ Austral Cook: PREMA |
|                   |                    | ■ Austral Cook: EM    |

Stracke 2012

# Composition of ocean island basalts

## Radiogenic isotope features

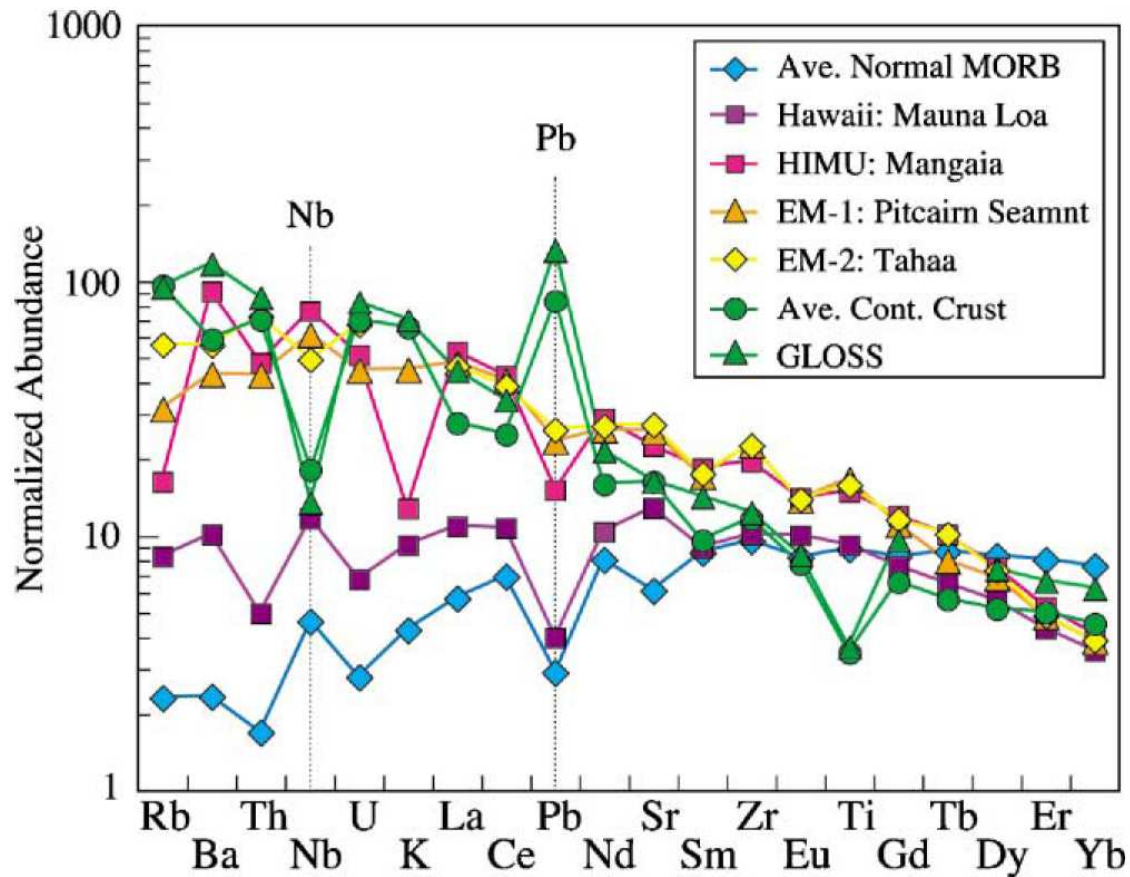


Stracke 2012

- |                   |                    |                       |
|-------------------|--------------------|-----------------------|
| ■ Atlantic MORB   | ● St. Helena       | ▼ Cape Verdes         |
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| ■ Indian MORB     | ● Pitcairn         | ● Madeira             |
| ◆ Hawaii          | ● Walvis Ridge     | ● Canary Islands      |
| ▲ Galapagos       | ● Tristan da Cunha | ● Ascension           |
| ● Society Islands | ▲ Gough            | ■ Iceland             |
| ▲ Samoa           | ● Comores          | ■ Austral Cook: HIMU  |
| ◆ Marquesas       | ● Azores           | ■ Austral Cook: PREMA |
|                   |                    | ■ Austral Cook: EM    |

# Composition of ocean island basalts

## Trace element features



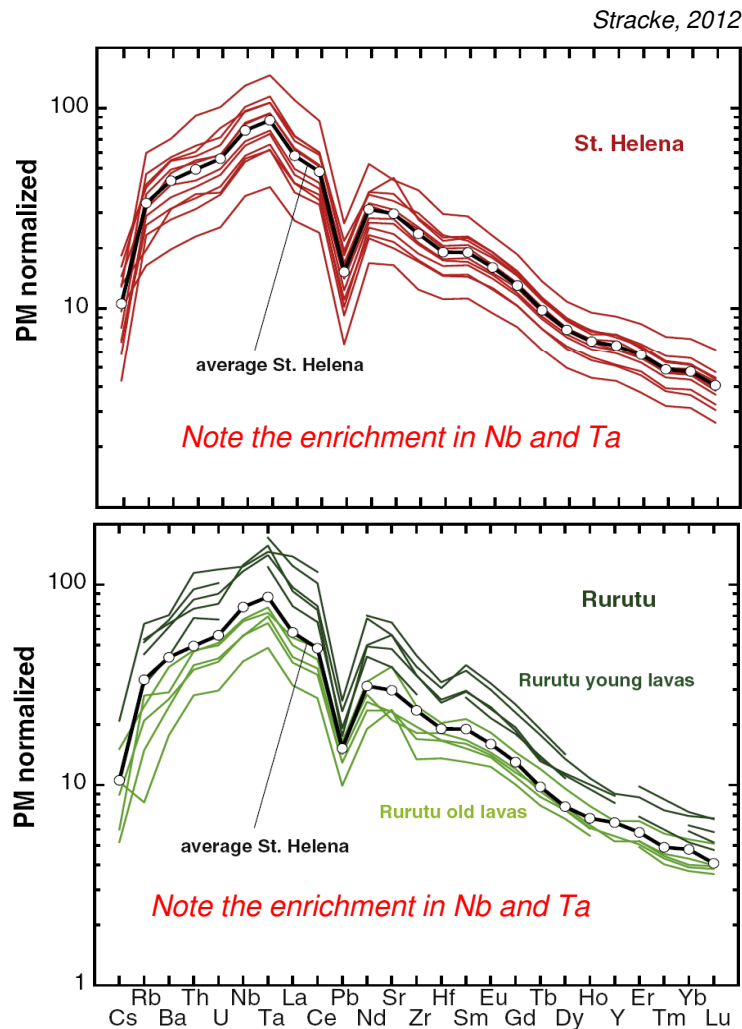
**Averaged PRIMA  
normalized trace-element  
concentrations in OIBs  
from different locations.**

GLOSS is the averaged composition of **globally subducted sediment**, which is very similar to average continental crust.

Note that **Hawaii OIBs** plot most closely to MORB (but have LOWER contents of HREE than MORB)!

# Composition of ocean island basalts

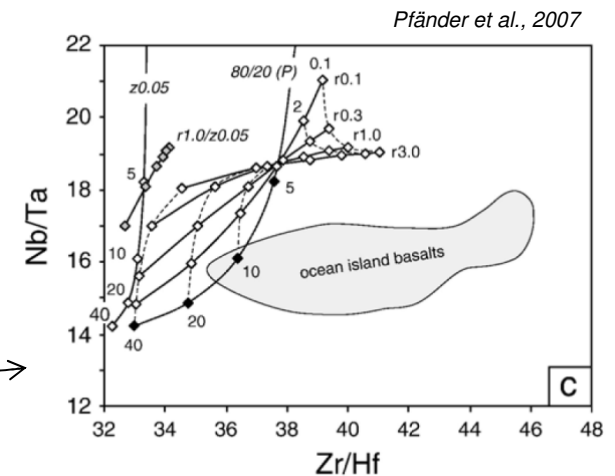
## Trace element features



**HIMU OIBs** are regarded as partial melts derived from a mantle source consisting of **PREMA** + highly altered **oceanic crust/lithosphere** (high U/Pb due to **hydrothermal Pb loss over U**).

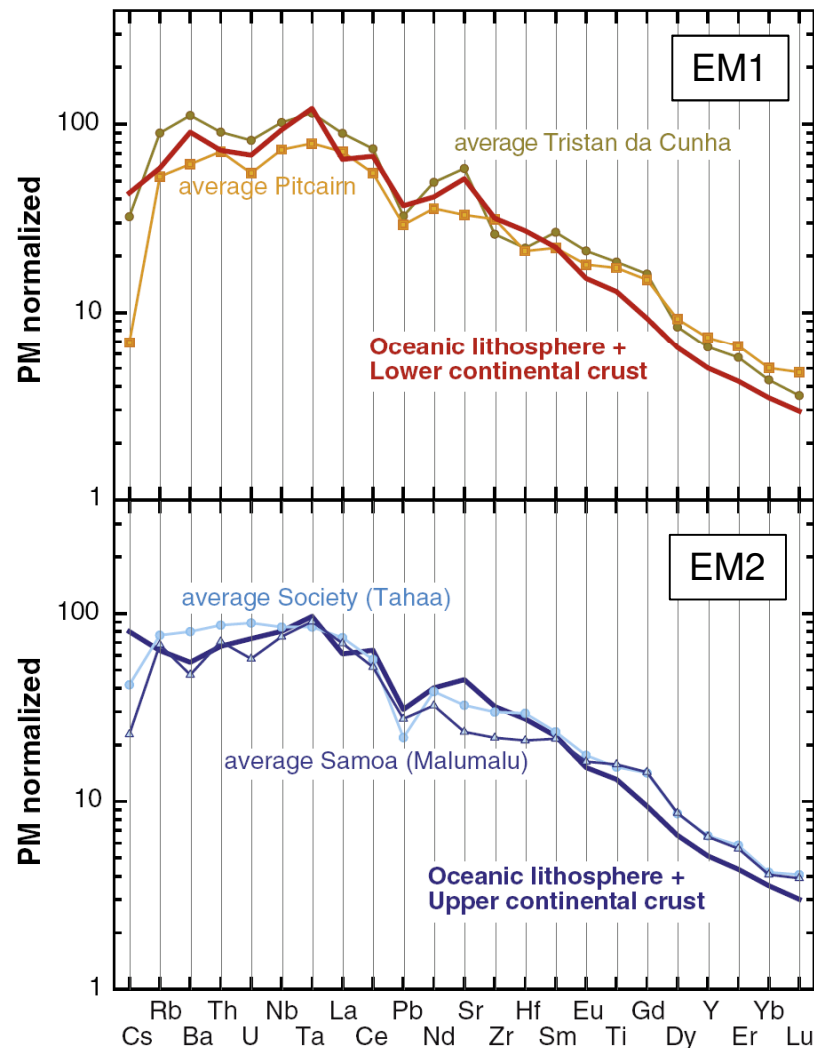
Also note the extreme **enrichment in Nb and Ta** over neighbouring elements (La, Th), which results from melting of Ti-phases out of the subducted slabs (now „eclogite“)!

About **constant Nb/Ta** at **variable Zr/Hf** in OIB suggest up to ~3% **rutile** in the **source of (HIMU) ocean island basalts**, which likely is hosted by **recycled oceanic crust** („eclogite“)

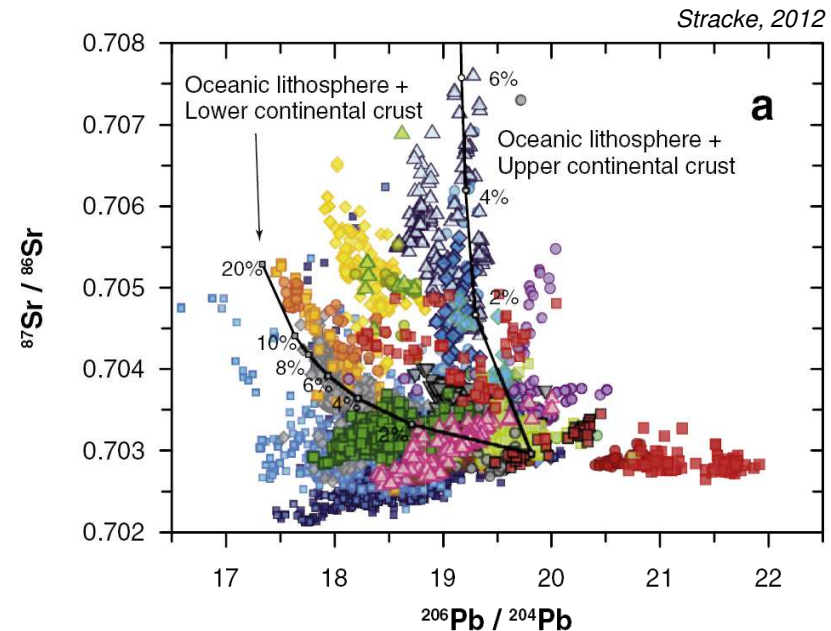


# Composition of ocean island basalts

## Trace element features



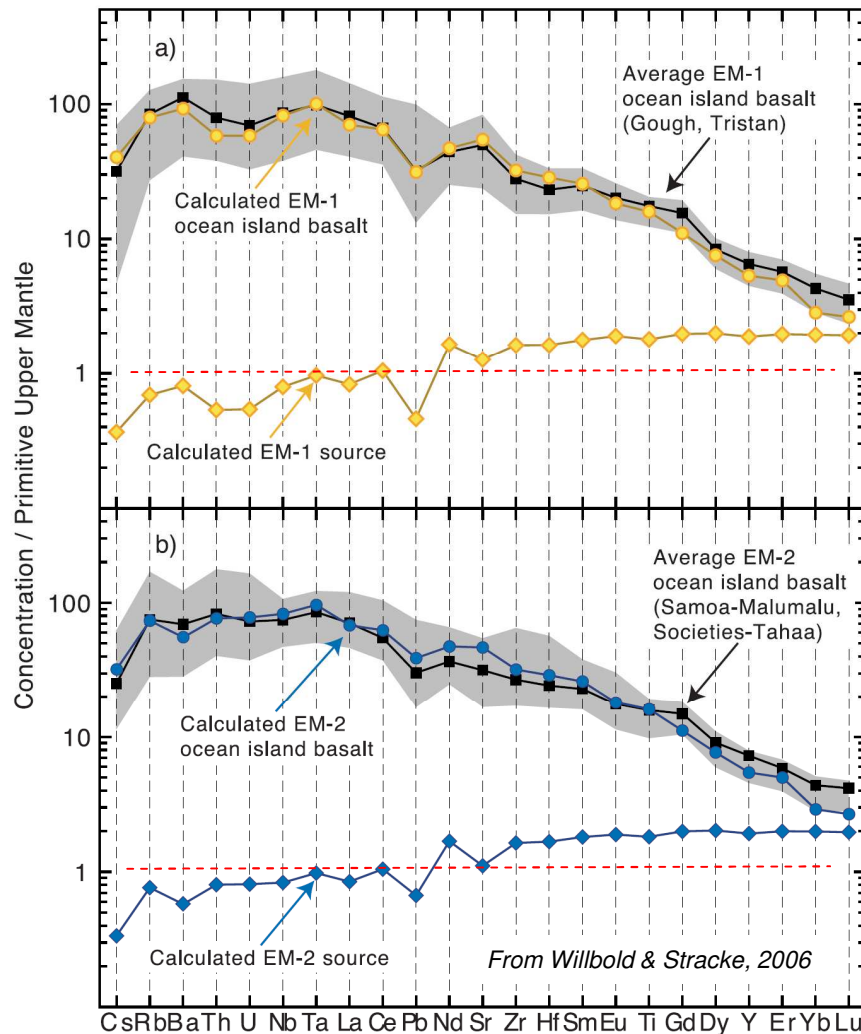
**EM1** and **EM2** compositions are interpreted as reflecting partial melts of a source consisting of **PREMA** + subducted oceanic lithosphere + **lower** (EM1) and **upper** (EM2) continental crust.



Stracke, 2012 (see also Willbold & Stracke, 2006)

# Composition of ocean island basalts

## Trace element features



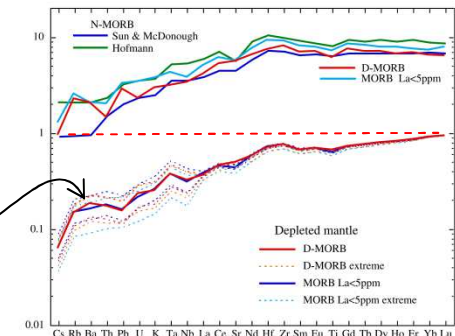
**EM1 and EM2 mantle sources** as inferred from trace-element compositions of EM1 and EM2 ocean island basalts, and applying **non-modal aggregate fractional melting** modelling (**F=1%**, melting in the **garnet** stability field).

**Note that source compositions are similar to PRIMA!**

**EM1 model source: 90% depleted mantle + 9% oceanic crust + 1% lower continental crust**

**EM2 model source: 90% depleted mantle + 9.8% oceanic crust + 0.2% upper continental crust**

*PRIMA (BSE) normalized trace-element abundances of the **depleted mantle source** for comparison (from Salters & Stracke, 2004)*



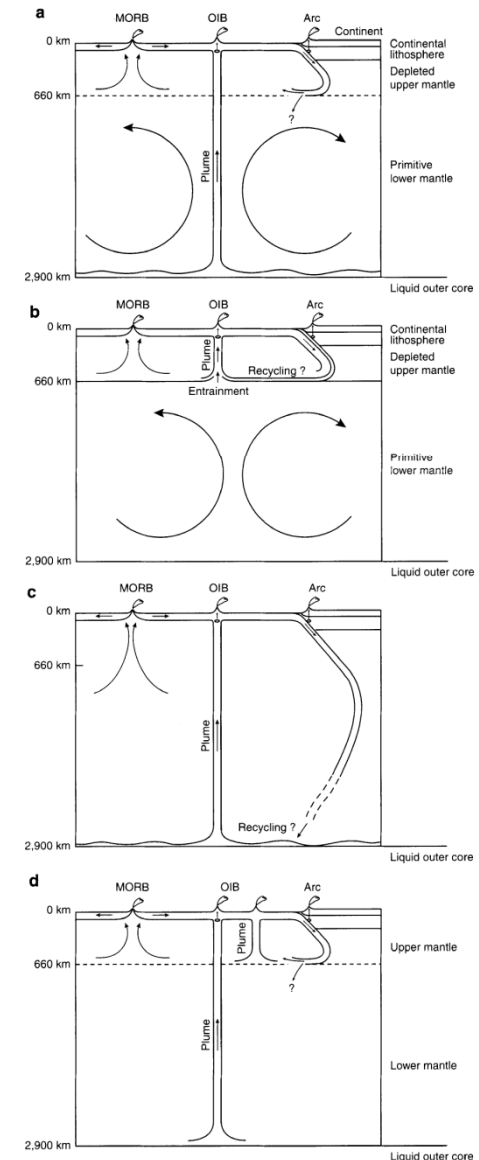
# Composition of ocean island basalts

## Fundamental first order conclusions

- The Earth's mantle is **heterogeneous**
- The MORB and OIB mantle sources have strongly **different geochemical compositions** that maintain **since the geological past** (>1 Ga)
- Unless the MORB source, the source of OIB is **enriched in trace-elements and radiogenic isotopes**
- A combination of variable proportions of primitive and/or depleted mantle **plus** subducted oceanic lithosphere **plus** upper and/or lower continental crust explains most of the observed features

For reviews see Hofmann (1994) *Nature*, and Stracke (2012) *Chemical Geology*

Hofmann, 1994



# Composition of ocean island basalts

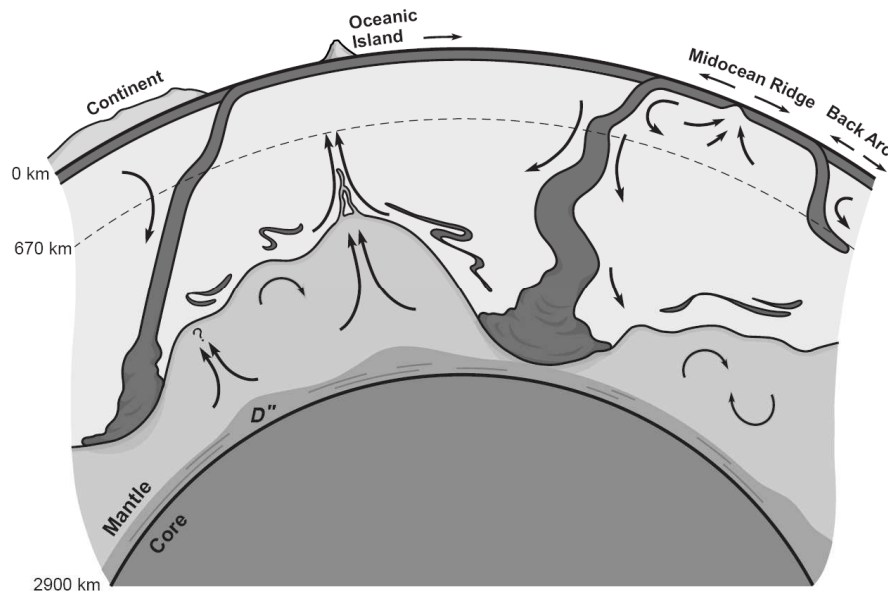
Fundamental first order conclusions

– implications from geophysics

**Geophysical data** (mostly

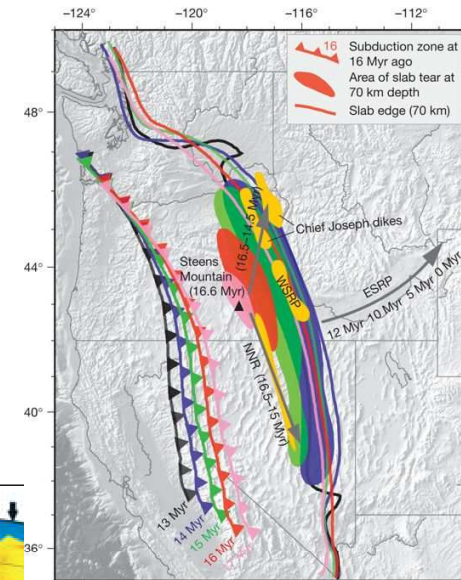
seismic tomography) support

the **recycling hypothesis!**

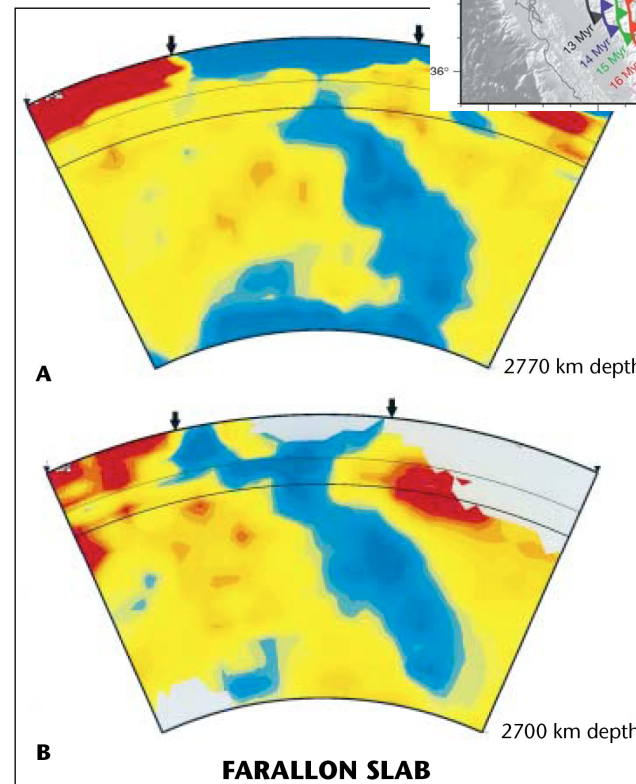


Kellogg et al., 1999, Science

Movement of Farallon Slab (Western USA, from Liu & Stegman, 2012, Nature)



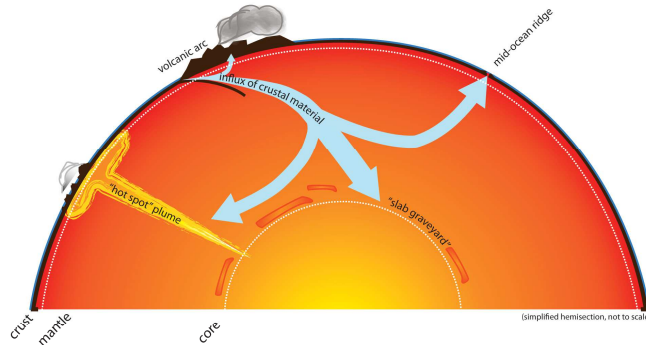
Grand et al., 1997, GSA Today



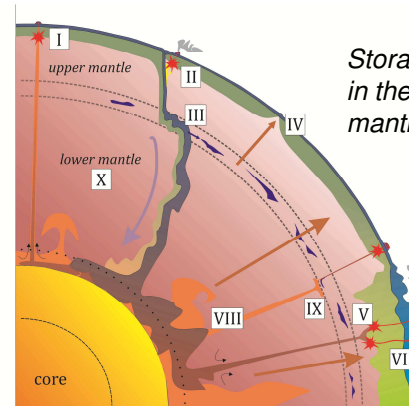
# Composition of ocean island basalts

... and the Cornucopia of mantle models that emerged

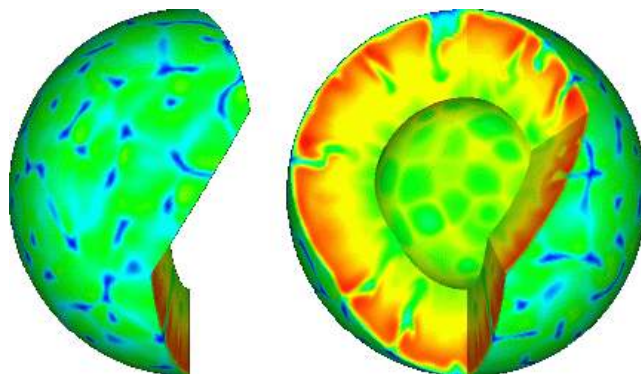
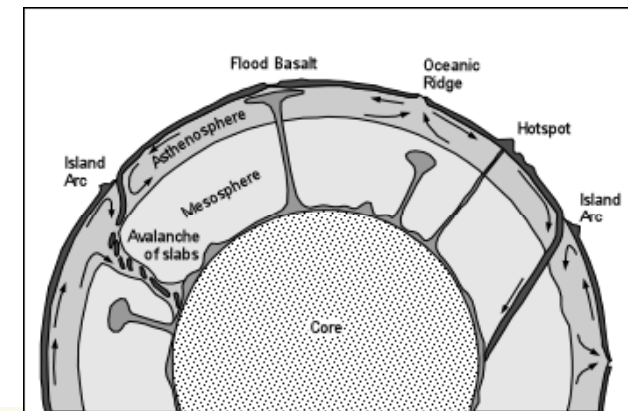
Where does subducted crust go?



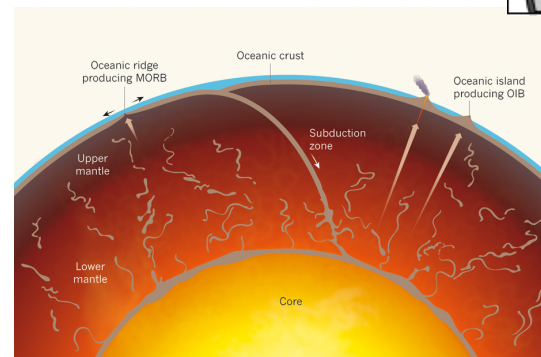
Recycled components in the plume **and** MORB source, and storage in the D'' layer



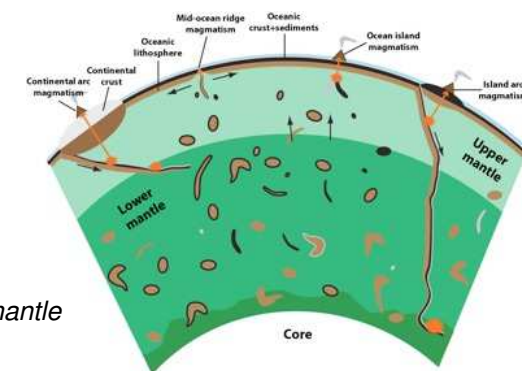
Storage of subducted lithosphere in the D'' layer, i.e. at the core – mantle boundary at ~2900 km



Numeric simulation of mantle convection (from Geophysics Homepage, LMU München, H.-P. Bunge)

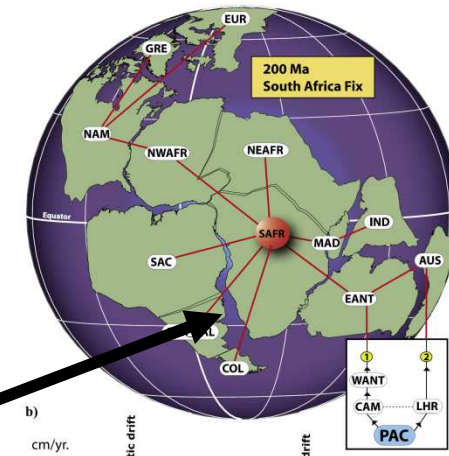
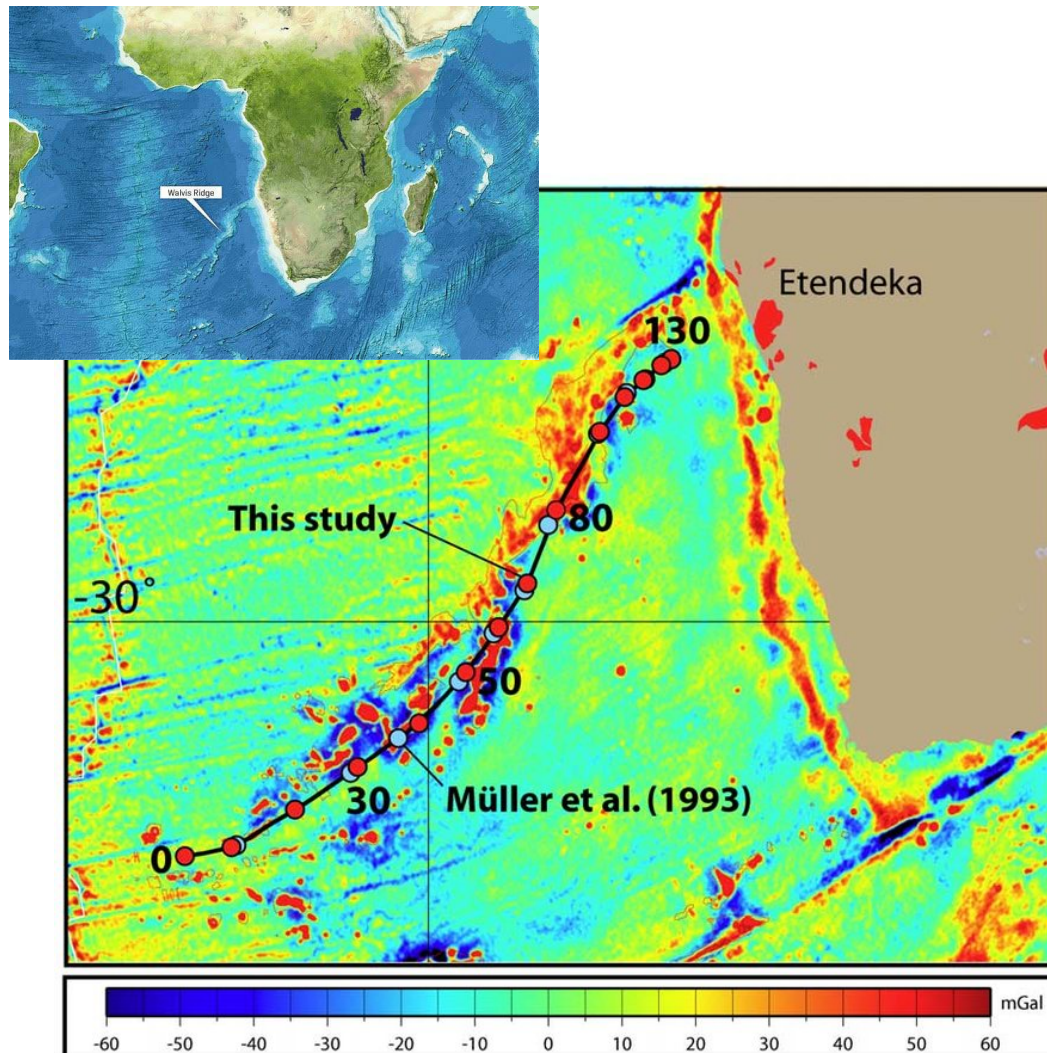


marble cake models of a multiscale heterogeneous mantle



# Mantle plumes and plate tectonics

... do mantle plumes disrupt continents?

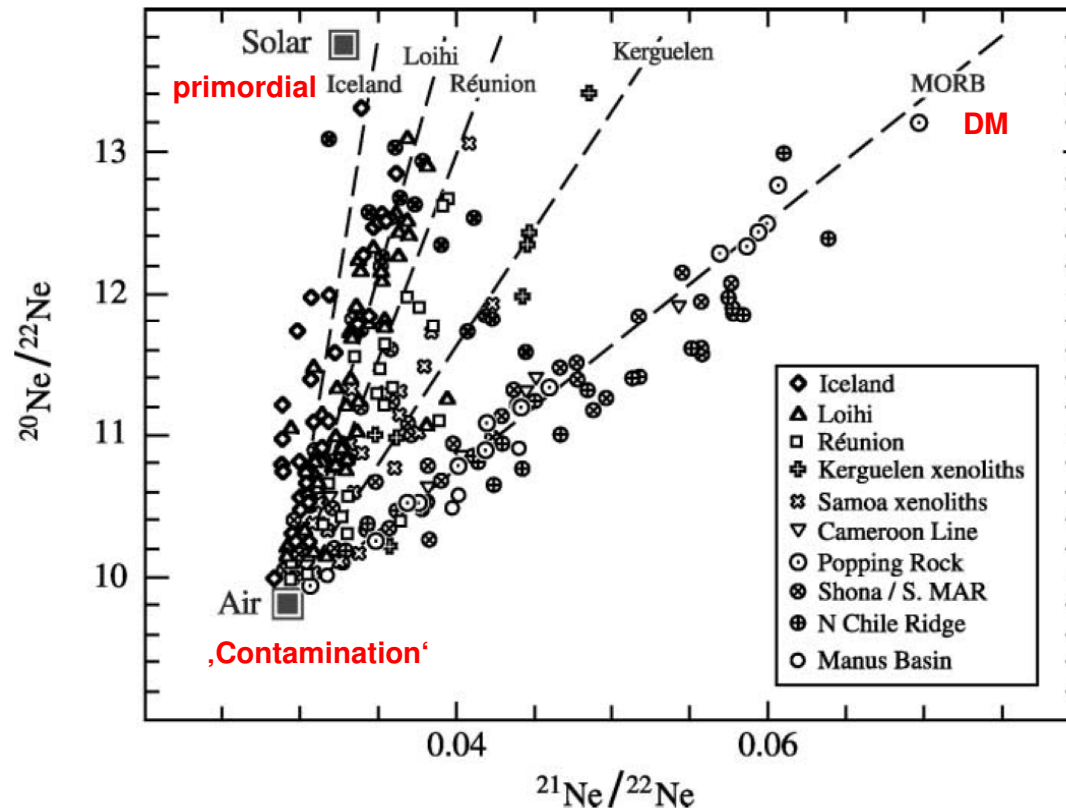


Breakup between **Africa** and **South America** as part of **Pangaea** caused by the **Tristan mantle plume**?

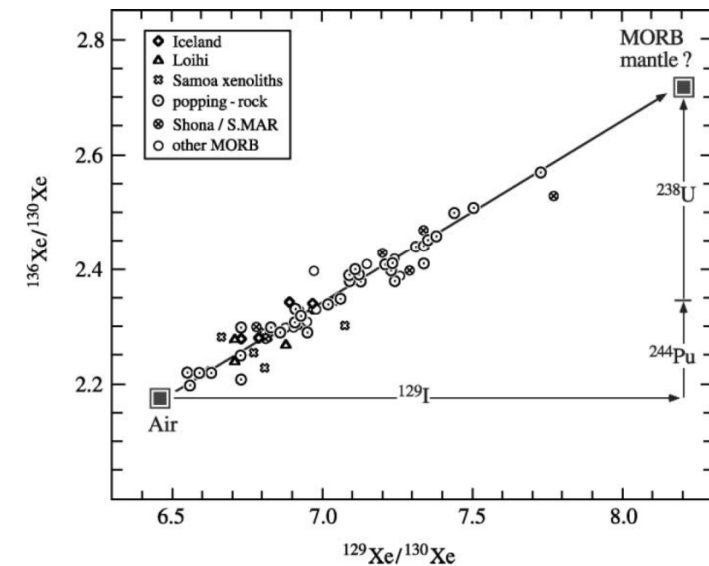
An age progression along the **Walvis ridge** makes this scenario very likely!

# Composition of ocean island basalts

Is there a primordial signature in the OIB source mantle?



Hilton & Porcelli (2003), *Treatise on Geochemistry*

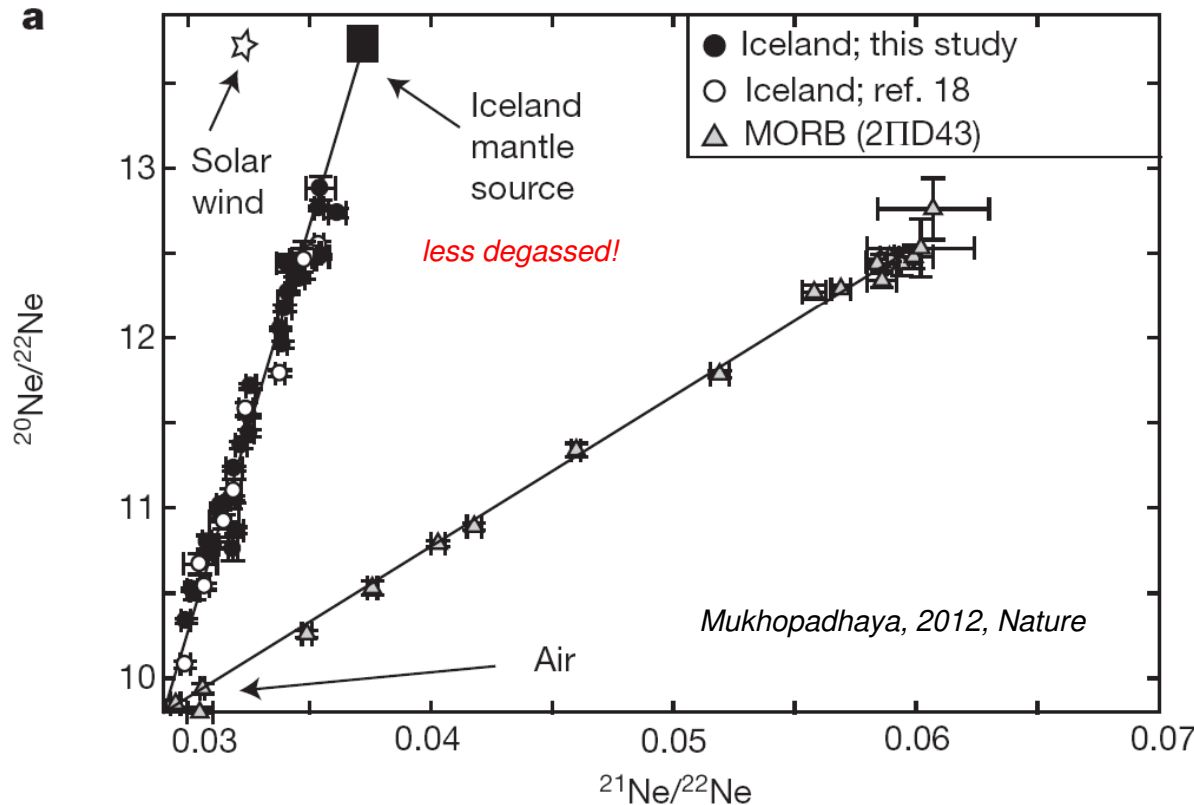


Or, what about the interaction between the OIB and the MORB source mantle? Noble gas constraints....

$^{20}\text{Ne}$  and  $^{22}\text{Ne}$  are primordial  
 $^{24,25}\text{Mg}(n,\alpha)^{21,22}\text{Ne}$   
 $^{18}\text{O}(\alpha,n)^{21}\text{Ne}$   
 $^{19}\text{F}(\alpha,n)^{22}\text{Na}(\beta)^{22}\text{Ne}$

# Composition of ocean island basalts

## Noble gas isotope features



The steeper slope of the **Iceland sample** indicates a higher amount of **primordial  $^{22}\text{Ne}$**  relative to **nucleogenic  $^{21}\text{Ne}$**  in the Iceland mantle source than in the MORB mantle source.

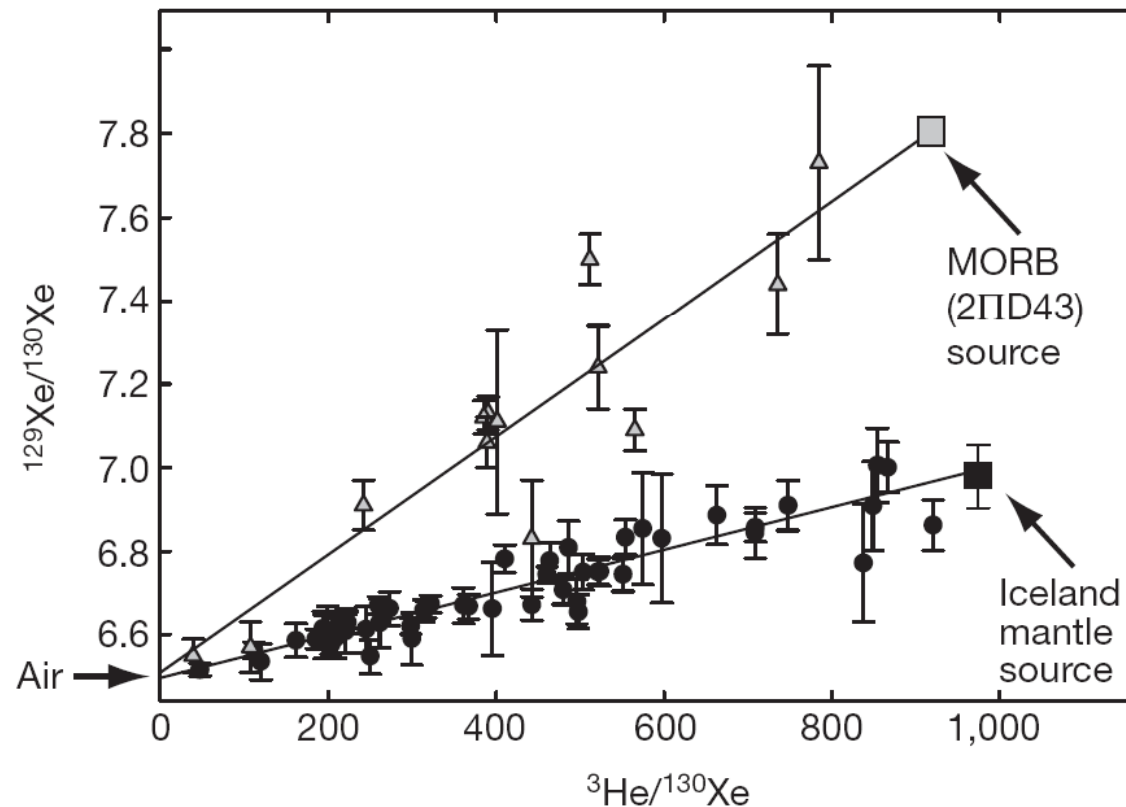
This suggests a **less degassed (deeper?) Iceland mantle source!**

Does this mirror a heterogeneous accretion of the Earth, and subsequent restricted 'mixing' since then?

$^{20}\text{Ne}$  and  $^{22}\text{Ne}$  are primordial  
 $^{24,25}\text{Mg}(n,\alpha)^{21,22}\text{Ne}$   
 $^{18}\text{O}(\alpha,n)^{21}\text{Ne}$   
 $^{19}\text{F}(\alpha,n)^{22}\text{Na}(\beta)^{22}\text{Ne}$

# Composition of ocean island basalts

## Noble gas isotope features



Lower  $^{129}\text{Xe}/^{130}\text{Xe}$  in the **Iceland mantle source** than in the **MORB mantle source** indicates **separation of both sources before ~4.45 Ga**, and only **very limited mixing** since then!

*In other words: It seems that the Earth's mantle sustained a large-scale heterogeneity since Earth formation*

Mukhopadhyay (2012), Nature

$^{129}\text{I}$  ( $\beta$ )  $^{129}\text{Xe}$  ( $T_{1/2} \sim 6 \text{ Ma}$ )

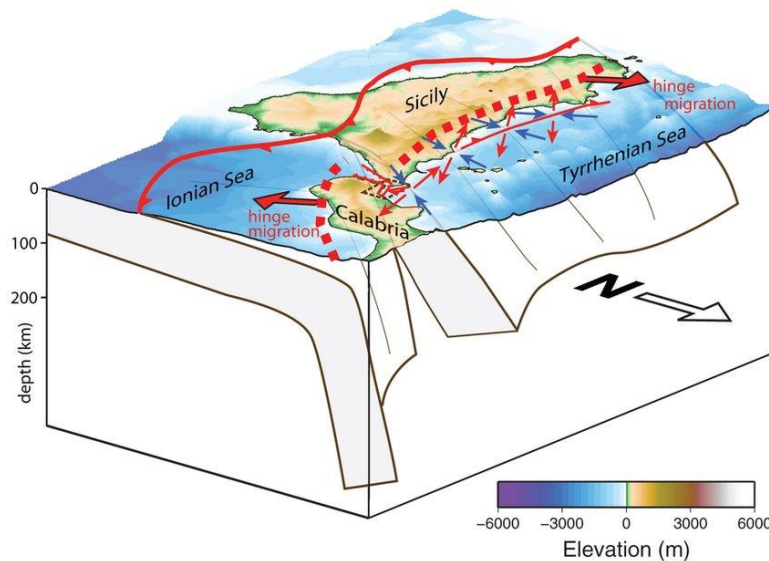
$^{244}\text{Pu}$  ( $^{238}\text{U}$ ) (fission)  $^{136}\text{Xe}$  ( $^{131}\text{Xe}$   $^{132}\text{Xe}$   $^{134}\text{Xe}$ )

Half life of  $^{244}\text{Pu}$ :  $\sim 80 \text{ Ma}$

## Chapter 5

# Continental (anorogenic) magmatism

... or, simply, **continental basalts** in a broader sense ...



Model of magmatism at Mt. Etna (Sicily).  
Magmatism is fed by a **mantle source** tapped  
through a **slab window**, which opened due to slab  
bending and subsequent **slab tear**

Note that the **Eolian Islands** north of Sicily are  
different in that they result from **island arc**  
**magmatism!**

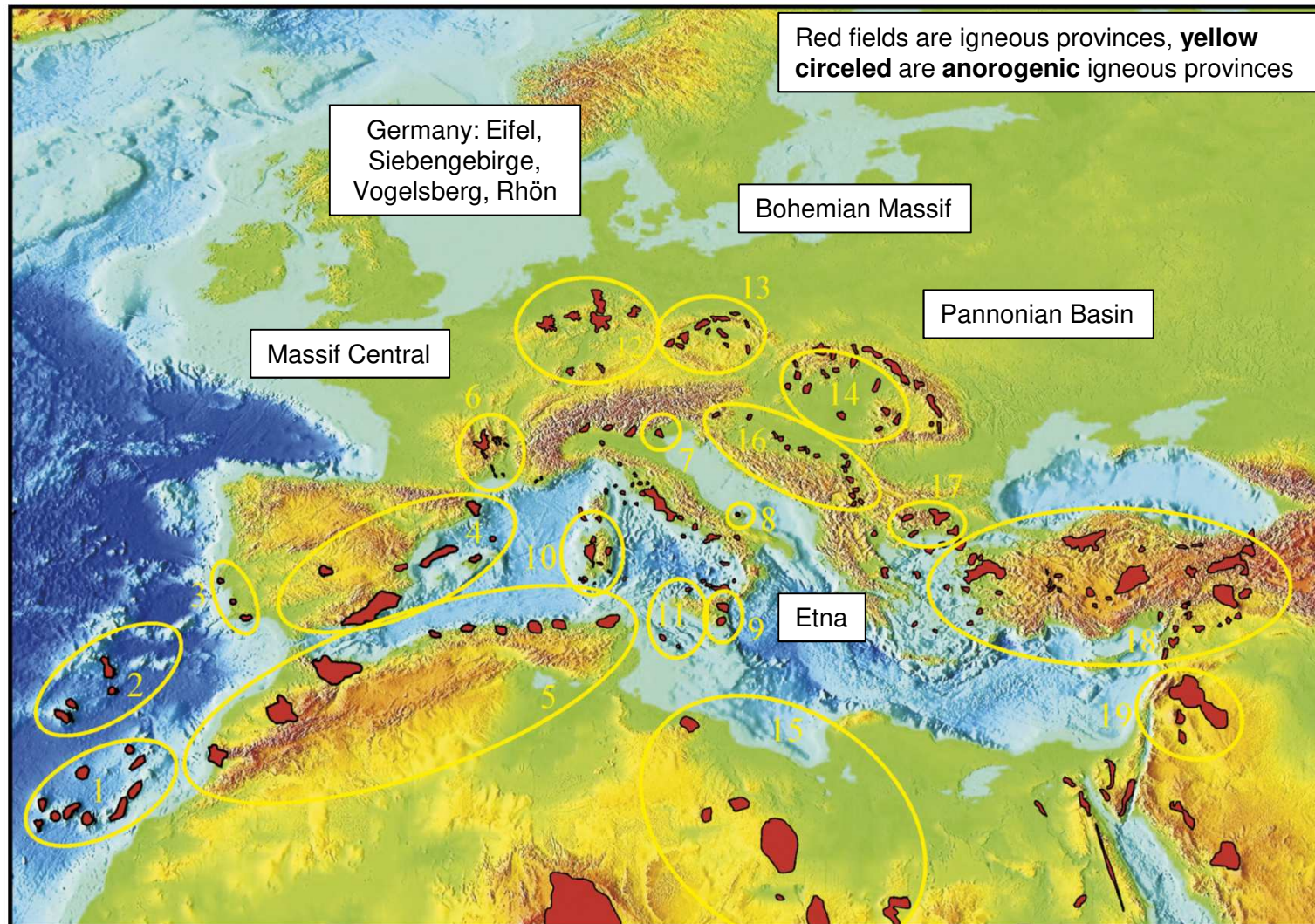
### Key questions

- What is the **geodynamic framework** in which CAM occurs?
- What **rock types** emerge from continental anorogenic magmatism (CAM)?
- What is the **source** of continental (anorogenic) magmatism?  
Is it the **asthenospheric** or the **lithospheric** mantle, or both?
- What is the **degree of melting**, what is the primary magma composition?
- What is the role of **crustal assimilation** and **fractional crystallisation**?

# Circum-Mediterranean (anorogenic) igneous provinces

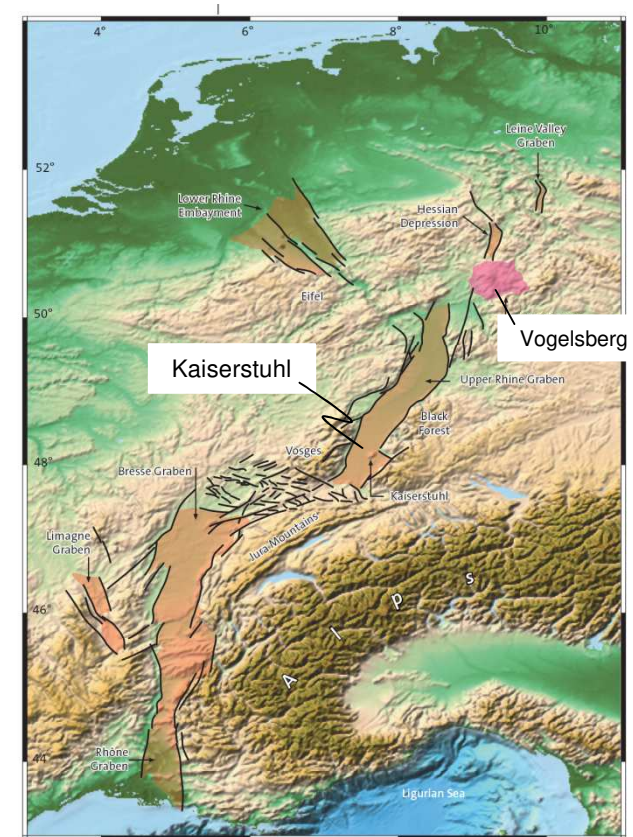
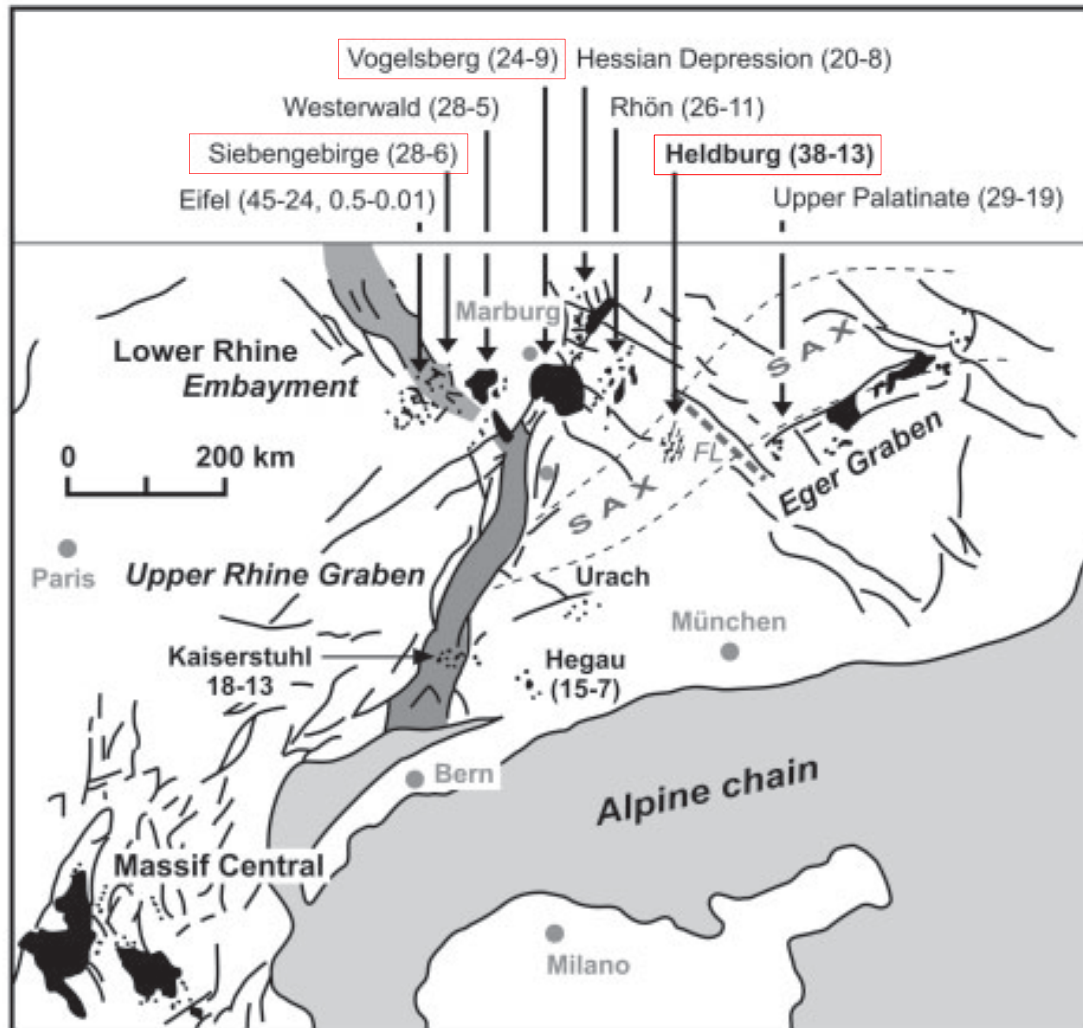
Cenozoic volcanic provinces – tectonic setting

*Lustrino & Wilson, 2007*



# Tectonic setting of CAM

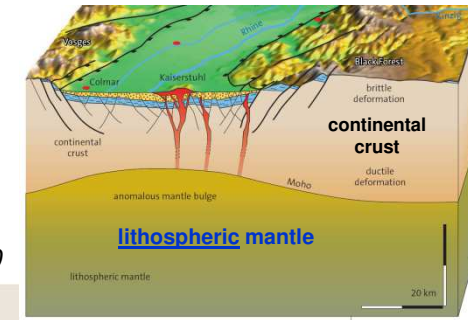
Here: Central European Volcanic Provinces – overview map



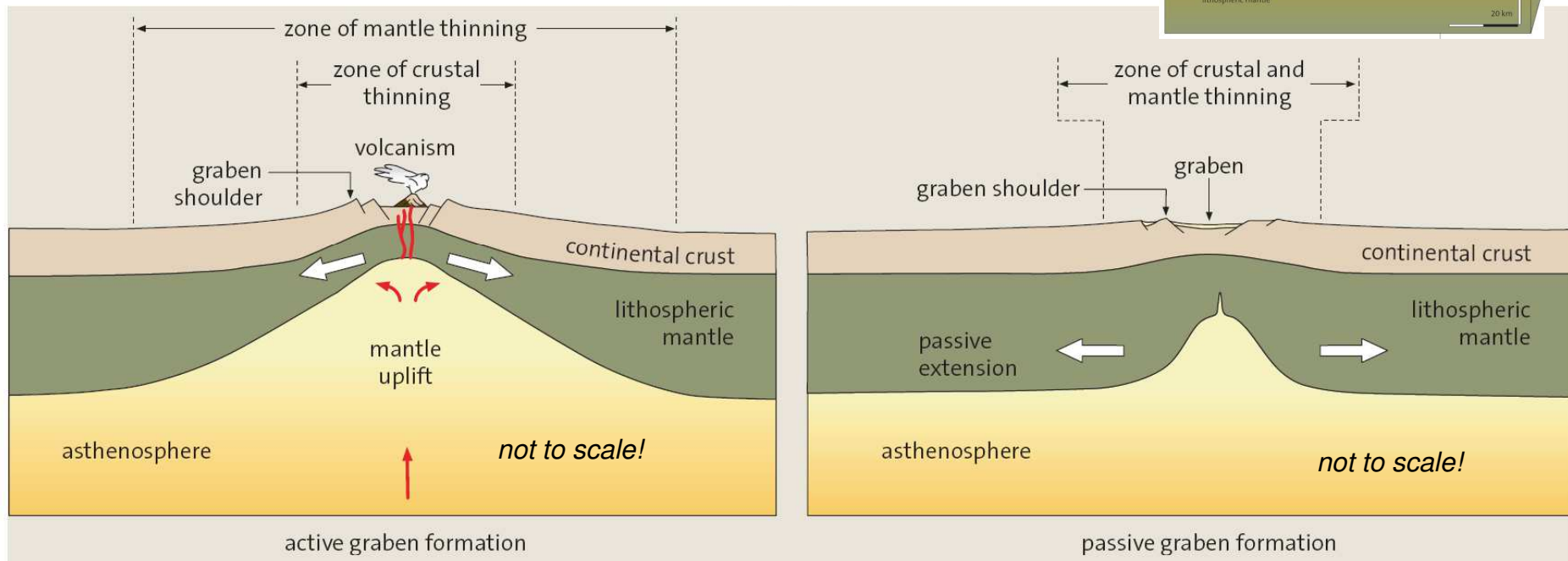
Lower Rhine graben, upper Rhine graben, Hessian depression, Vogelsberg, Bresse graben (from Frisch et al., 2010)

## Tectonic setting of CAM

Magma origin – working hypothesis



From Frisch et al., 2010



Rifting caused by mantle upwelling

Rifting caused by lithosphere extension

So, what causes melting of what, and why? Remember:



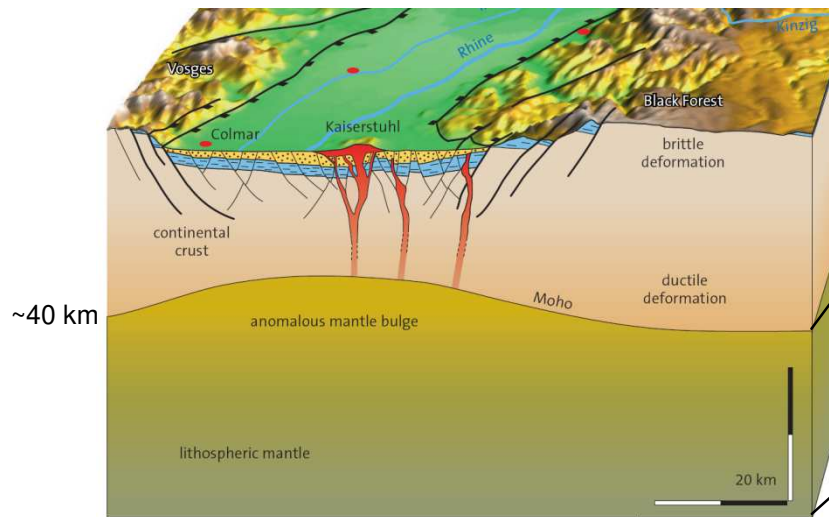
Melting the mantle?  
Temperature is, what matters!

$$\left(\frac{\partial T}{\partial z}\right)_S = \frac{g\alpha_f T}{C_P}$$

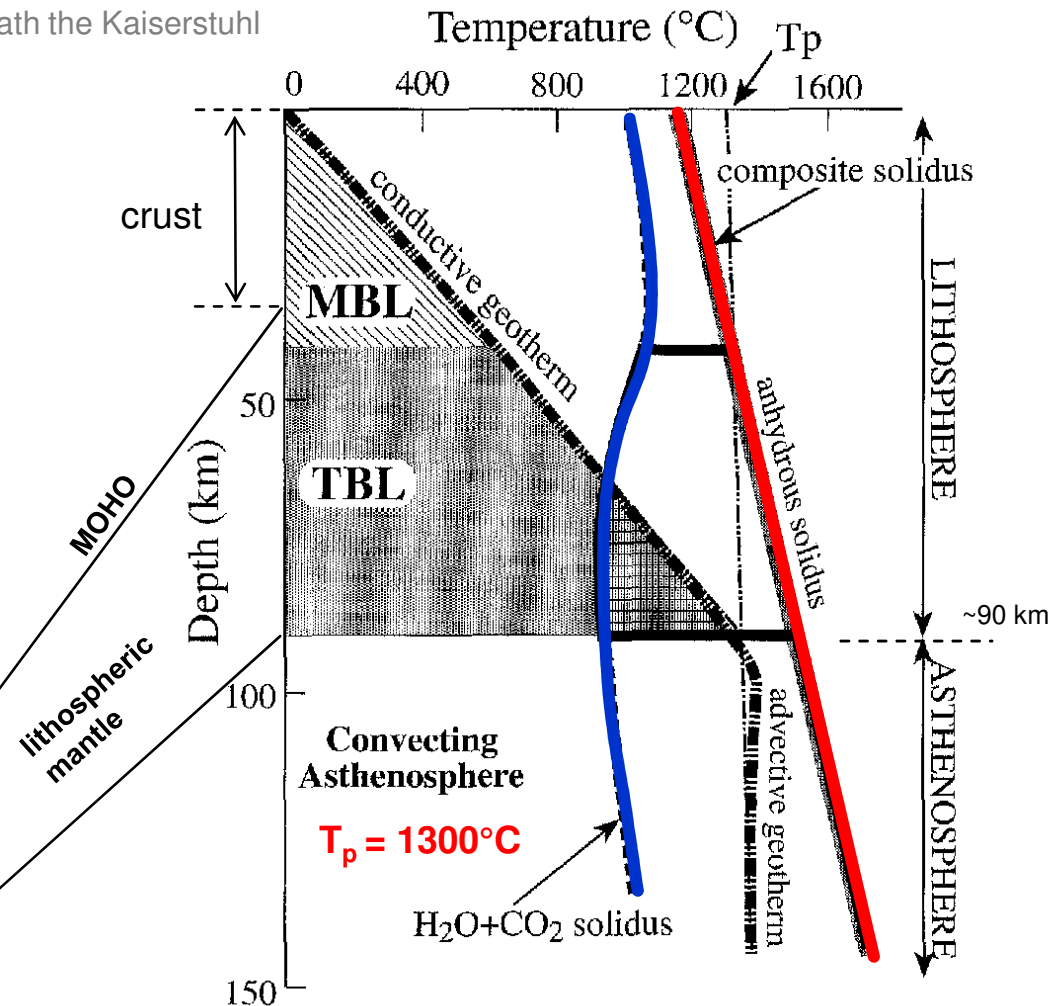
## Tectonic setting of CAM

Temperature distribution in the lithospheric and asthenospheric mantle beneath the Kaiserstuhl

MBL = Mechanical boundary layer  
 TBL = Thermal boundary layer  
 Conductive geotherm calculated based on  $T_p = 1300^\circ\text{C}$



From Frisch et al., 2010

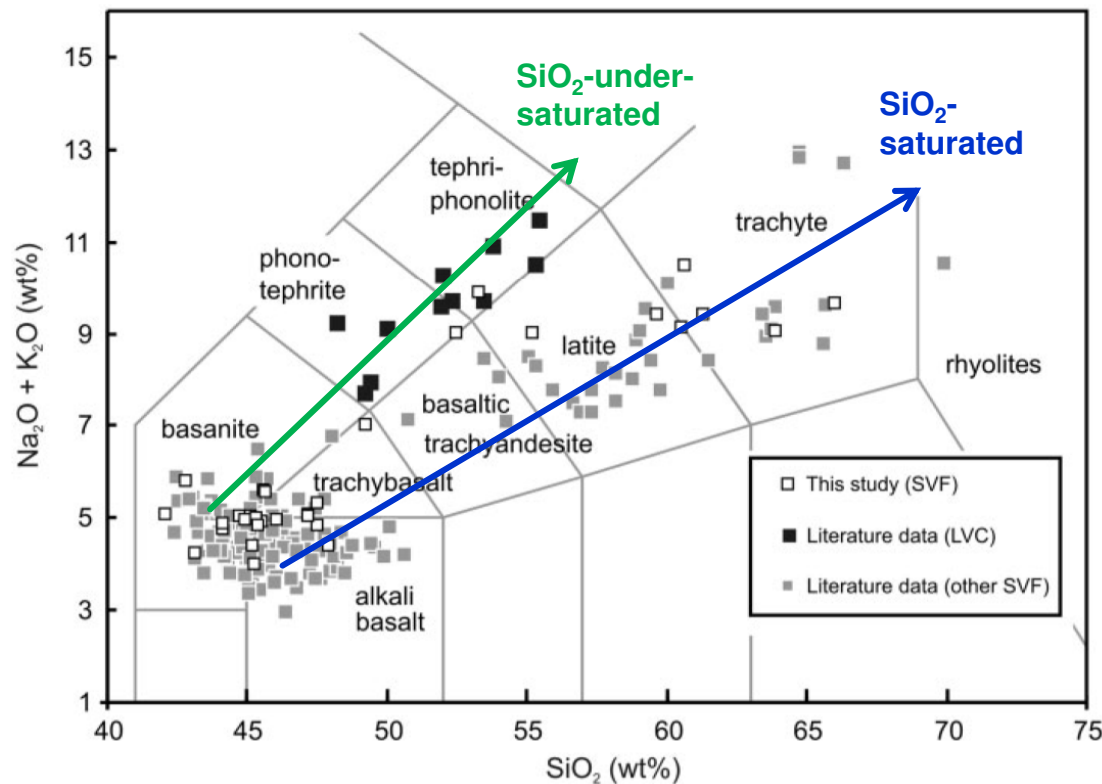


Wilson et al. (1995)

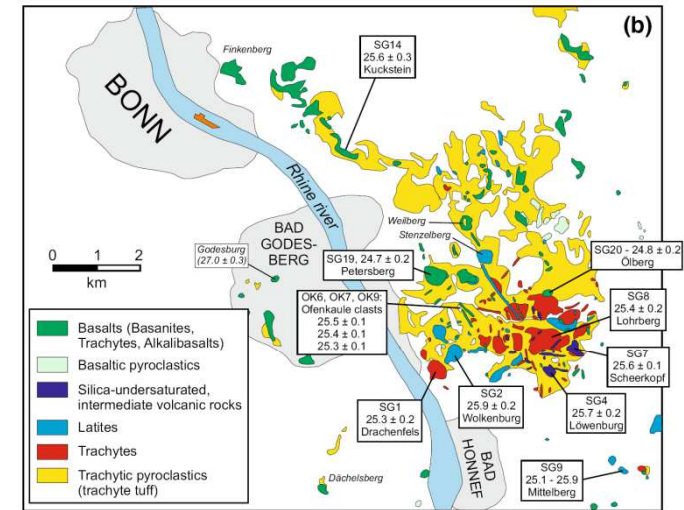
## Rock types in continental magmatic settings

Siebengebirge volcanic province (Central Germany)

*Przybyla et al., 2017*



*Kolb et al., 2012; Jung et al., 2012*

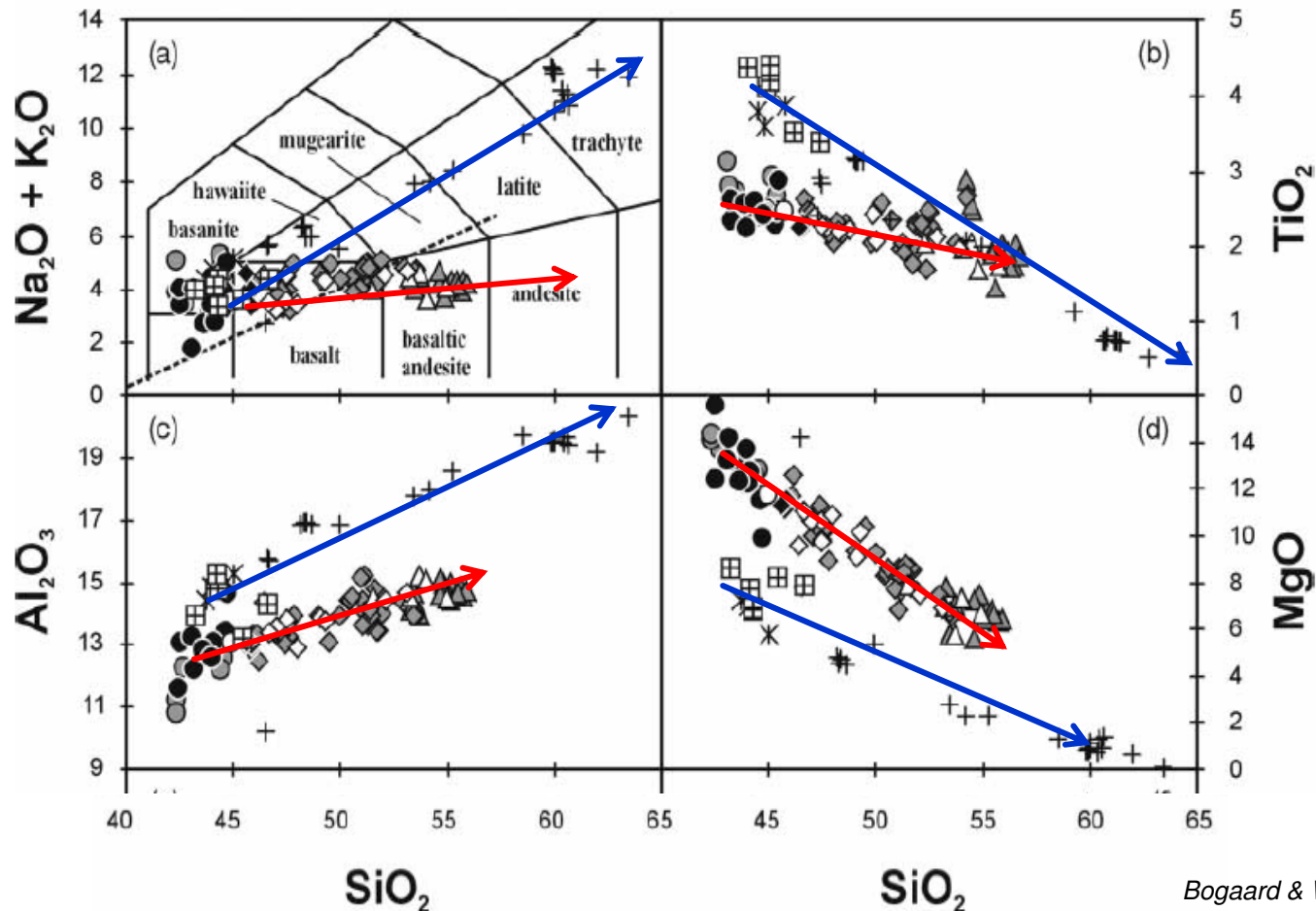


*Siebengebirge volcanic province in Central Germany near Bonn (Rhein) (From Przybyla et al., 2017)*

Anorogenic continental magmatism comprises both, SiO<sub>2</sub>-undersaturated and SiO<sub>2</sub>-saturated alkaline series

# Rock types in continental magmatic settings

Vogelsberg volcanic province (Central Germany)

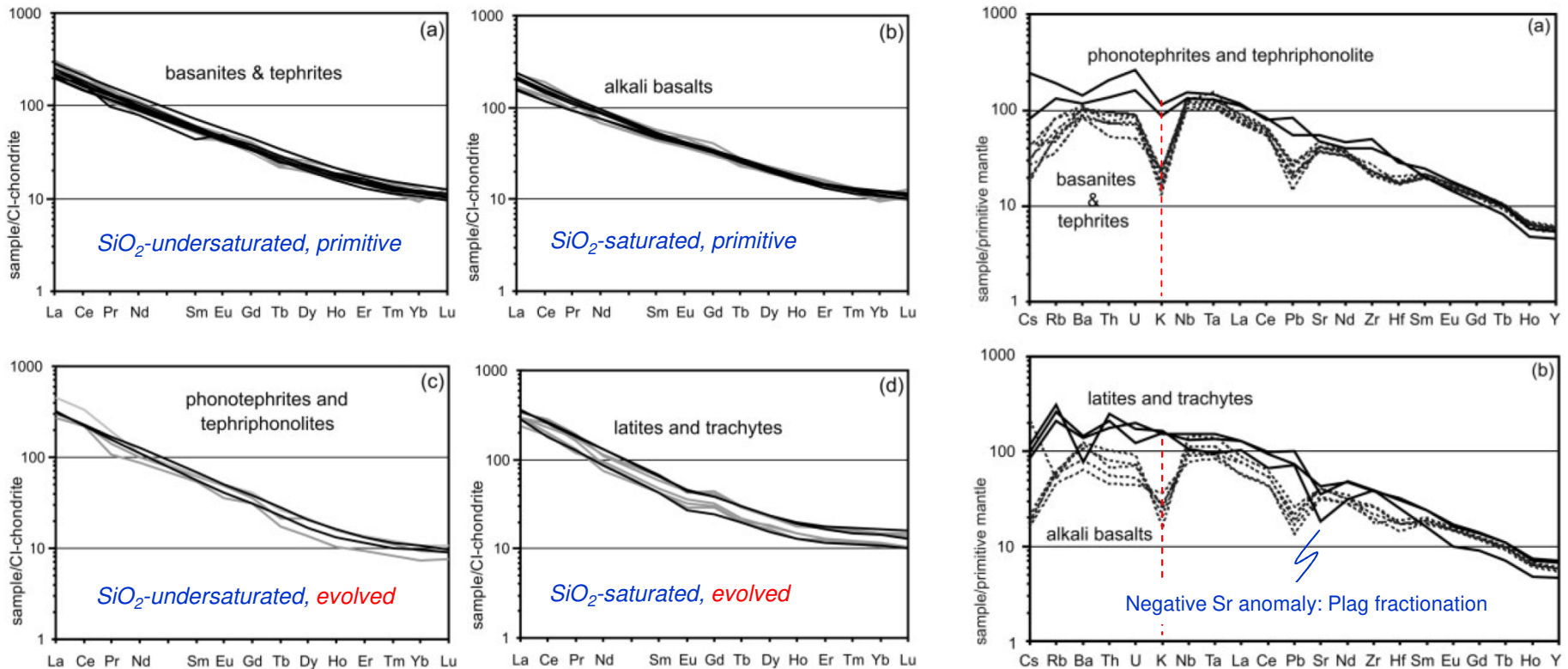


In the **Vogelsberg** volcanic province, magmatism erupted both, **alkaline series** and **tholeiite series** rocks

## Rock types in continental magmatic settings

Trace element and isotope characteristics

Kolb et al., 2012

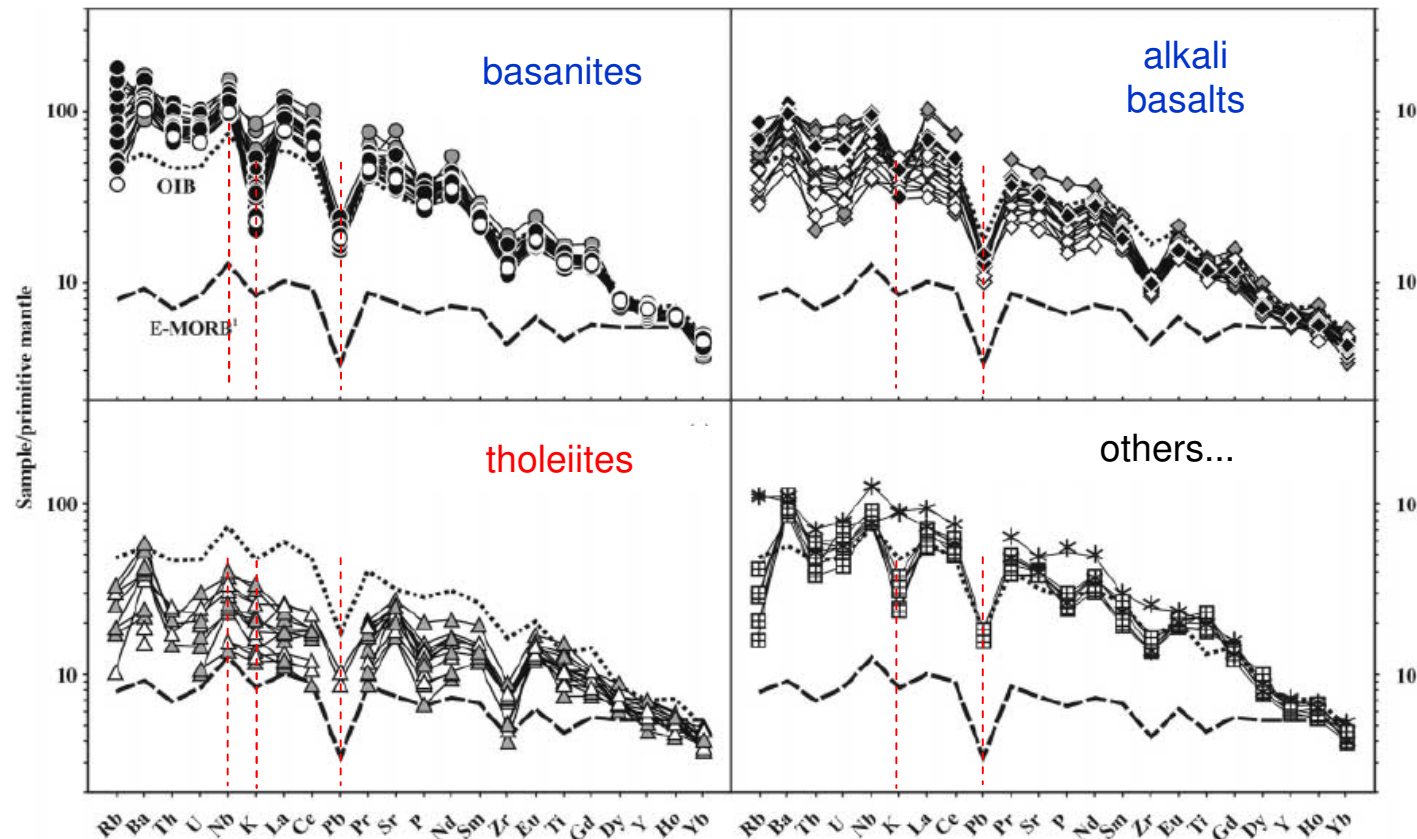


Chemical characteristics of **volcanic rocks** from the **Siebengebirge**. Trace element enrichment, **Nb/La ~1 (or >1)** as well as **negative (Cs, Rb), K and Pb anomalies** in the basanites/tephrites and alkali basalts are obvious. Differentiated serie have higher TE amounts.

## Rock types in continental magmatic settings

Trace element and isotope characteristics

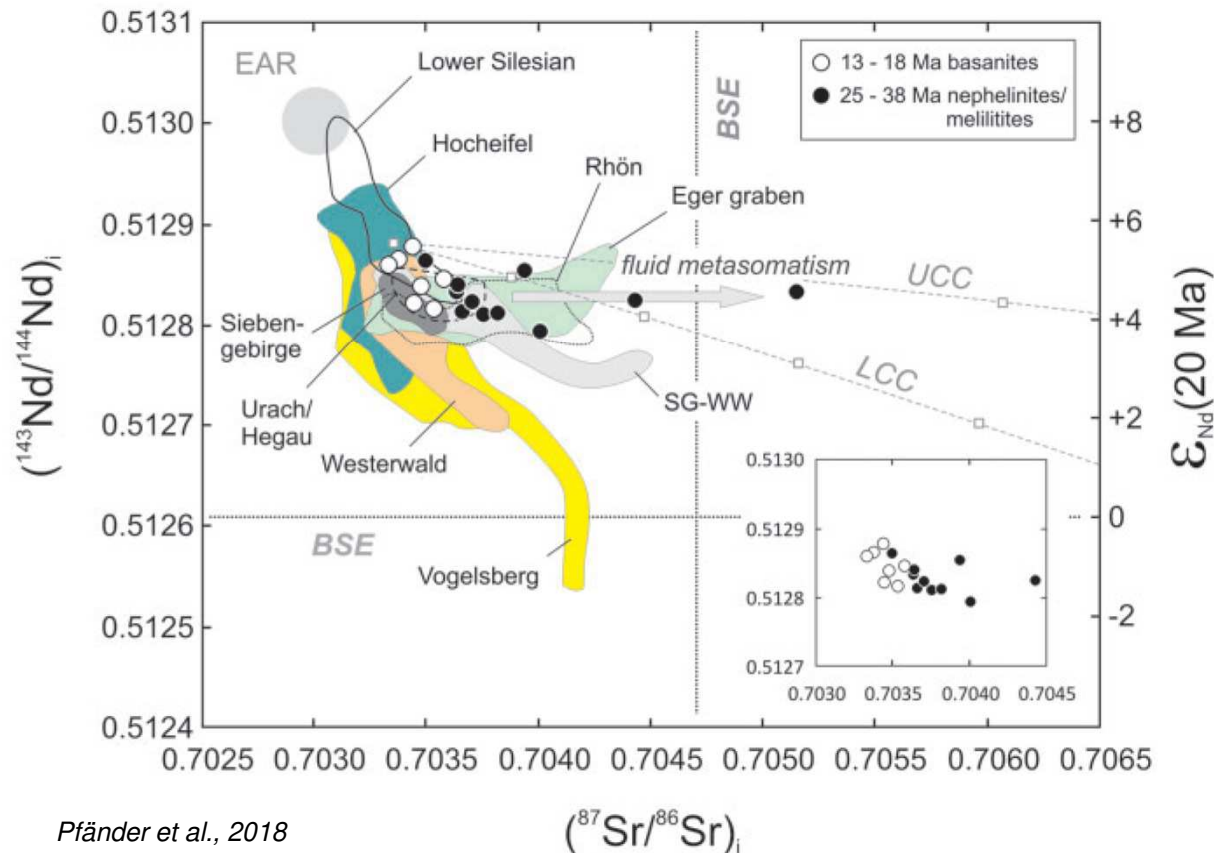
*Bogaard & Wörner, 2003*



Chemical characteristics of **volcanic rocks** from the Vogelsberg. Note the negative K- and Pb-anomalies in the **alkaline rocks** (that strongly resemble OIBs), but the **absence of a negative K-anomaly in the tholeiites**

# Rock types in continental magmatic settings

Trace element and isotope characteristics



Pfänder et al., 2018

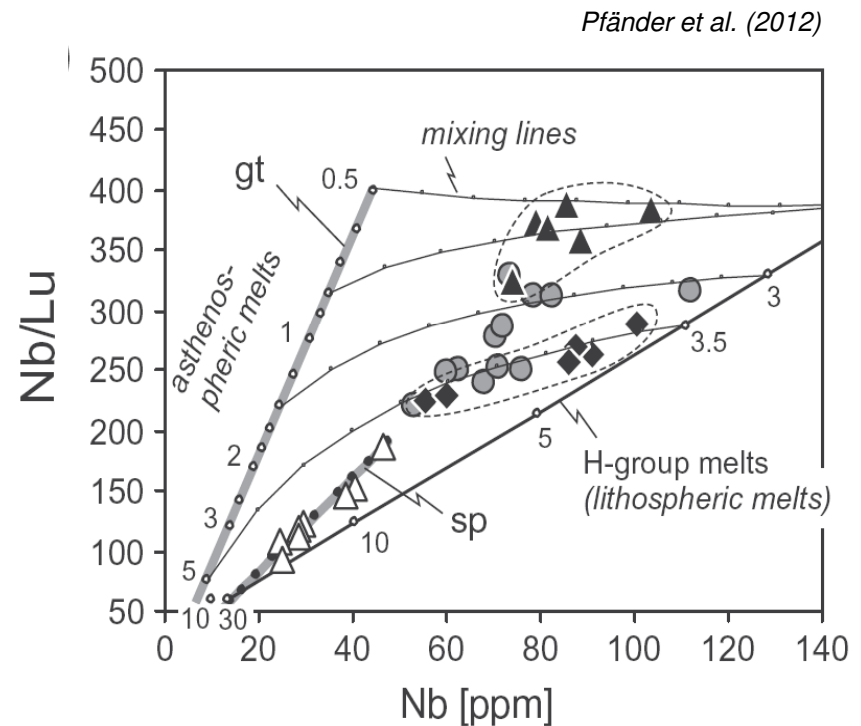
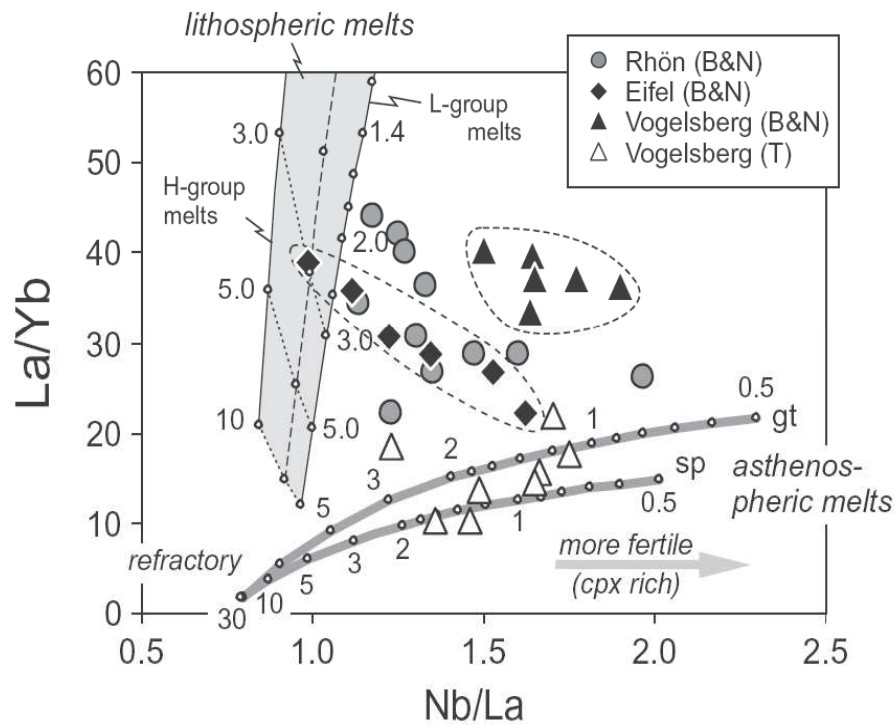
Nd-Sr isotope characteristics of **intraplate basalts** from Central to East Germany and the Silesian and Eger volcanic provinces.

Dots are **basanites and nephelinites** from the Heldburg region (N-Bavaria and S-Thuringia).

Note that the **isotope compositions** resemble – very broadly – the **PREMA composition of OIBs**: *EAR = European Asthenospheric Reservoir*

## Mantle sources involved in CAM

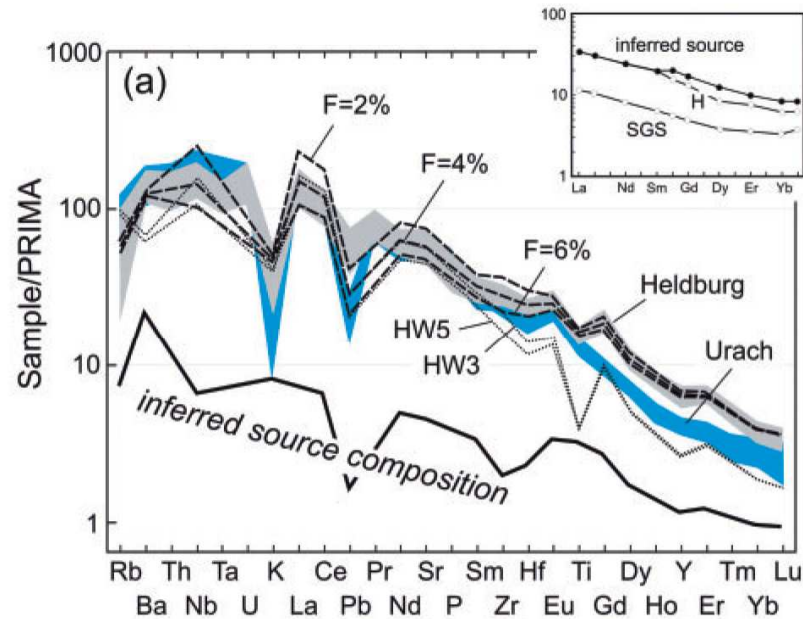
Lithospheric vs. asthenospheric mantle source



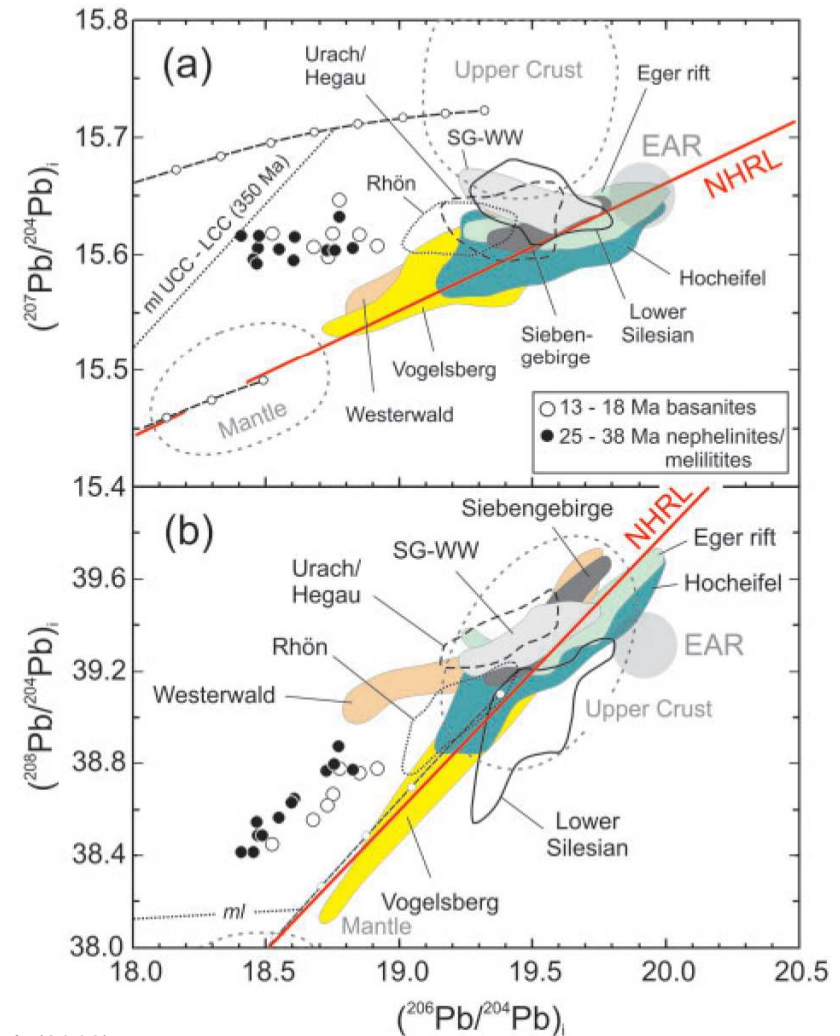
Trace-element modelling applied to **primitive mafic rocks** from the Rhön, Eifel and Vogelsberg magmatic provinces indicate the involvement of **at least three mantle source components** (primitive to slightly-depleted asthenospheric garnet- and spinel-peridotite mantle components as well as enriched spinel-peridotite lithospheric mantle)

# Evolution of CAM

Regional examples and conclusions: Heldburg region

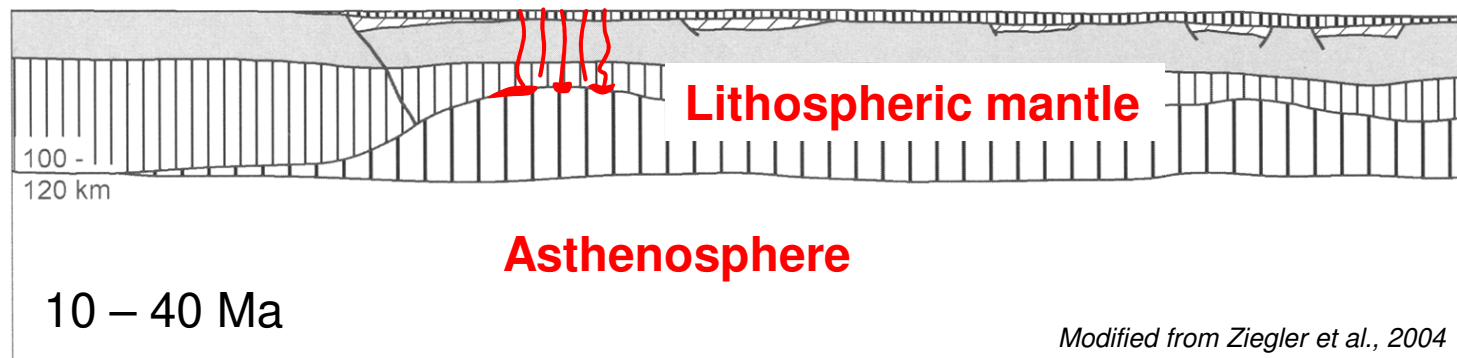
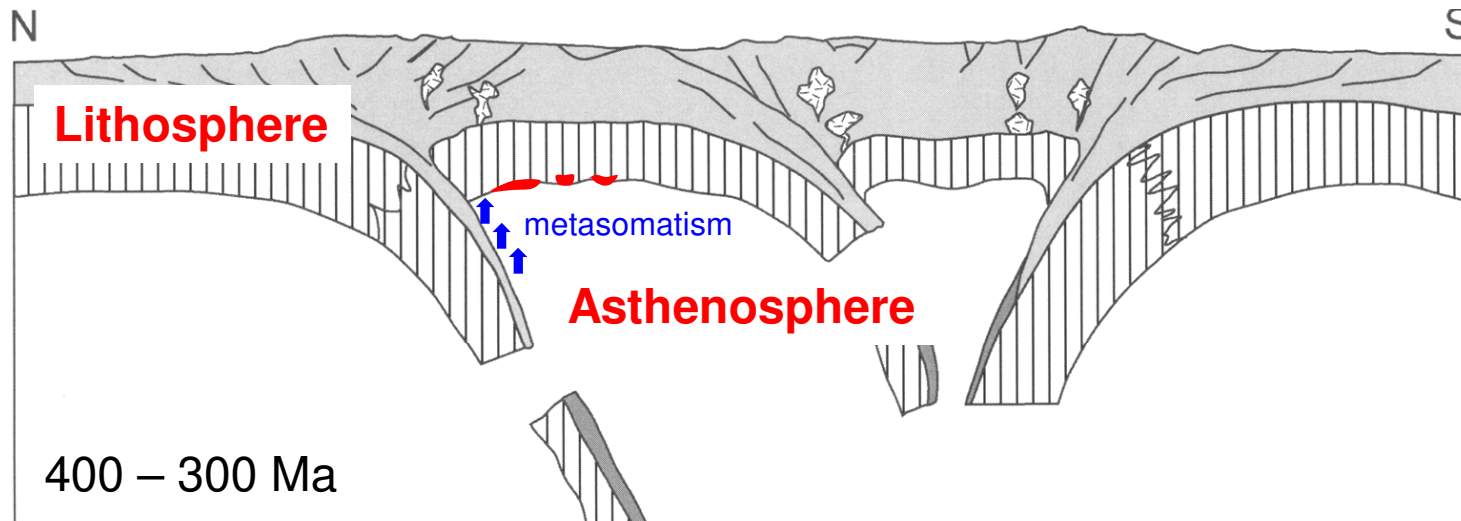


The HELDBURG region seems to be a rare example of nearly pure (low-degree) melting of a strongly metasomatised lithospheric mantle



# Evolution of CAM

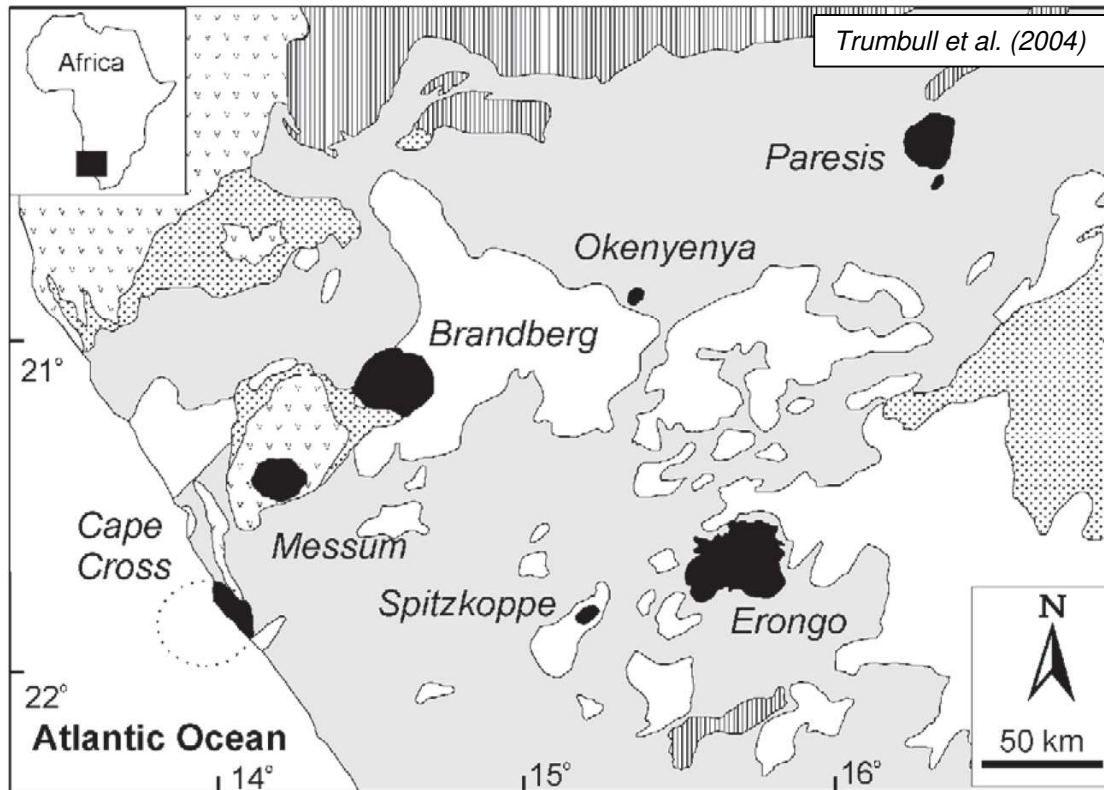
Regional examples and conclusions: Heldburg region









Modified from Ziegler et al., 2004

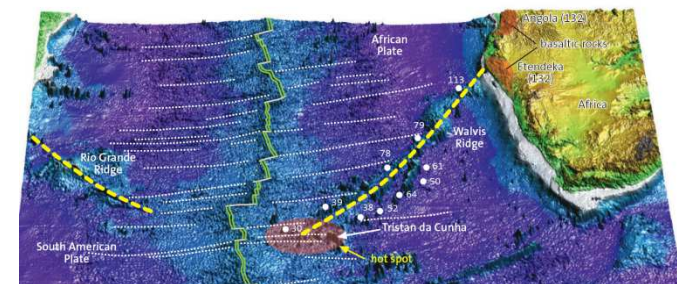
## Evolution of CAM – more regional examples

Messum, Brandberg, Erongo, Paresis alkaline complexes



- |   |  |
|---|--|
|  Damaraland complexes    |  Damara granites      |
|  Etendeka volcanic rocks |  Damara metasediments |
|  Karoo sediments         |  Pre-Damara basement  |

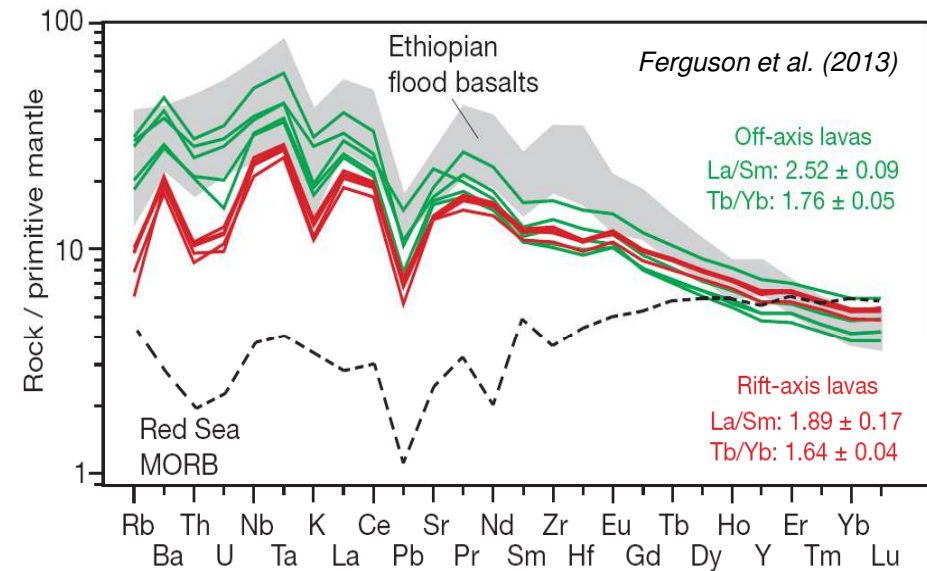
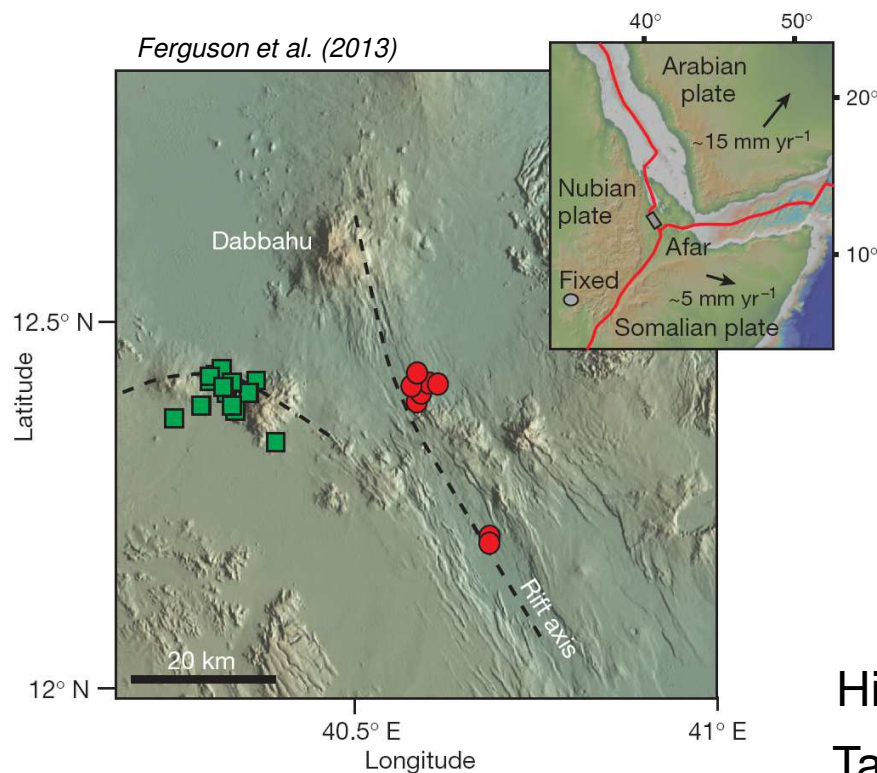
These **early Cretaceous** Complexes (137-125 Ma) comprise a broad variety of **volcanic to plutonic rocks** (basanites, tephrites, alkali basalts, gabbros, Qz- and Neph-syenites, alkali-granites, etc.)



Tristan da Cunha mantle plume and 'hot spot track' (Walvis ridge) SW of Namibia (from Frisch et al., 2010)

# Hot spot induced intraplate magmatism

Afar Hot Spot – Afar Triple Junction – Ethiopian Rift



High contents of LILE, but in particular Nb and Ta indicate the plume contribution in the EFBs. Assimilation of continental crust would reduce the Pb anomaly and would lower Nb and Ta

### ,Exotic' rock types in continental magmatic

**settings:** Carbonatites, Kimberlites, Lamproites, Lamprophyres

#### Carbonatites

Rocks from mantle derived ultra-low degree melts consisting of >50 mod.% primary Ca, Mg, Na, (K) – carbonate

Mantle source is CO<sub>2</sub>-rich (metasomatism)



#### Lamprophyres

Ultramafic (often diamond bearing) rocks, intermediate between Carbonatite and Lamproite



#### Lamproites

Ultramafic silicic rocks derived from a H<sub>2</sub>O-rich (but mostly CO<sub>2</sub>-free) mantle source.

Mantle source possibly contains some CH<sub>4</sub>

#### Kimberlites

Ultramafic alkaline, often diamond bearing rocks from the deepest parts of the cratonic lithosphere

Such **highly SiO<sub>2</sub>-undersaturated** rocks typically represent **low- to ultra-low degree melts** from the interface between the **lowest parts** of the metasomatised **continental lithospheric mantle** and the **convecting asthenosphere**

## ‘Exotic’ rock types in continental magmatic settings: Lamprophyres from Beaver Lake, Antarctica

Table 1. Modal mineral abundances in the Beaver Lake ultramafic lamprophyre facies

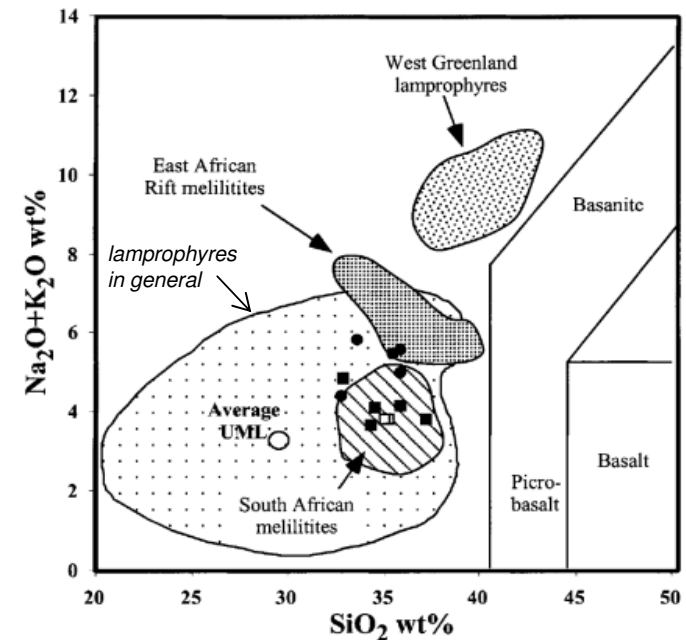
Intrusion type	Ol	Ne	Mel	Phl	Ti-Mag	Prv	Ap	Carb	Cpx	Spl	Glass	Mtc
Dyke	10–15	7–10	30–40	7–10	7–10	3–5	<1	5–10	–	2–3	3–5	–
Plug	7–10	10–15	20–30	15–20	7–10	7–10	<1	3–5	+**	+	10–15	–
Sill	5–7	20–25	15–20	30–40	10–15	10–15	2–3	3–5*	–	<1	2–3	+

\* Carbonate is mainly secondary calcite; \*\* clinopyroxene occurs only within leucocratic globules

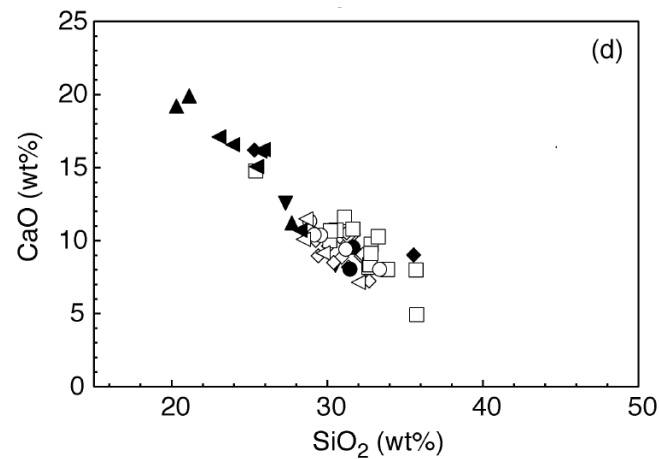
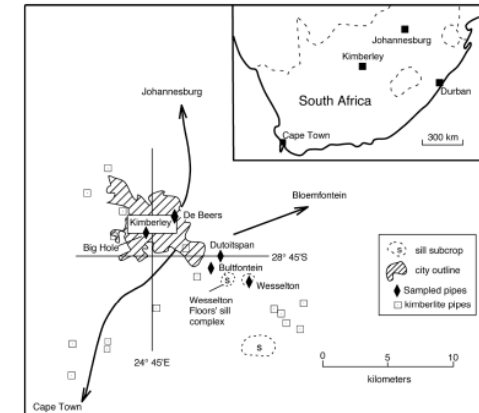
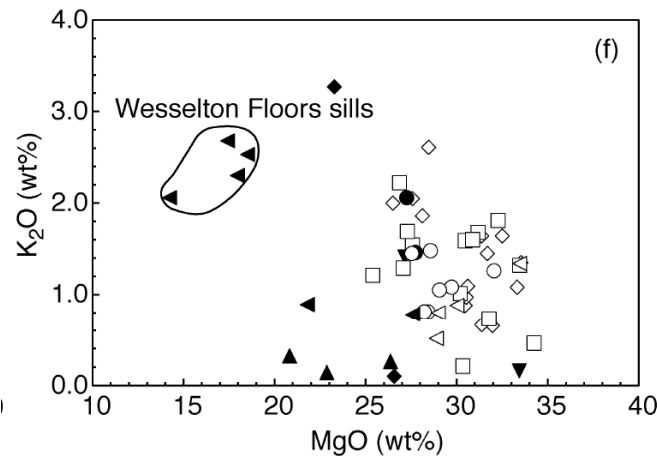
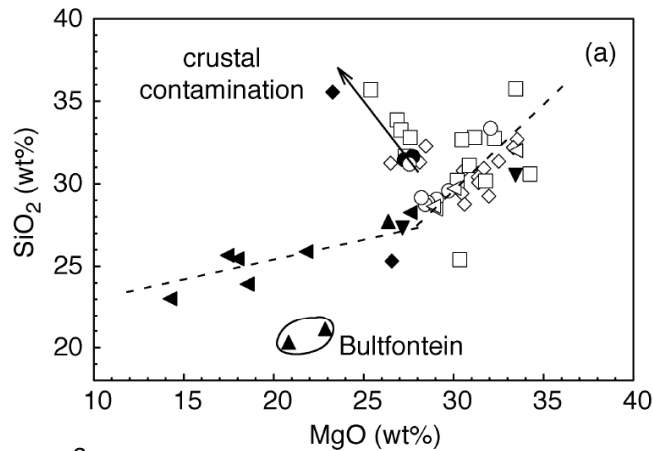
Foley et al., 2002

Nepheline	$\text{Na}(\text{AlSiO}_4)$ or $\text{Na}_3\text{K}(\text{Al}_4\text{Si}_4\text{O}_{16})$
Melilite	$(\text{Ca},\text{Na})_2(\text{Al},\text{Mg},\text{Fe}^{2+})(\text{Si},\text{Al})_2\text{O}_7$
Phlogopite	$\text{KMg}_3[(\text{AlSi}_3)\text{O}_{10}(\text{F},\text{OH})_2]$
Perovskite	$\text{CaTiO}_3$
Apatite	$(\text{Ca},\text{Ba},\text{Pb},\text{Sr},\text{etc.})_5(\text{PO}_4,\text{CO}_3)_3(\text{F},\text{Cl},\text{OH})$

**Lamprophyres from Antarctica in a TAS diagram. Note the strongly  $\text{SiO}_2$ -undersaturated nature that results from very low degrees of melting (<2%) and a great melting depth (>120 km)**



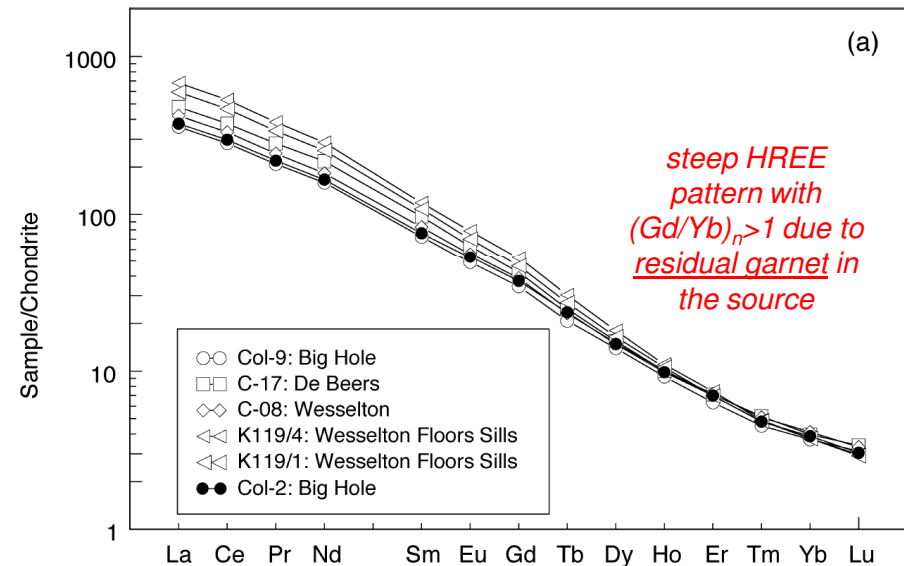
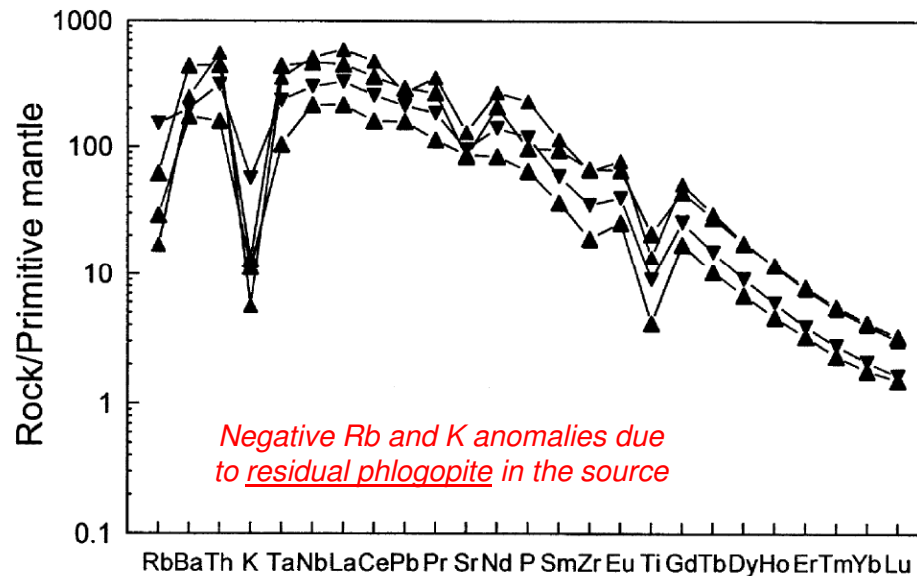
## ‘Exotic’ rock types in continental magmatic settings: Kimberlites from Kimberley, South Africa



**Kimberlites** originate from the interface of the convecting mantle and the lowermost **cratonic continental lithosphere** – they represent the deepest known melting depths, and ultra-low degrees of melting (~1%)

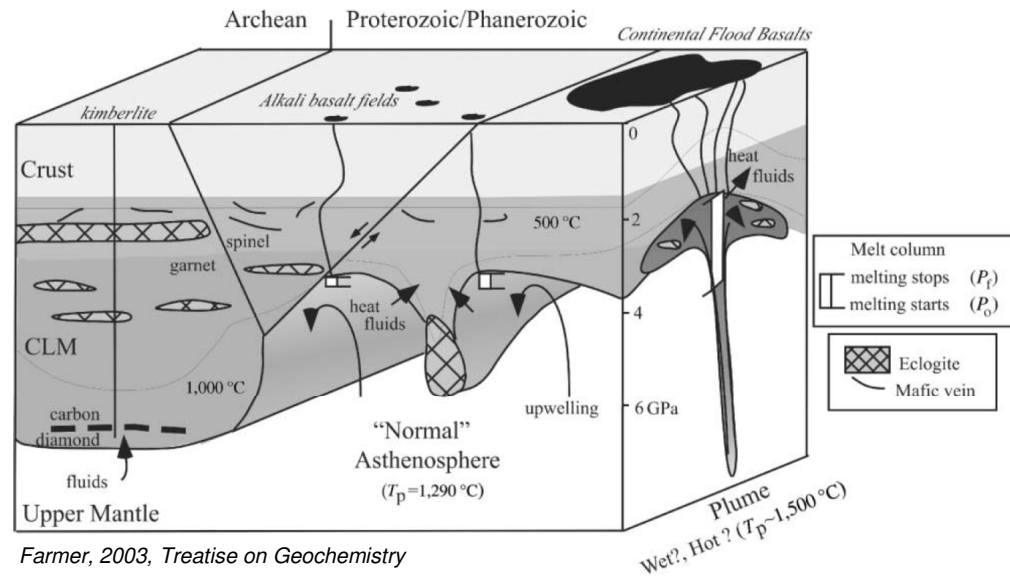
## ‘Exotic’ rock types in continental magmatic settings: Kimberlites from Kimberley, South Africa

Le Roex et al., 2003, *Journal of Petrology*



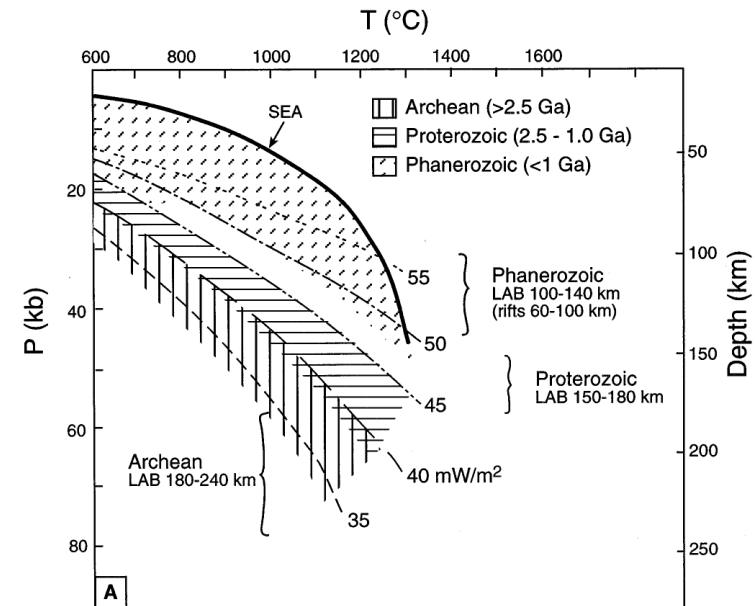
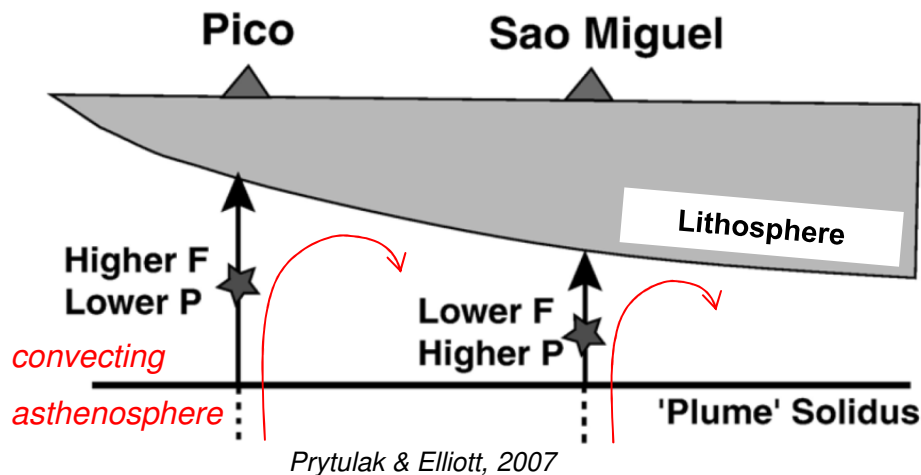
Extremely enriched trace-element patterns (very high La/Yb-ratios) due to an enriched (metasomatised), carbonate-bearing **garnet-peridotite** source (cratonic mantle) and **very low degrees of melting** (great melting depths **>160 km**)

## The role of lithosphere thickness



Beside the **potential temperature** of the convecting mantle, the **degree of melting** broadly depends on the **thickness of the lithosphere**

*Geothermal gradients in Phanerozoic, Proterozoic and Archean Continental Regions estimated from heat flow measurements*



### Summary I: Continental (anorogenic) magmatism

- **Rock types** are predominantly **alkaline** SiO<sub>2</sub>-undersaturated and saturated volcanic and plutonic series (basanites, tephrites, alkali basalts, ...)
- CAM is mostly observed in **extensional geodynamic settings** where **lithosphere thinning** causes **mantle upwelling** and **decompression melting**
- During melting, **asthenospheric** and (metasomatised) **lithospheric** mantle sources are involved, leading to overall **enriched** trace-element compositions

### Summary II: Continental (anorogenic) magmatism

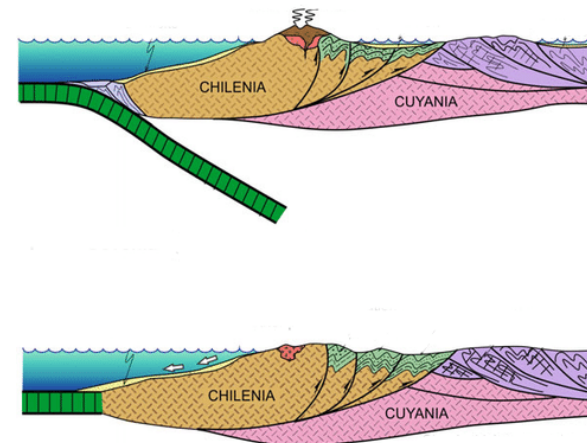
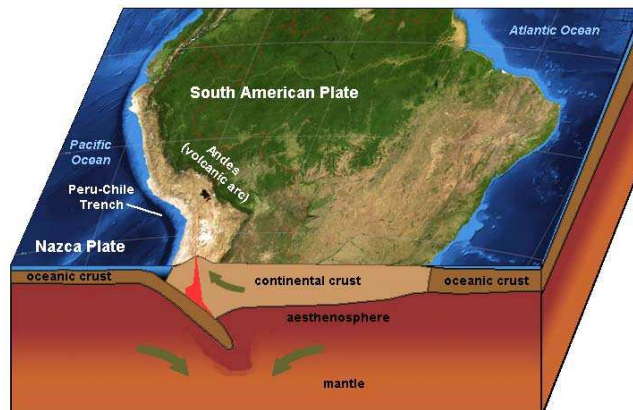
- The degrees of melting are lower (3-7%) than in oceanic settings, dependent on mantle potential **temperature**, lithosphere **thickness** and **volatile content**
- **Crustal assimilation** and **fractional crystallisation** (AFC) play a role where primary magmas **differentiate** to more evolved series (phonolites, trachytes, ...)
- ,Exotic' rocks such as **Carbonatites**, **Kimberlites** or **Lamprophyres** form by ultra-low degrees of melting from highly volatile-rich, **very deep** (lithospheric) mantle sources

Chapter 6

# Convergent plate margins

## *Continental Arcs*

## ***Continental Collision Zones***



**How can we generate  
continental crust from a mantle,  
whose primary melting  
products are basaltic?**

*Two aspects need to be addressed:*

- 1) ,Real' formation of juvenile continental crust from mantle melting*
- 2) Remelting and internal differentiation of pre-existing (continental) crust*

# What is continental crust from a geochemical point of view?

Estimates from Rudnick & Gao, 2003

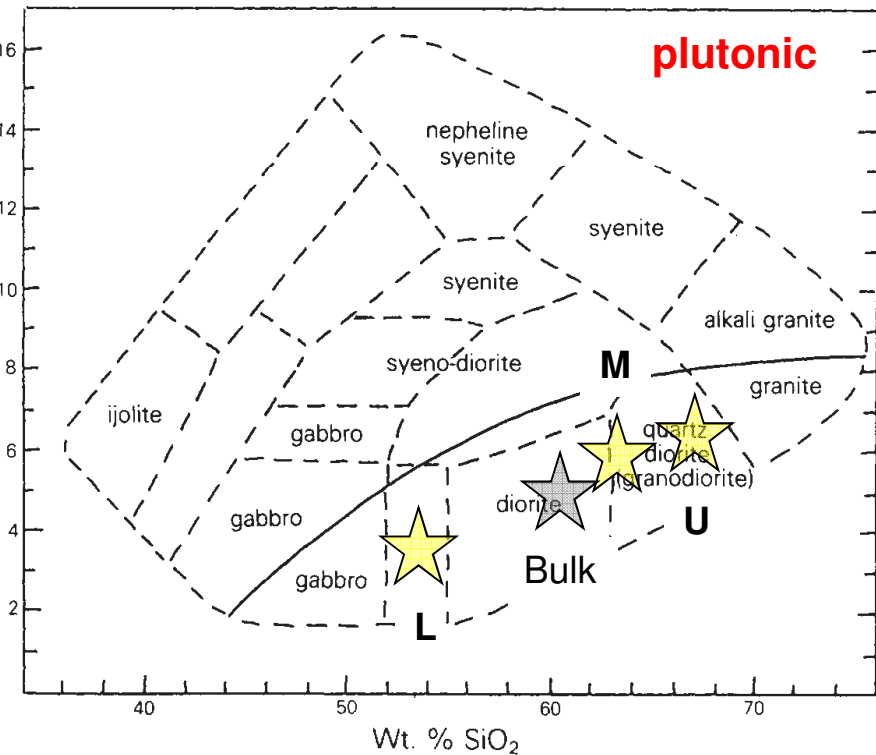
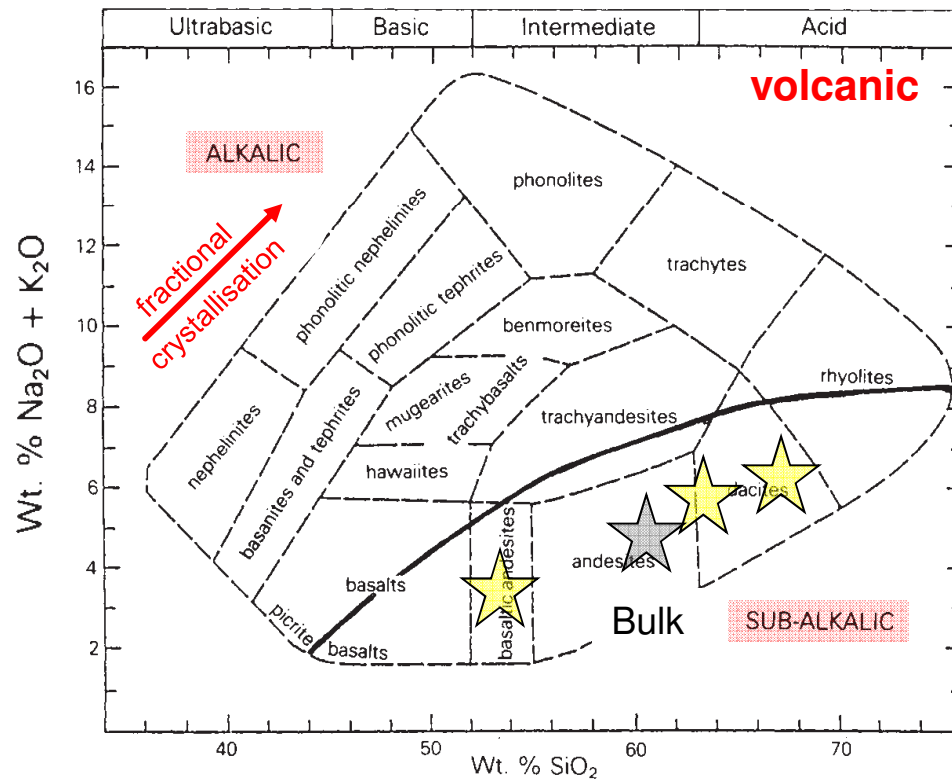
**Table 11** Comparison of the upper, middle, lower and total continental crust compositions recommended here.

<i>Element</i>	<i>Upper crust</i>	<i>Middle crust</i>	<i>Lower crust</i>	<i>Total crust</i>
SiO <sub>2</sub>	66.6	63.5	53.4	60.6
TiO <sub>2</sub>	0.64	0.69	0.82	0.72
Al <sub>2</sub> O <sub>3</sub>	15.4	15.0	16.9	15.9
FeO <sub>T</sub>	5.04	6.02	8.57	6.71
MnO	0.10	0.10	0.10	0.10
MgO	2.48	3.59	7.24	4.66
CaO	3.59	5.25	9.59	6.41
Na <sub>2</sub> O	3.27	3.39	2.65	3.07
K <sub>2</sub> O	2.80	2.30	0.61	1.81
P <sub>2</sub> O <sub>5</sub>	0.15	0.15	0.10	0.13
Total	100.05	100.00	100.00	100.12
Mg#	46.7	51.5	60.1	55.3

# What is continental crust from a geochemical point of view?

Estimates from Rudnick & Gao, 2003

The continental crust comprises **0.57 %** of the BSE, but accounts for **20-70%** of incompatible and highly incompatible elements

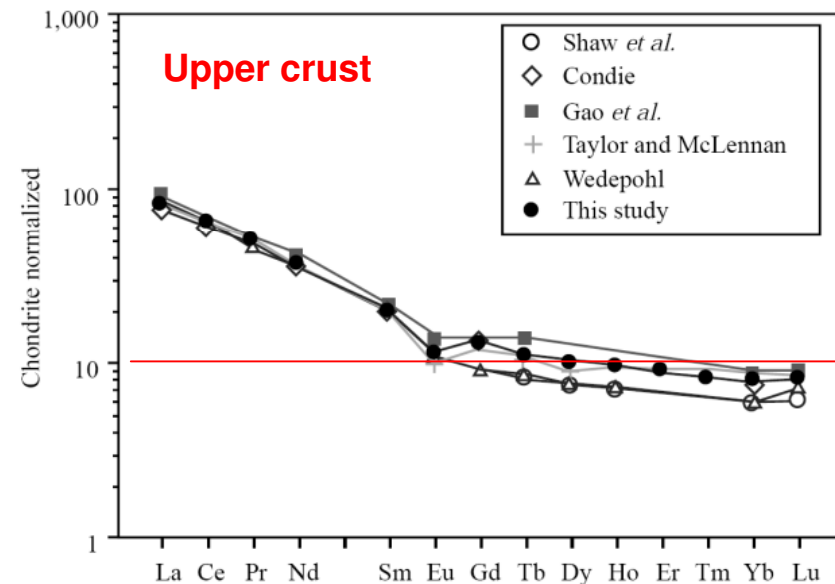
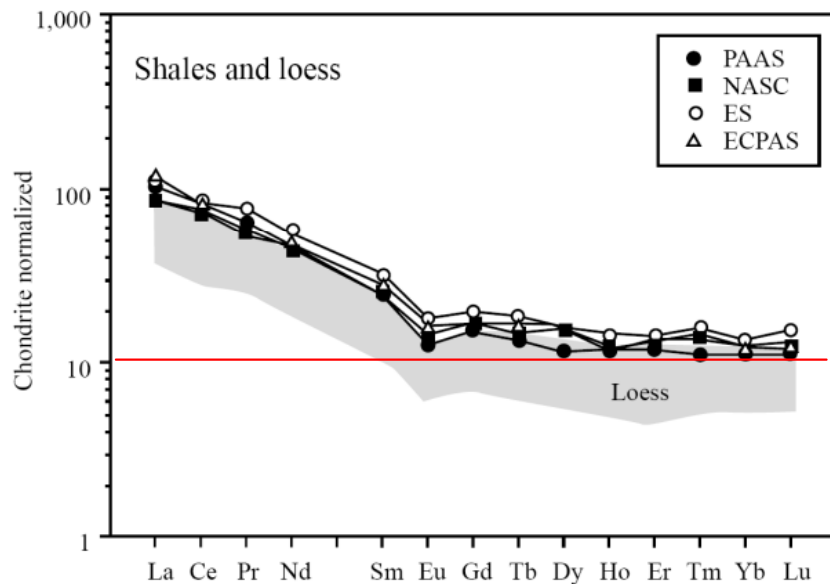


From: Wilson, M.: *Igneous Petrogenesis*, Springer.  
Crustal estimates from Rudnick & Gao, 2003

L = lower, M = middle, U = upper, Bulk = whole continental crust

# What is continental crust from a geochemical point of view?

Upper continental crust

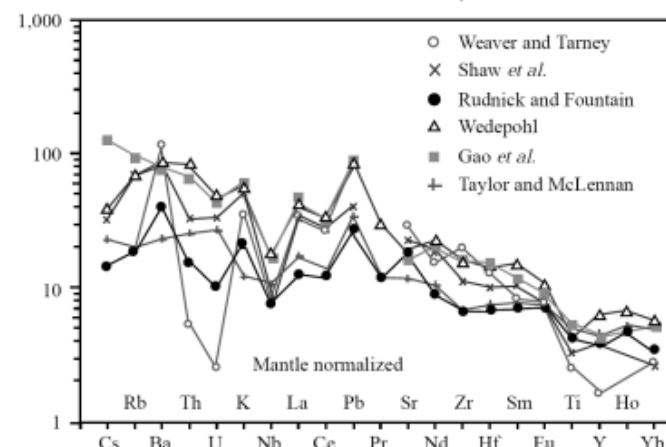
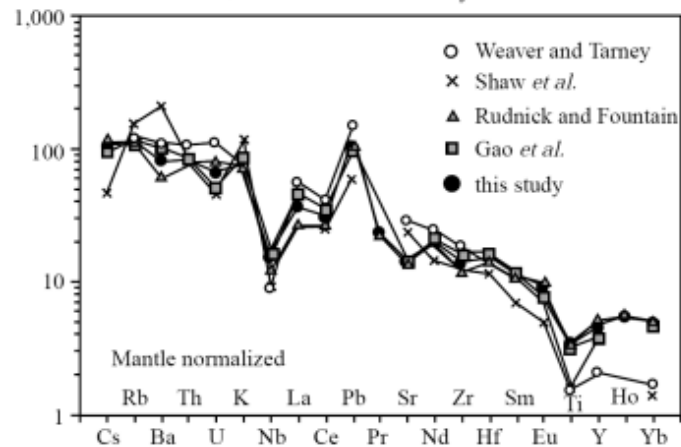
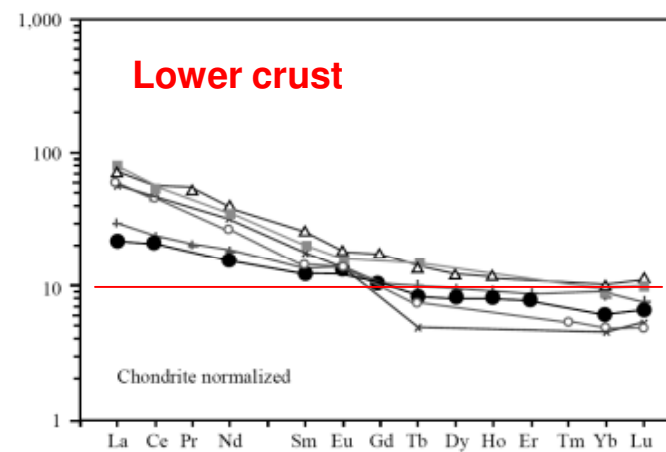
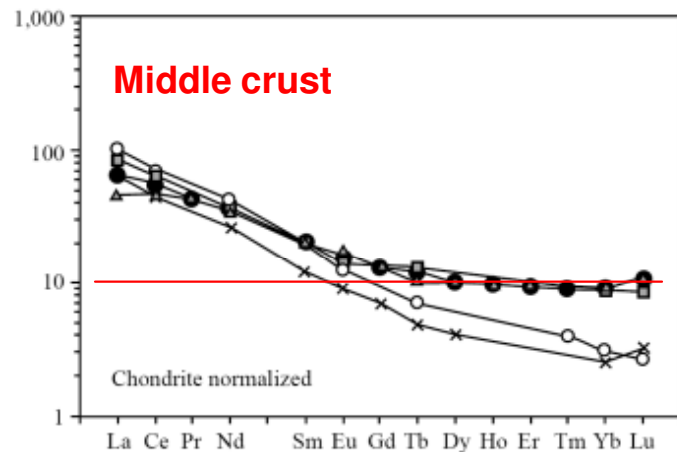


PAAS = Post Archean Australian Shale  
 NASC = North American Shale Composition  
 ES = European Shale Composition  
 ECPAS = Eastern China post-Archean Shale

**Trace-element composition of the upper continental crust as estimated from shales**

## What is continental crust from a geochemical point of view?

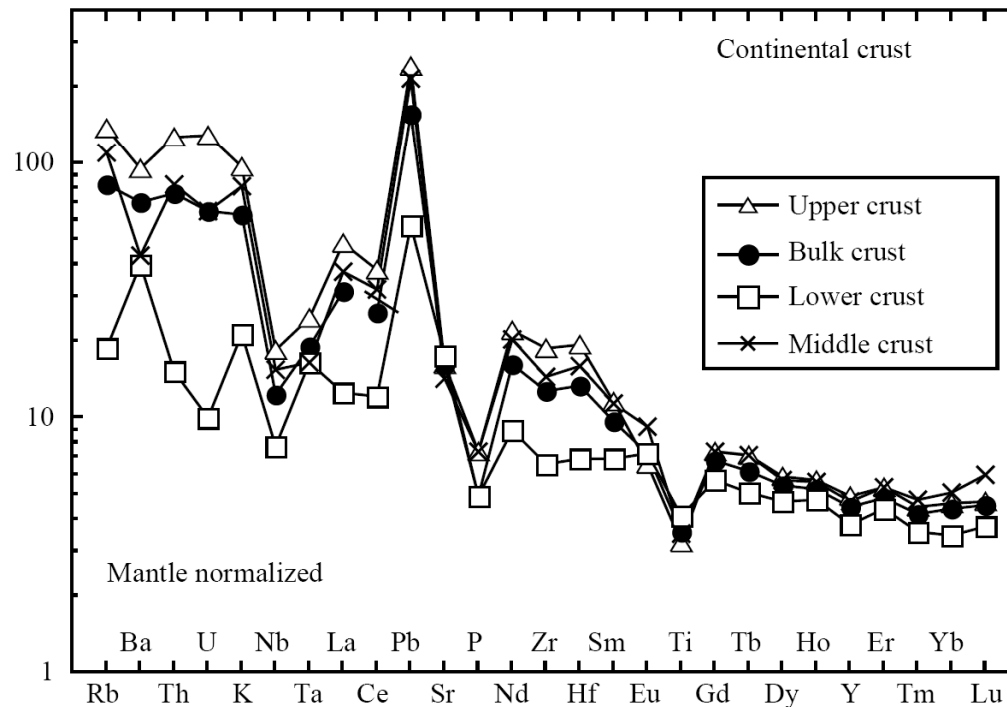
Middle and lower continental crust



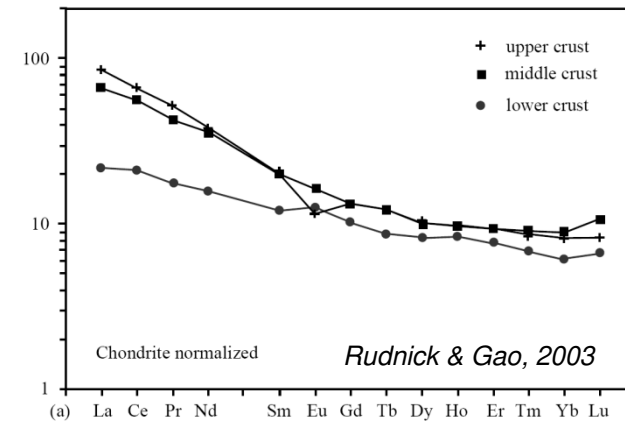
Note that different authors provide fairly **different compositions**, in particular for moderately incompatible elements (e.g. REE; „**sampling bias**“)

# What is continental crust from a geochemical point of view?

Upper, middle and lower continental crust



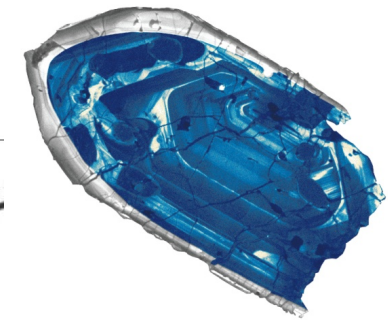
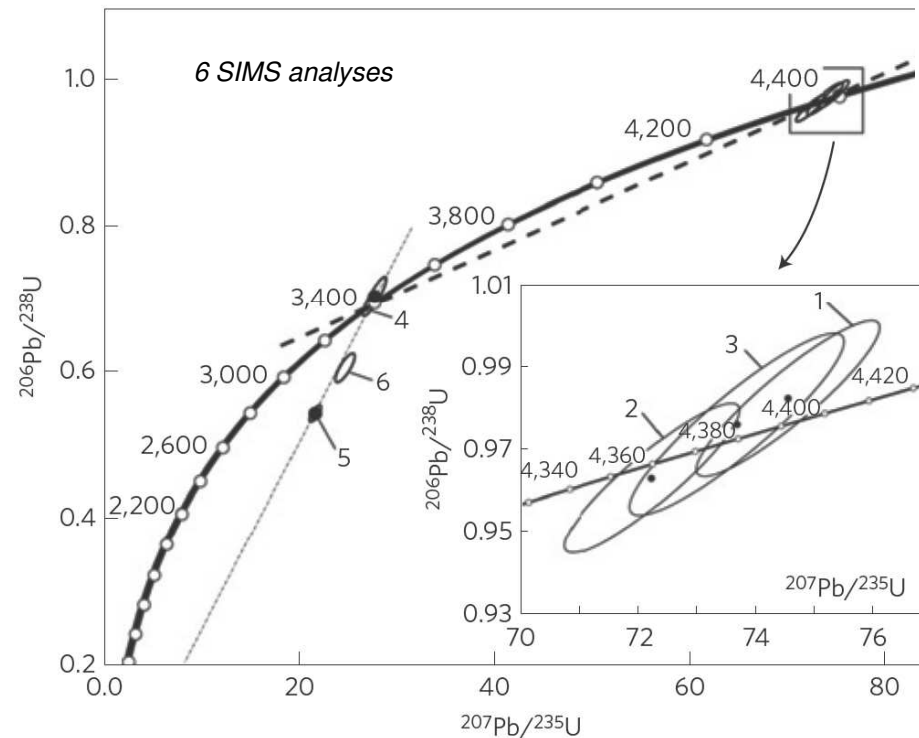
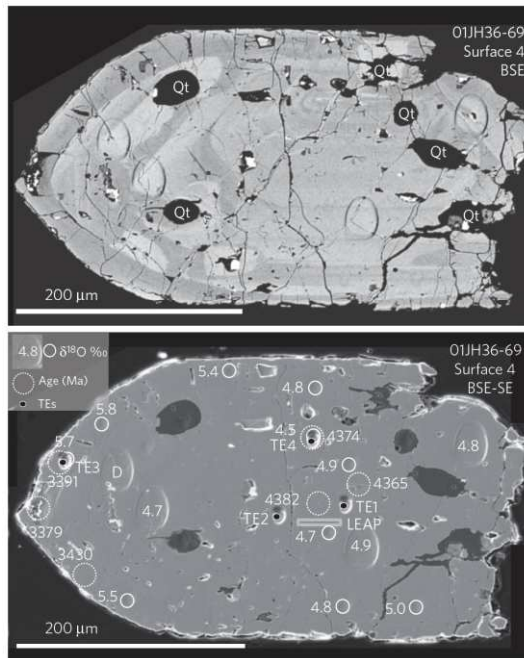
Kemp & Hawkesworth, 2003, *Treatise on Geochemistry*



Bulk continental crust has an overall **andesitic composition**, and is **enriched** in incompatible trace-elements. Therefore, beside other **accessory phases** (monazite, apatite, xenotime), intermediate to felsic rocks contain abundant **zircon**

## The earliest (continental?) crust on Earth

Zircons from Jack Hills, Australia

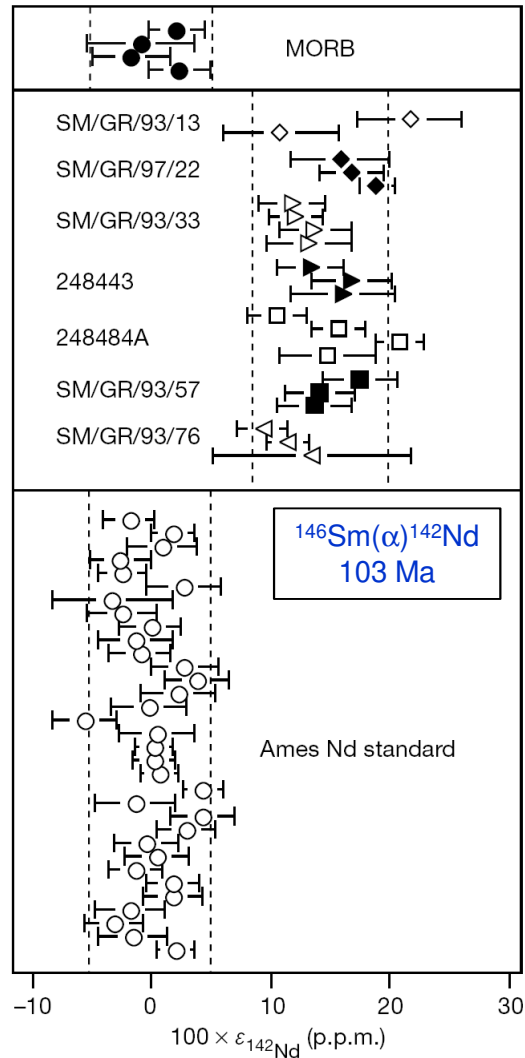


Zircon from Jack Hills, Yilgarn Craton, Western Australia, with a **core** age of ~4.37 Ga

Valley et al., 2014, Nature

Using SIMS and APT, Valley et al. (2014) determined a **concordant** age of **4.374 ± 0.006 Ga** for the **core of a detrital zircon** from the **Jack Hills, Western Australia**. This indicates that **silicate differentiation** on Earth already took place less than ~200 Ma after Earth accretion

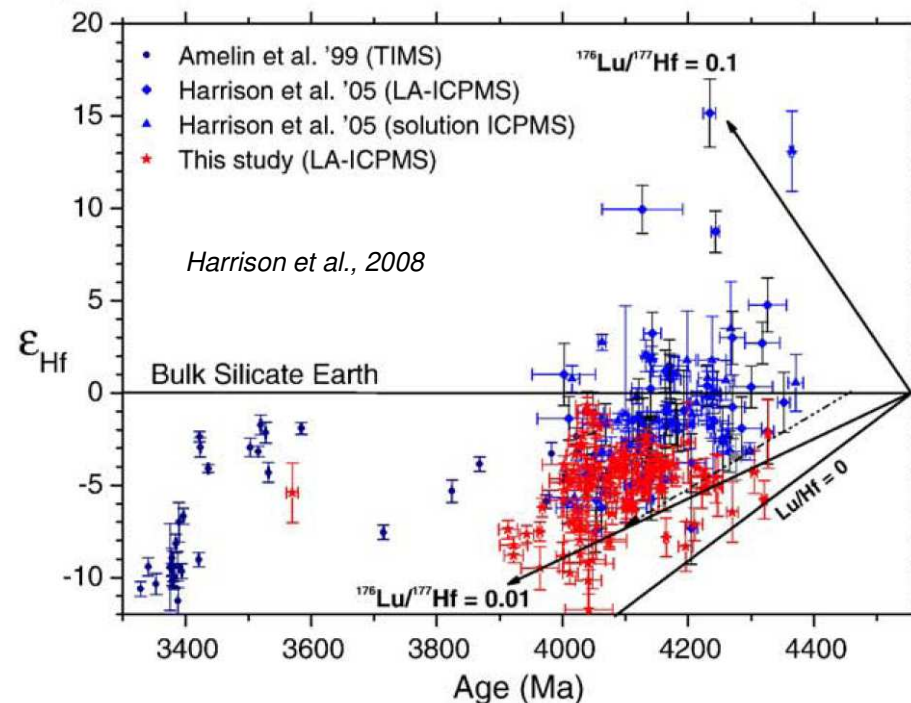
## The earliest (continental?) crust on Earth



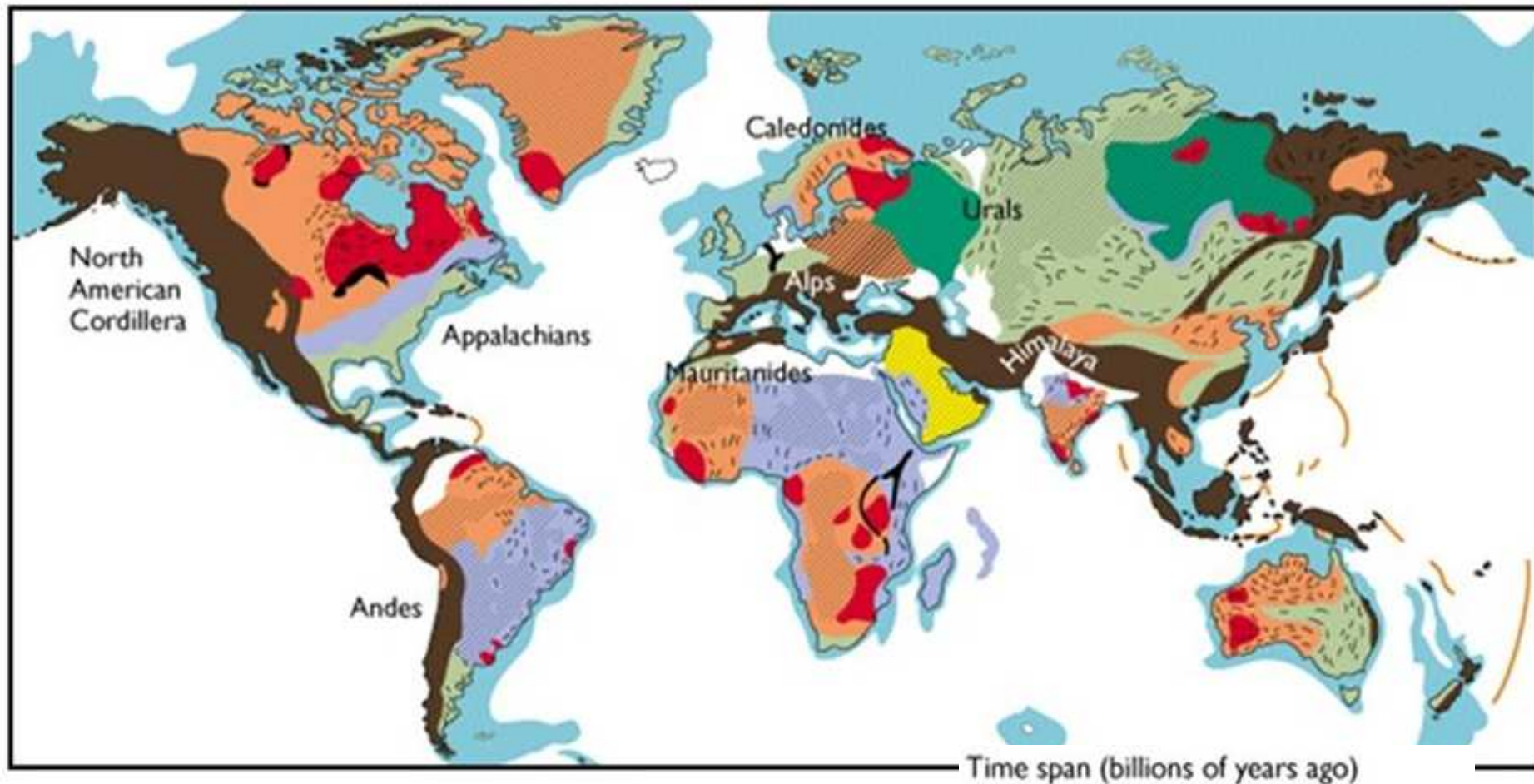
Caro et al., 2003, Nature

Positive  $^{142}\text{Nd}$  anomalies ( $^{142}\text{Nd}$  excess) in meta-sediments (~3.7 – 3.8 Ga old) from the **Isua Greenstone Belt** (West Greenland) indicate very early mantle differentiation at **~4.460 Ga ( $\pm$  0.115)**.

$^{176}\text{Hf}/^{177}\text{Hf}$  anomalies (negative  $\epsilon_{\text{Hf}}$ ) in zircons from Jack Hills confirm very early silicate (mantle-crust) differentiation.

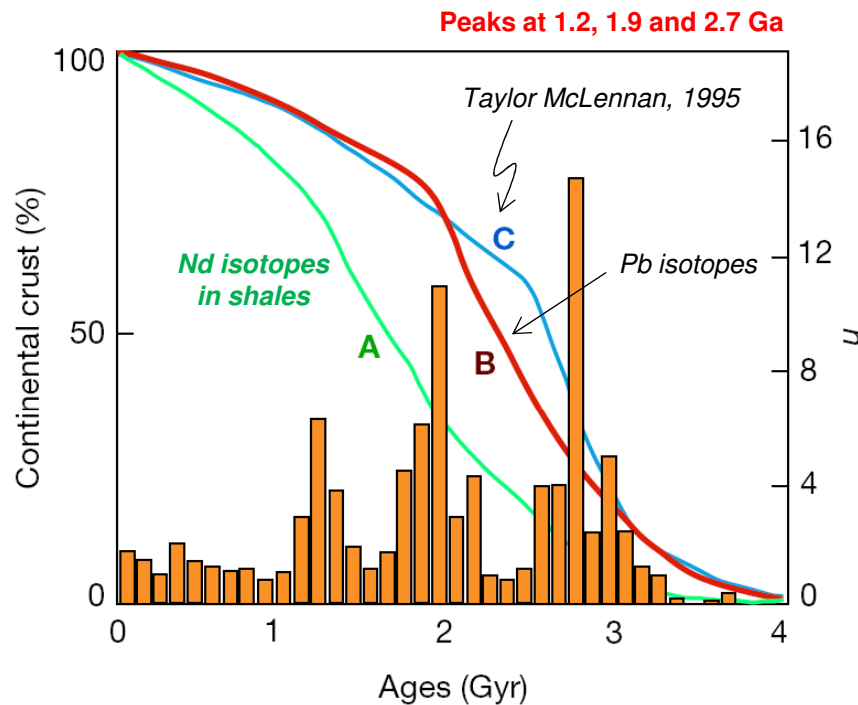


## Age distribution of the continental crust

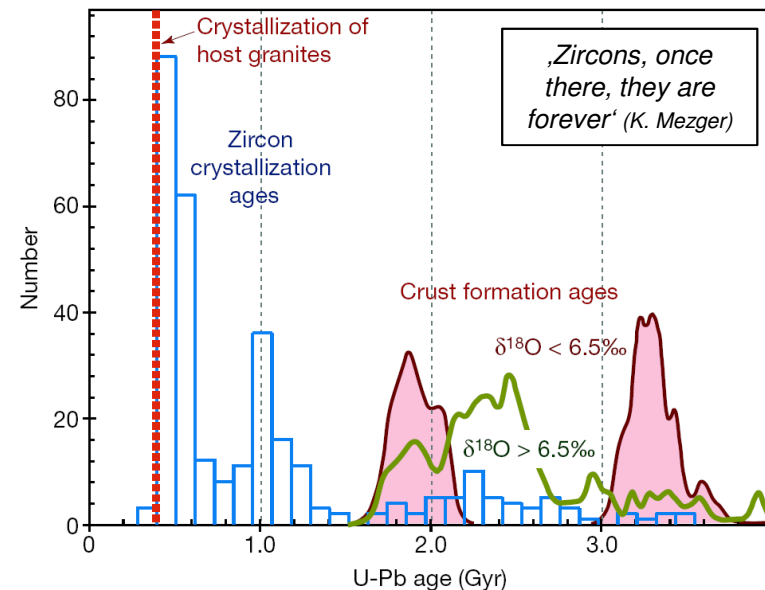
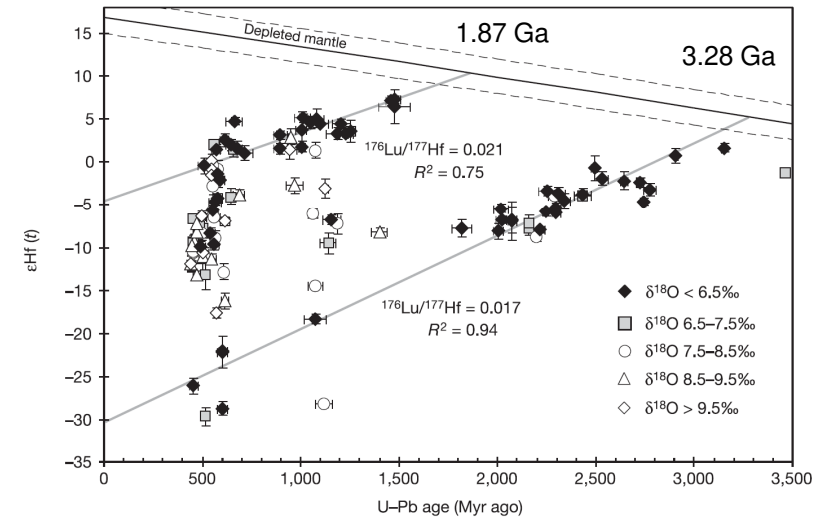


Did the volume of the continental crust continuously increase over time or was it about constant?

## Growth rates of the continental crust

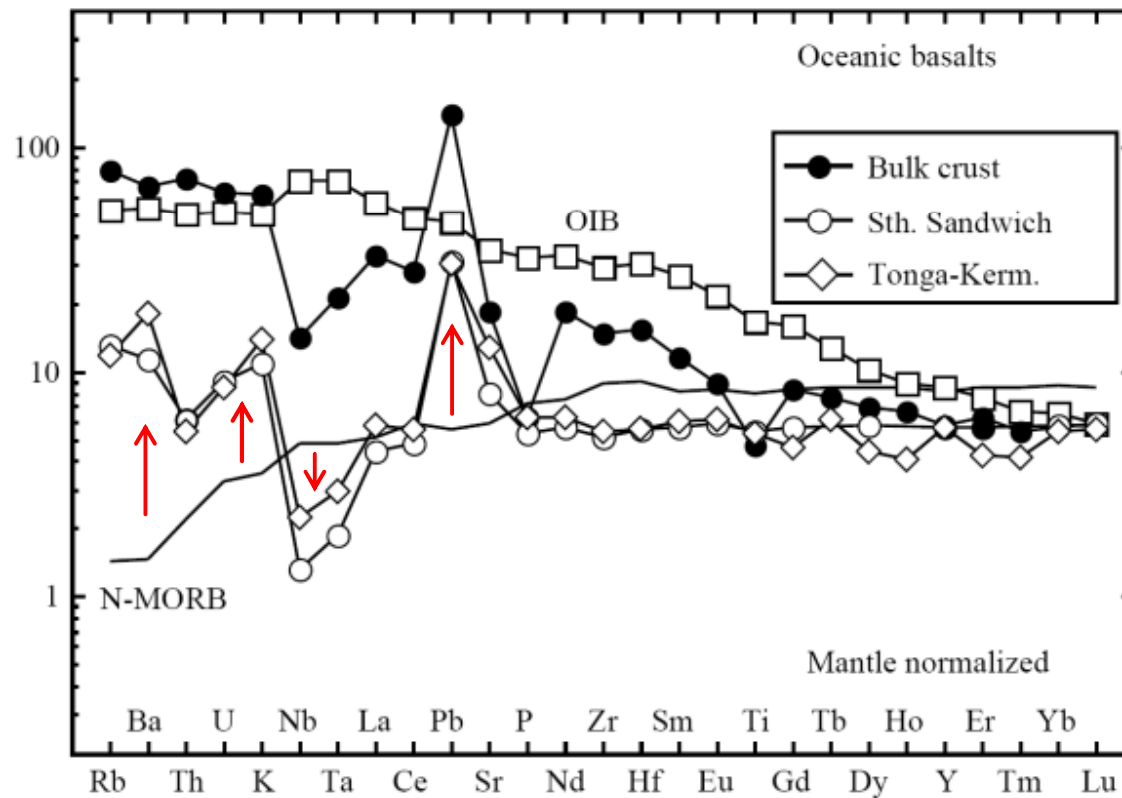


Histogram of ages of **crustal igneous rocks** with mantle-like initial isotope compositions and crustal growth curves based on different approaches



# Comparison of the composition of the bulk crust with that of island arc magmas

Crust formation in island arcs



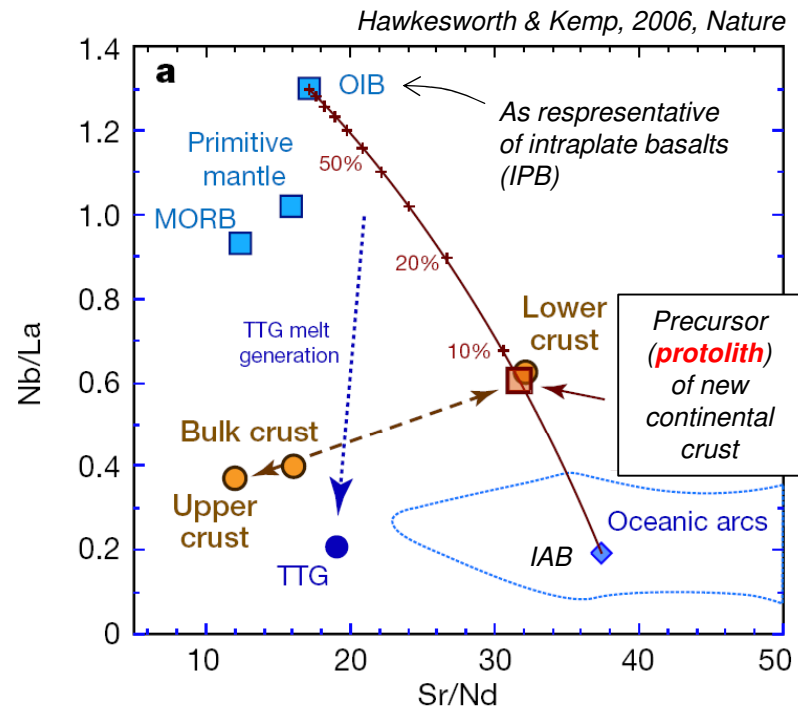
Kemp & Hawkesworth, 2003, *Treatise on Geochemistry*

Note the **LILE, Th, U and Pb enrichment**, but **Nb, Ta and HREE depletion** in IAB relative to MORB („slab-effect“)

Several features (**Nb, Ta and Pb anomalies**) are also observed in **bulk crust**, but overall concentrations of incompatible trace-elements are higher.

**The precursors of new continental crust result from subduction zone magmatism**

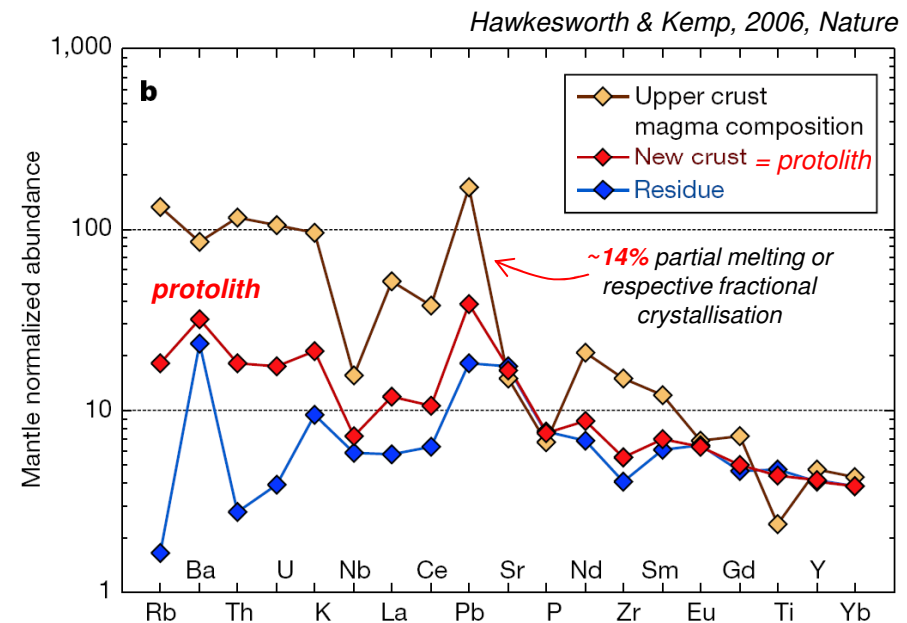
## First order Formation mechanisms of continental crust



**Nb/La** used as a „source“ proxy, **Sr/Nd** mirrors intracrustal differentiation (residual plagioclase retains Sr!).

### PROBLEM: RESIDUUM

Beside island arc magmatism, **intraplate-magmatism** may contribute in some way to the formation of new continental crust



*No or little garnet involved during average intracrustal differentiation!  
I.e., no thickened crust required .....*

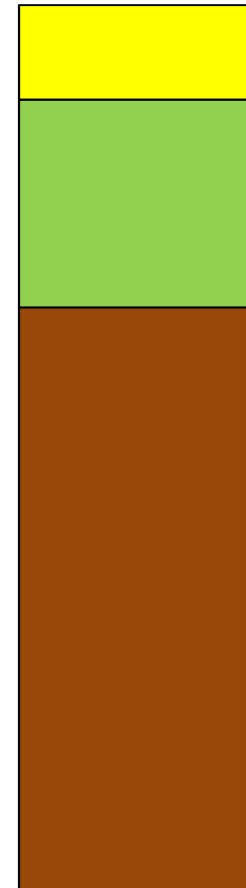
## First order Formation mechanisms of continental crust

### Summary

*based on Hawkesworth & Kemp, 2006, Nature*

Residuum is **founded back**  
into the **mantle** due to a **density**  
**inversion** caused by the  
formation of **high-density**  
**phases** such as garnet

**Back to the mantle**  
(recycling)



**Upper crust**, ~12.5 km thick,  
solidified partial melt of a  
basaltic precursor

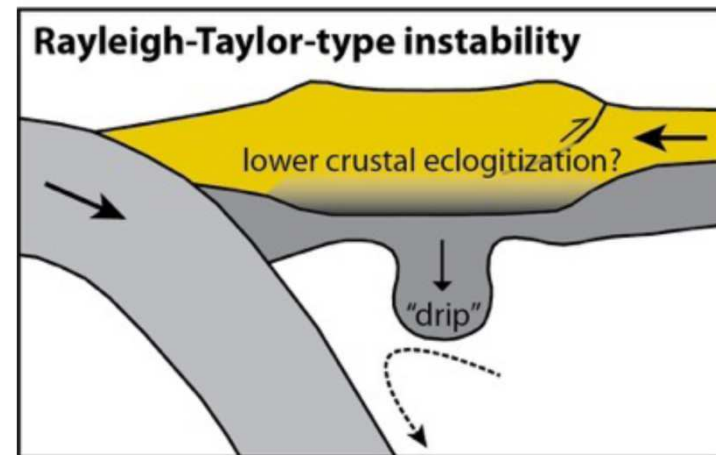
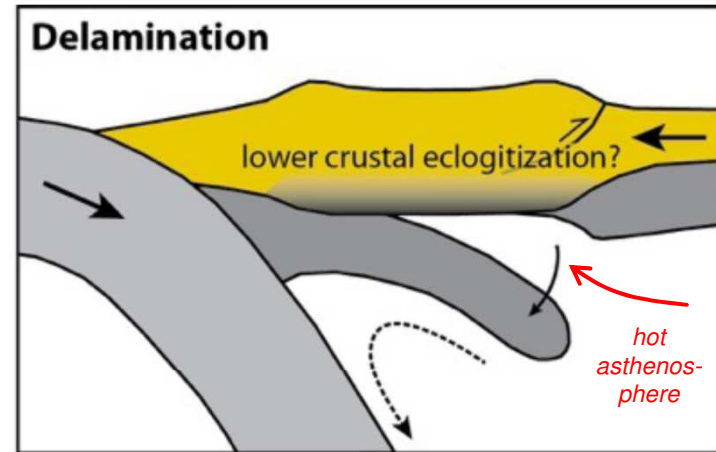
**Middle and lower**  
**crust**, ~27.5 km thick,  
mixture of upper crust,  
precursor and residuum

**Residuum** of basaltic  
precursor after ~14%  
partial melting, about  
77 km thick

## First order Formation mechanisms of continental crust

### DELAMINATION

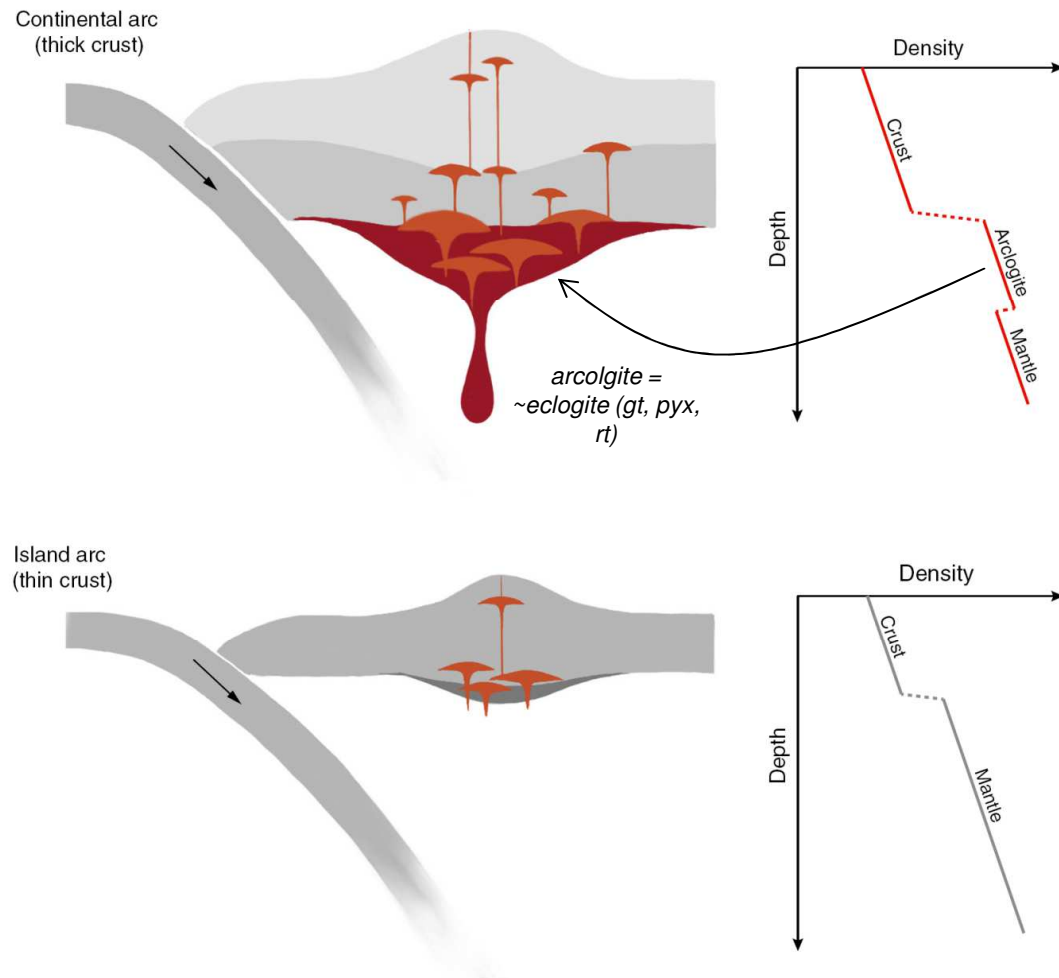
**Delamination** (foundering) of **lower crust** needs to result in (1) **rapid uplift** due to isostatic disequilibrium, and (2) intense **magmatism** if hot asthenosphere rises to fill the „gap“





## First order **Formation mechanisms of juvenile continental crust** Role of crustal thickness and density inversion

According to this model, only thick crust as observed in **continental arcs** enables **density inversion** and **delamination**, and recycling of **eclogite** (arclogite) back into the mantle



## Magmatism in (continental) collision zones

**How were crustal rocks, either primary**

(diorites, granodiorites), **intermediate** (meta-igneous

felsic to mafic) **or secondary** (meta-sedimentary)

**processed to form more evolved**

**lithologies** (granites, pegmatites, ...) ?

How can we distinguish between **new**

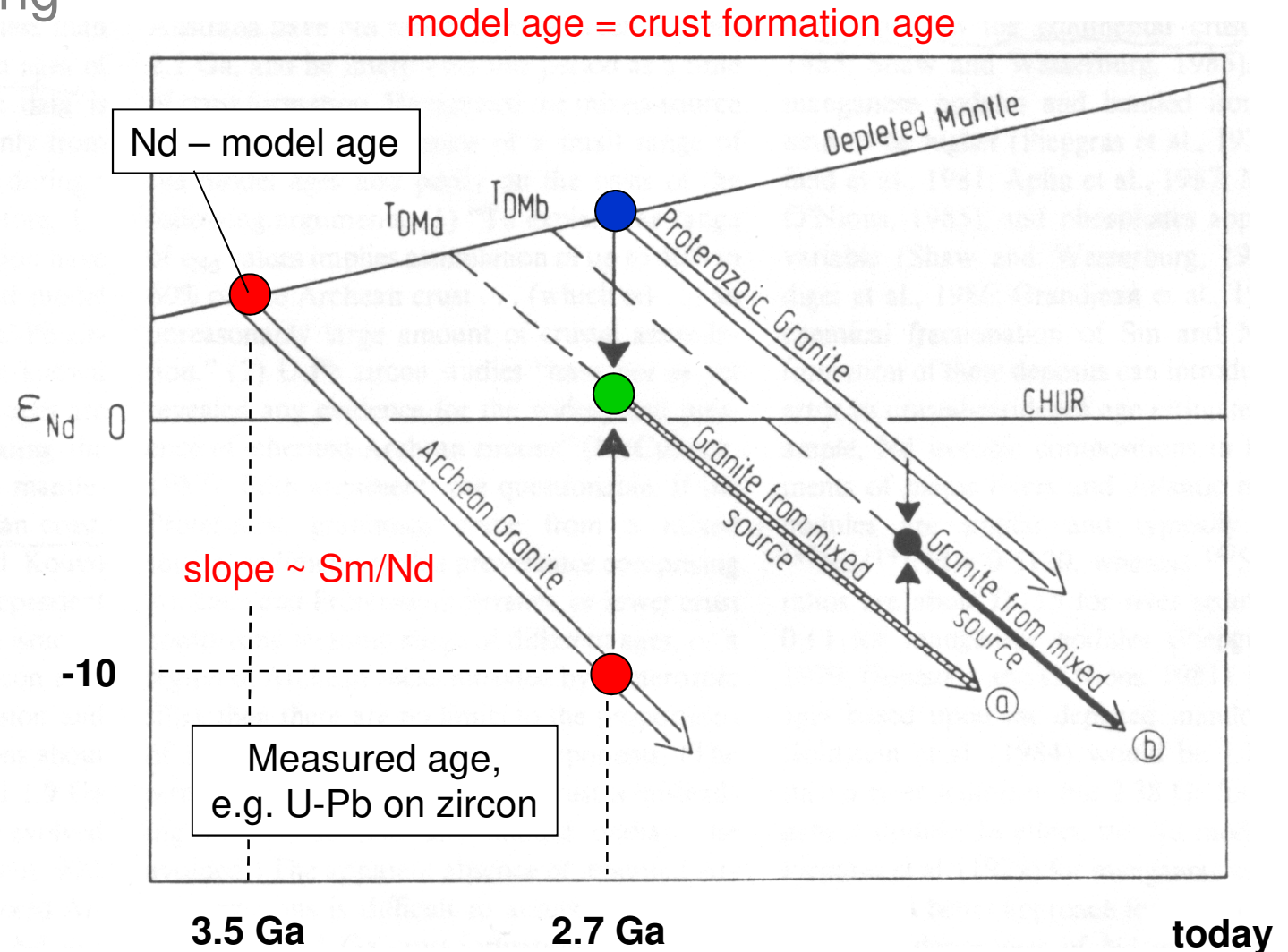
(juvenile) and **reworked** crust?

## Magmatism in (continental) collision zones

Crustal reworking

**Crust formation vs. crustal remelting** ideally resolvable from **Nd-isotope composition** and **zircon age** of a rock sample.

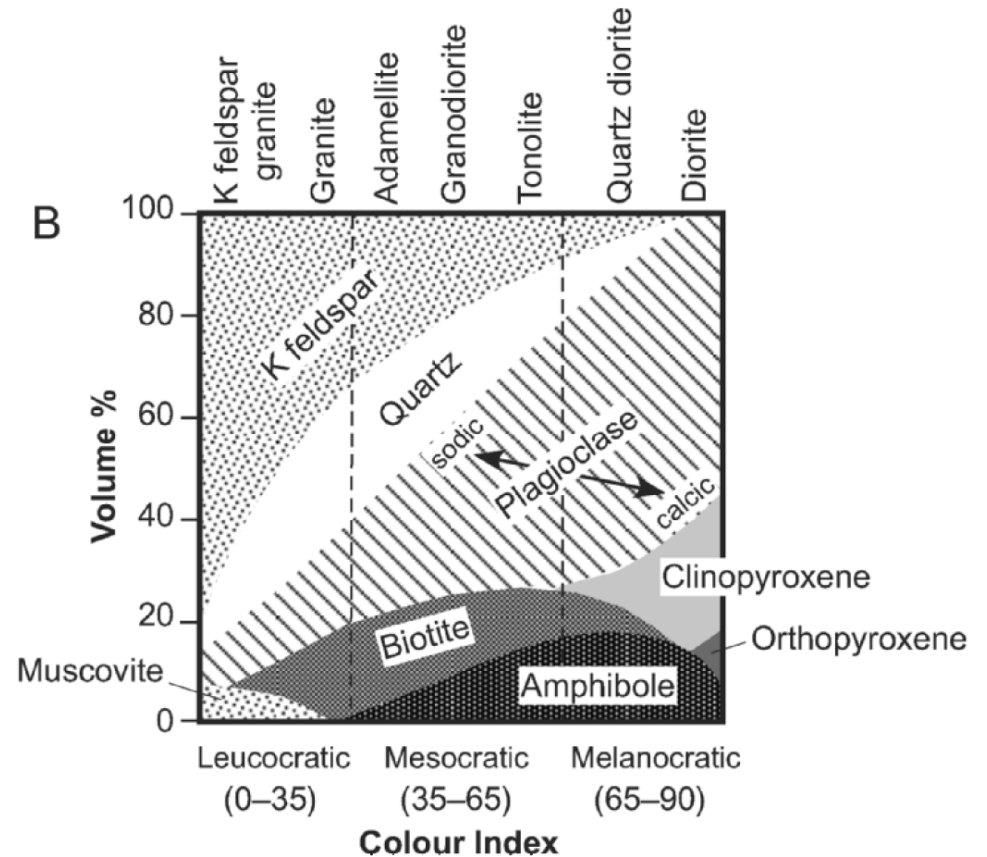
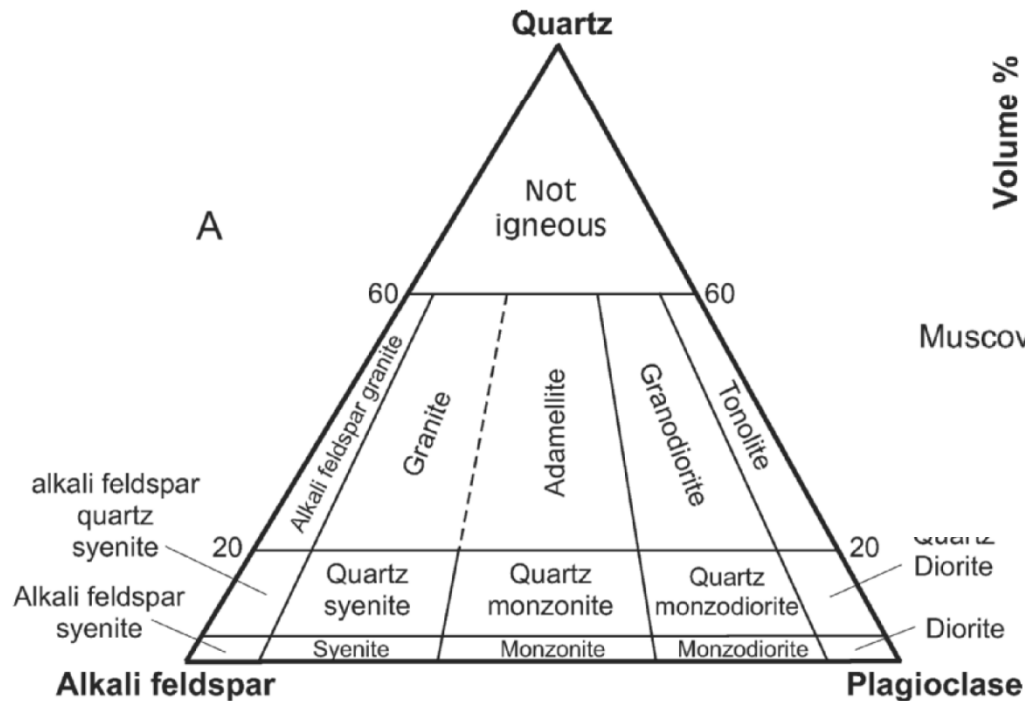
*„If the Nd-model age and the zircon age coincide, the rock is a piece of juvenile crust“*



## Magmatism in (continental) collision zones

### Mineralogy of granite rocks

Chen & Grapes (2007)  
 Classification following Streckeisen



# Magmatism in (continental) collision zones

Terms used to classify granitic rocks (based on Al- and Alkali-content)

**Peraluminous:**  $\text{Molar Al}_2\text{O}_3 > \text{Molar CaO} + \text{K}_2\text{O} + \text{Na}_2\text{O}$  [A > CKN]

(Plag + Kfs have ~1:1). Bt, Ms, Cd, Sil are peraluminous minerals

**Metaluminous:**  $\text{Molar CaO} + \text{K}_2\text{O} + \text{Na}_2\text{O} > \text{Molar Al}_2\text{O}_3 > \text{Molar K}_2\text{O} + \text{Na}_2\text{O}$  [CKN > A > KN]

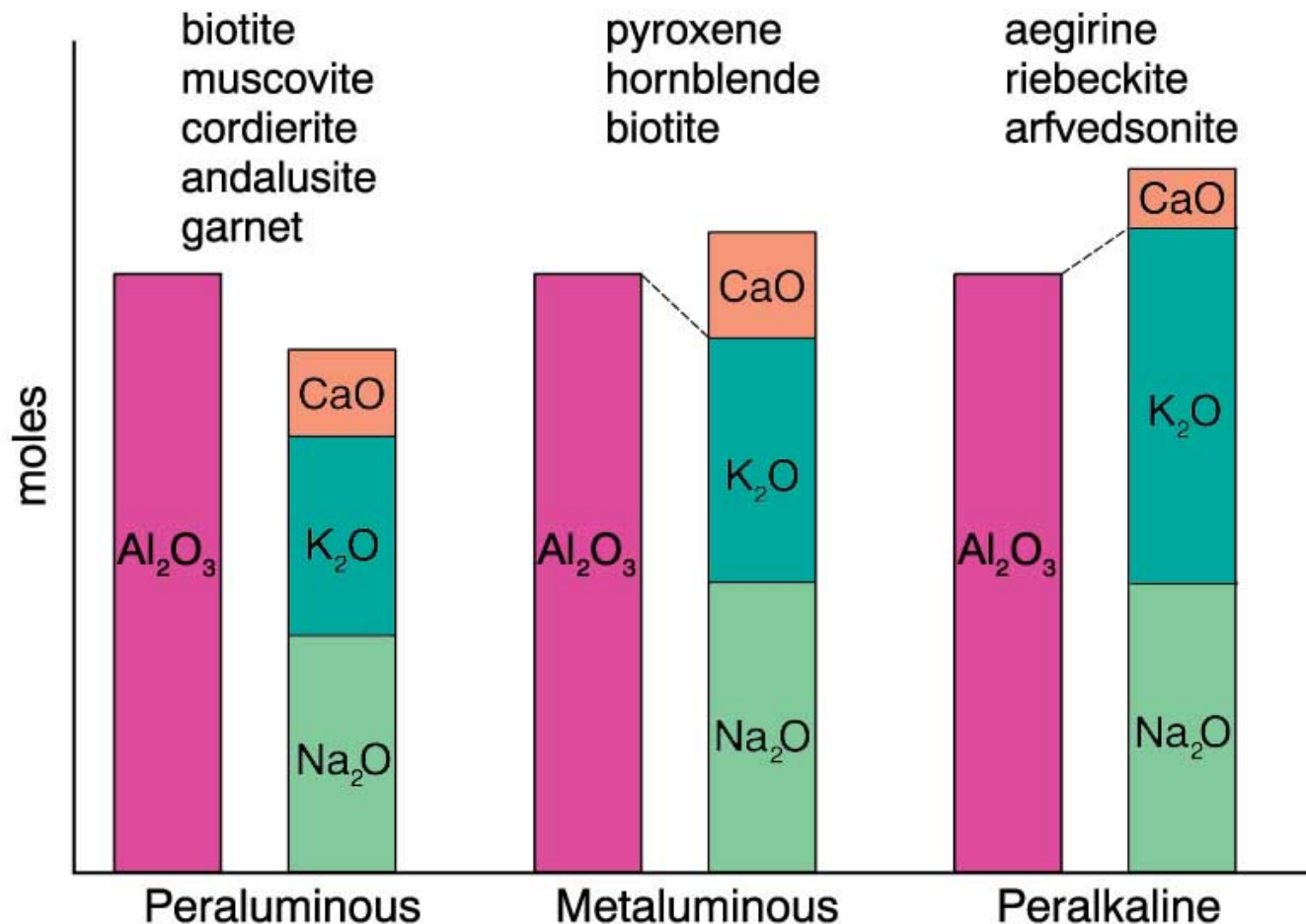
**Subaluminous:**  $\text{Molar Al}_2\text{O}_3 \sim \text{Molar K}_2\text{O} + \text{Na}_2\text{O}$  [A~KN]

**Peralkaline:**  $\text{Molar Al}_2\text{O}_3 < \text{Molar K}_2\text{O} + \text{Na}_2\text{O}$  [A<KN]

**MALI (modified alkali lime Index):**  $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO}$  (nicht molar!)

## Magmatism in (continental) collision zones

Terms used to classify granitic rocks (based on Al- and Alkali-content)



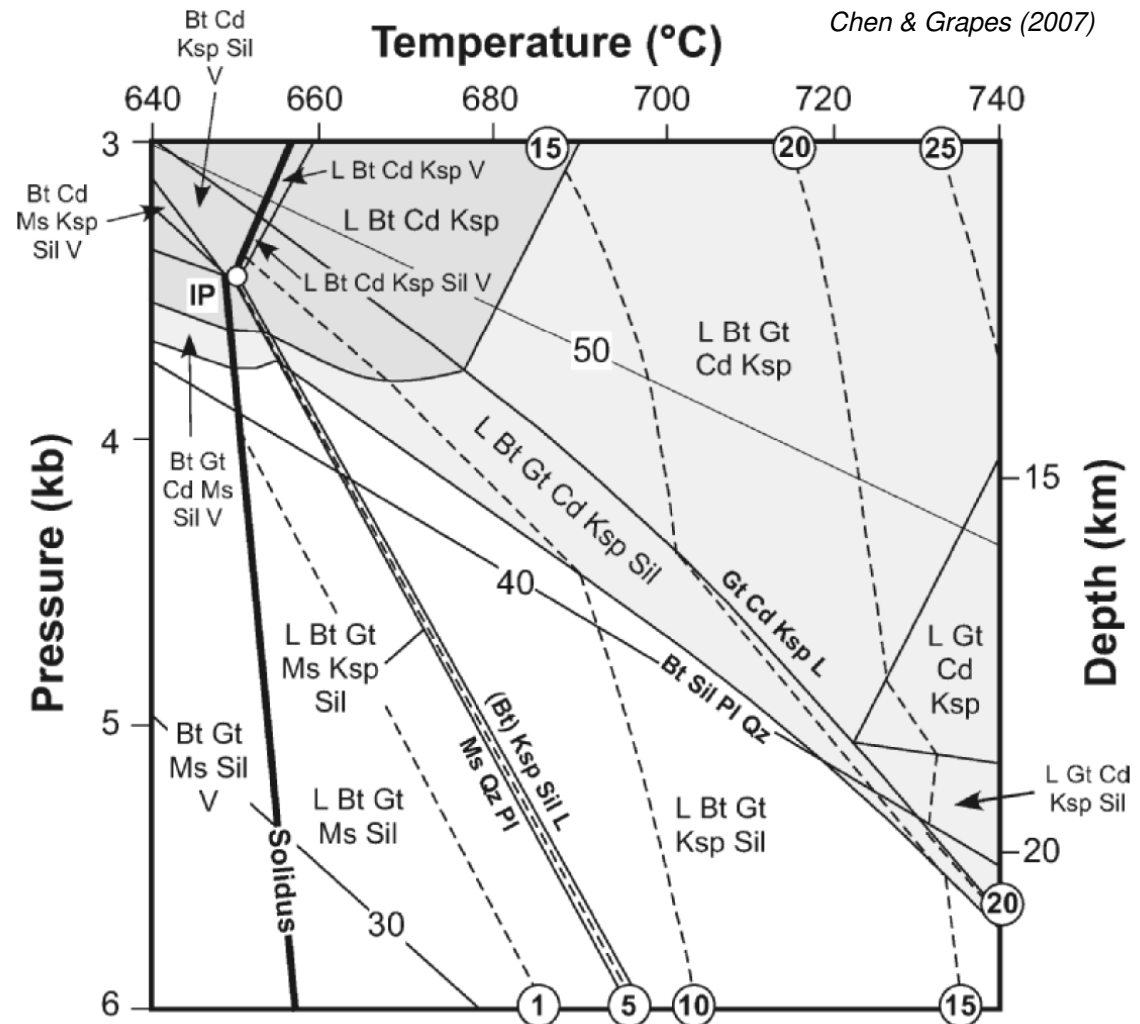
## Magmatism in (continental) collision zones

High-grade metamorphic (metapelite) rocks and anatexis

Equilibrium **phase-relations** (including **partial melts**) in the **MnNCKFMASH** System representing a typical average (meta-) **pelite composition**.

Stable phases – dependent on p-T, are Qz, Ksp, Plag, Bt, Ms *plus* Gt, Cd, Sil and melt.

(Meta-) Pelite (~mol%): 73.6 SiO<sub>2</sub>, 11.9 Al<sub>2</sub>O<sub>3</sub>, 7.2 FeO, 0.10 MnO; 3.9 MgO, 0.35 CaO; 1.4 Na<sub>2</sub>O; 3.1 K<sub>2</sub>O

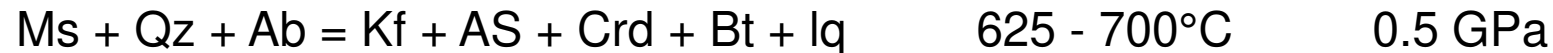


## Magmatism in (continental) collision zones

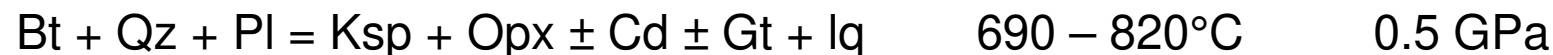
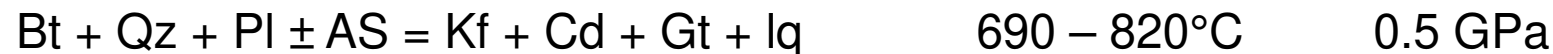
High-grade metamorphic (metapelite) rocks and anatexis

**Partial melting** is either triggered by **free water** (~640°C at 1 GPa) or by the **breakdown of water bearing phases**. Important experimental constraints:

*Muscovite breakdown melting:*



*Biotite breakdown melting:*



*Amphibole breakdown melting:*



## Magmatism in (continental) collision zones

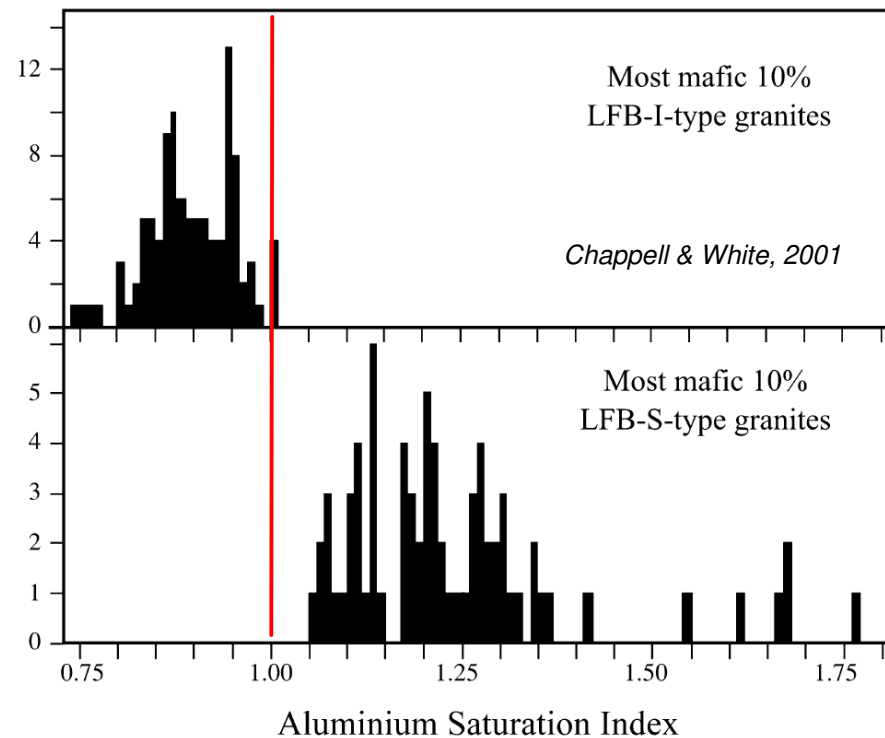
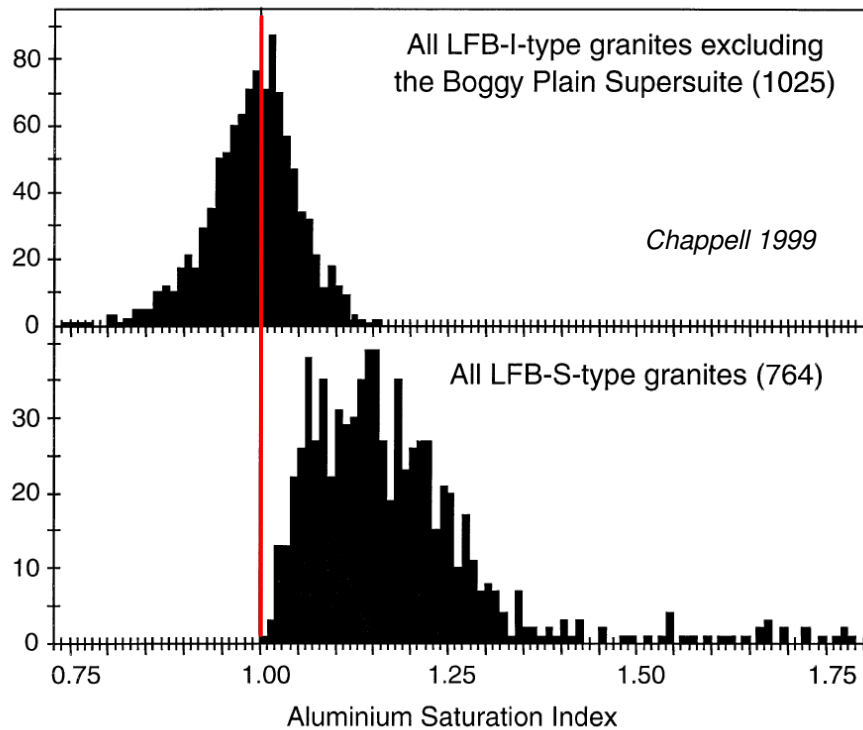
High-grade metamorphic rocks and migmatites

**Anatexis** (of meta-sedimentary or meta-igneous rocks)

and the formation of **‘granitic melts’** is an important process of **crustal reworking** and **intracrustal differentiation** during orogeny, but may **not** produce any juvenile crust!

## Magmatism in (continental) collision zones

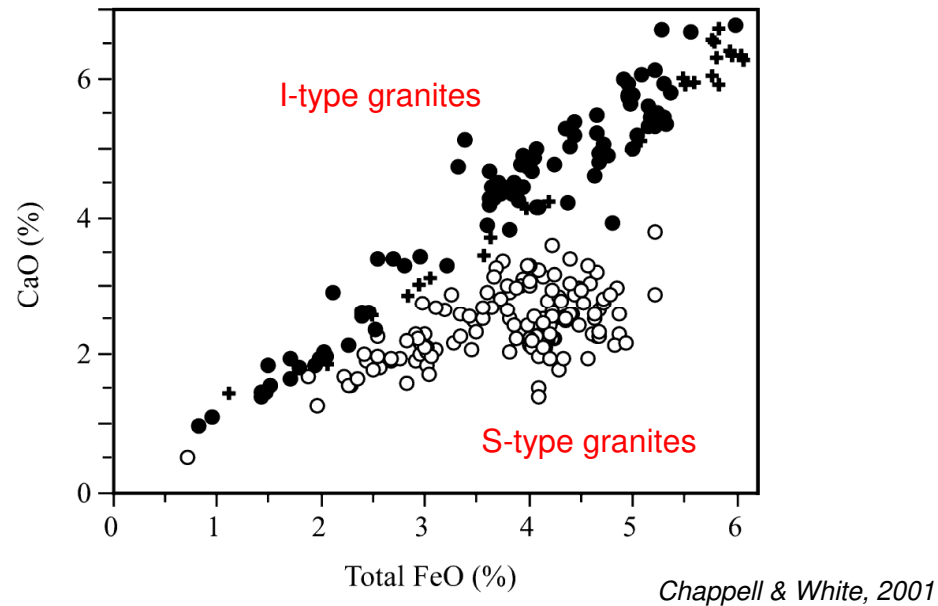
Origin and types of granites – importance of Al-saturation



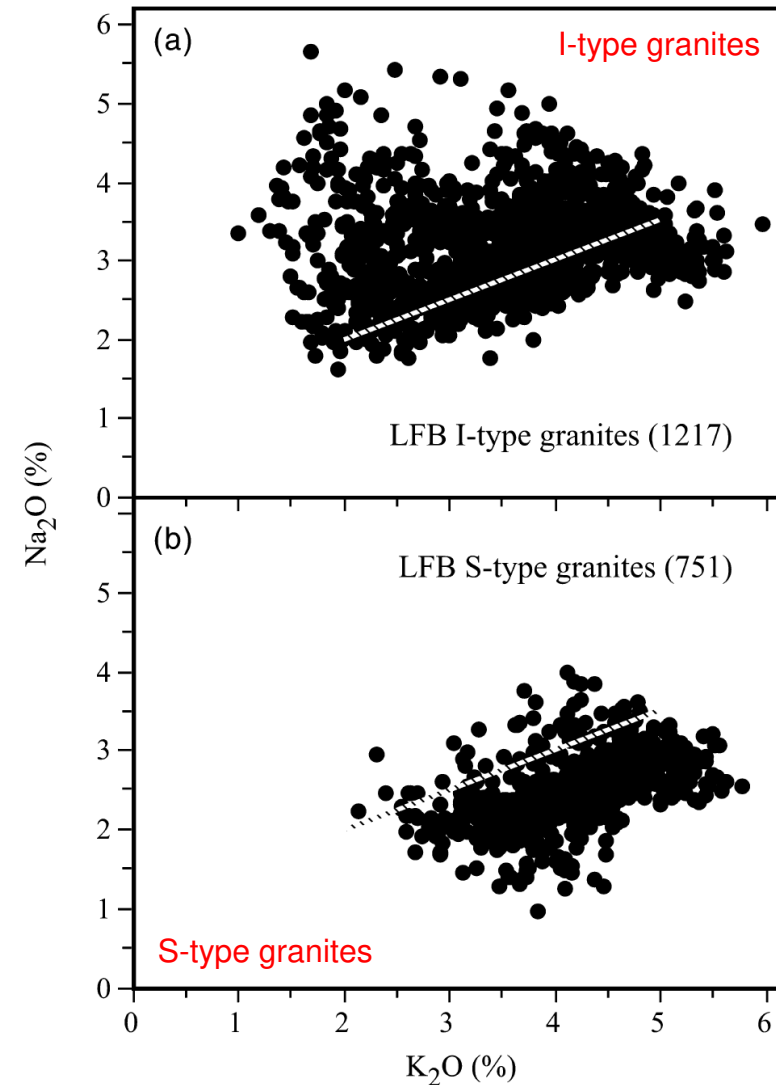
Major element characteristics based on ~2000 analyses on granites from the Lachlan foldbelt. Chappell & White (1974) first suggested predominantly meta-igneous sources for I-type and meta-sedimentary sources for S-type granites

## Magmatism in (continental) collision zones

### Origin and types of granites

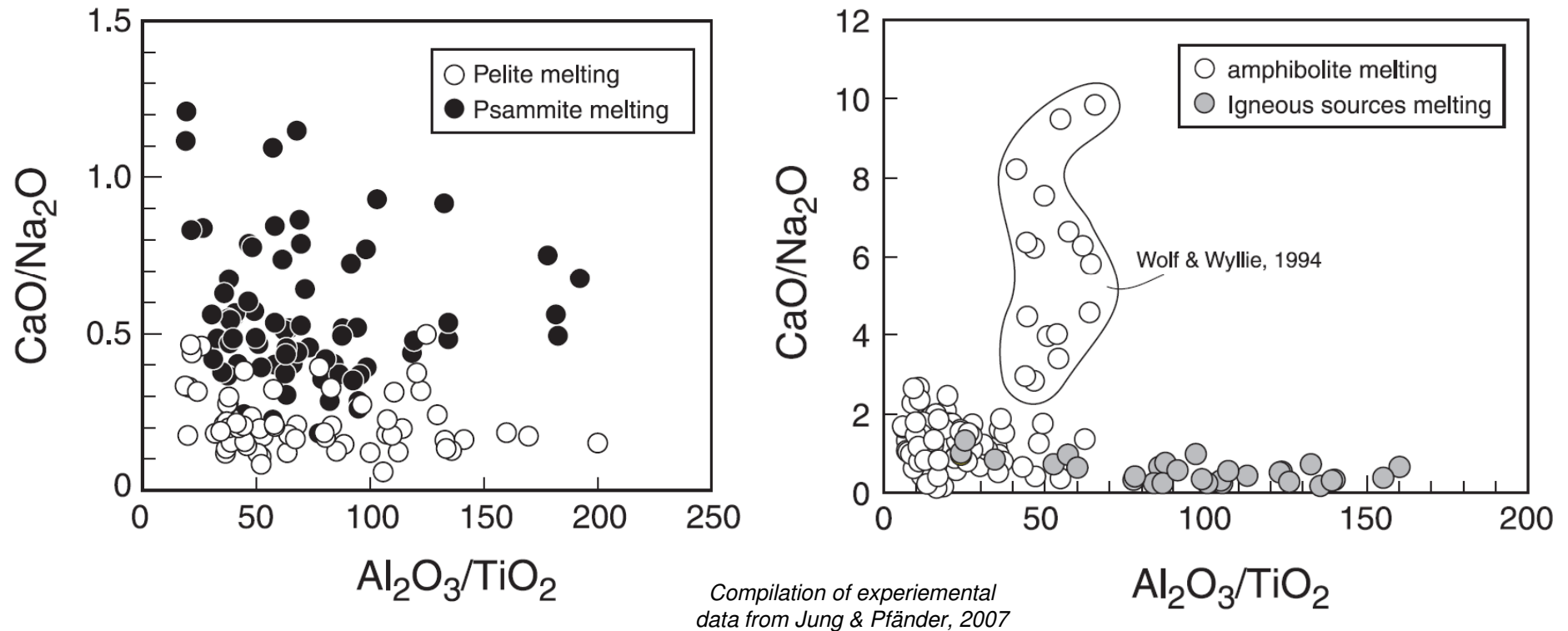


Major element characteristics  
of I- and S-type granites from  
the **Lachlan fold belt**, SE-  
Australia



## Magmatism in (continental) collision zones

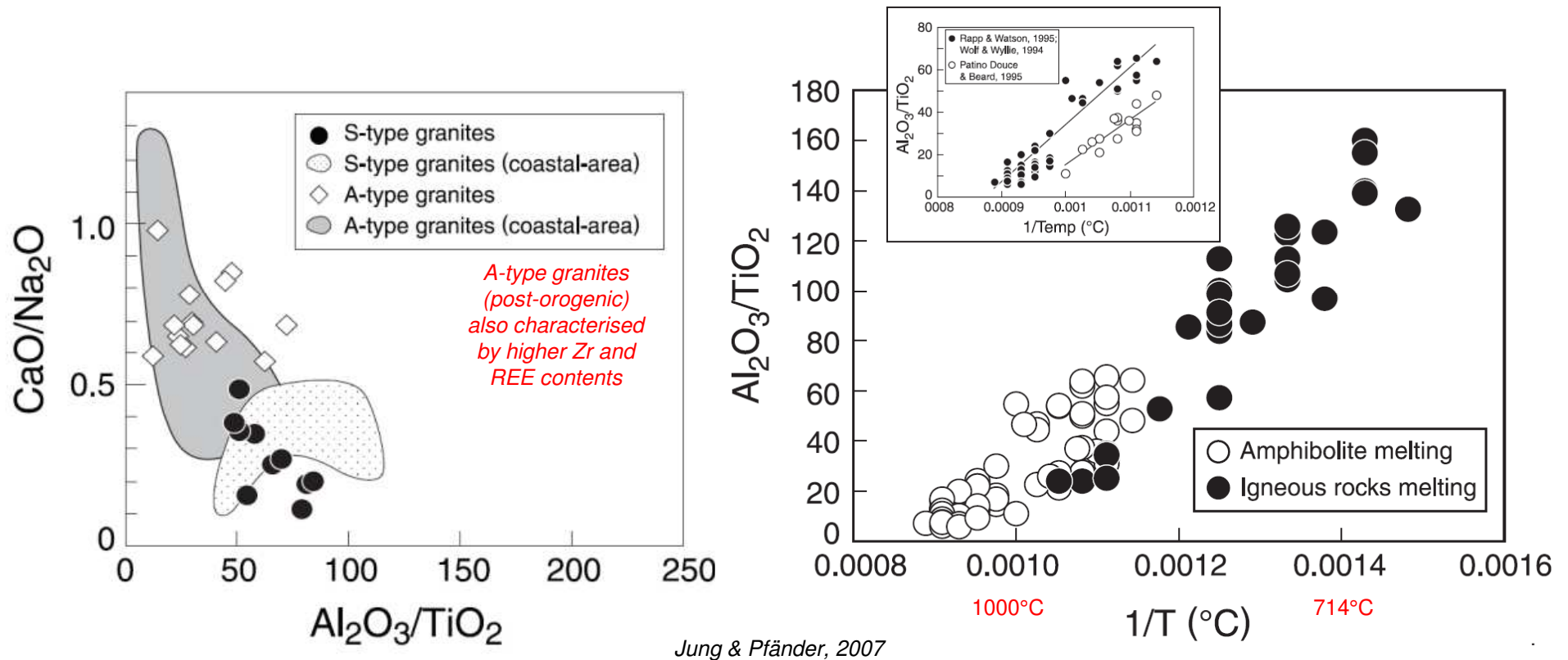
Origin and types of granites – experimental constraints



CaO/Na<sub>2</sub>O vs. Al<sub>2</sub>O<sub>3</sub>/TiO<sub>2</sub> of **granites** can be used as first-order constraints on **protolith composition** and **melting temperatures**

## Magmatism in (continental) collision zones

Origin and types of granites – experimental constraints



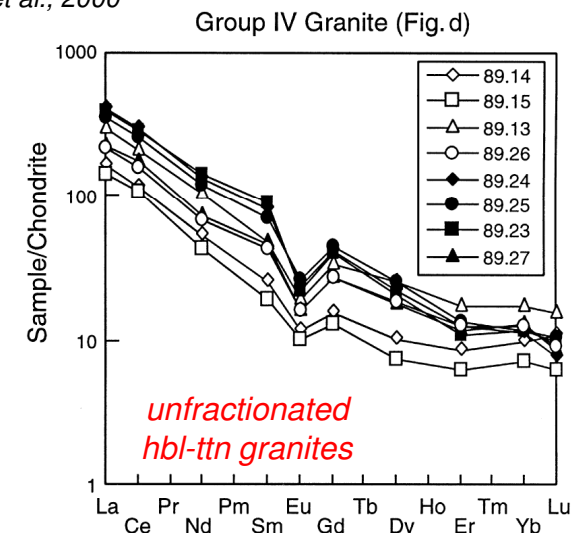
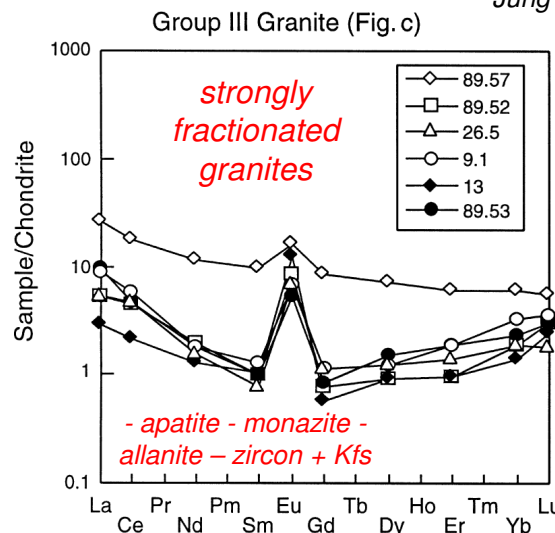
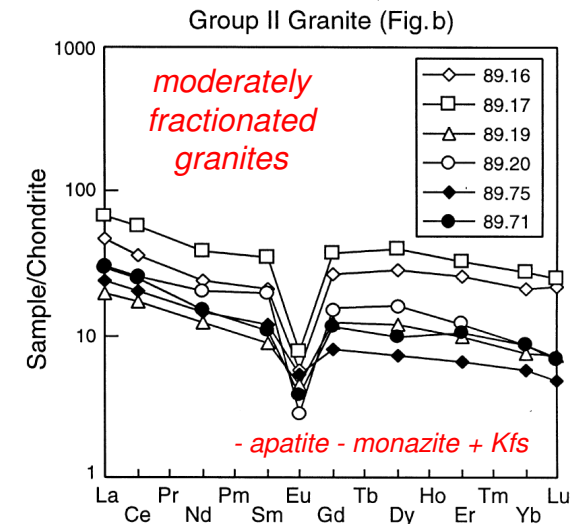
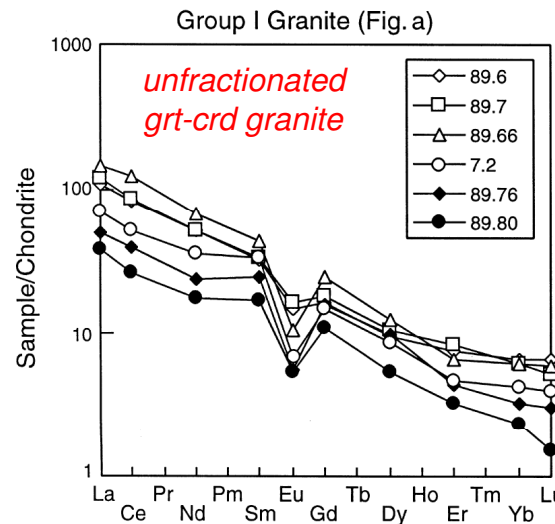
$\text{CaO}/\text{Na}_2\text{O}$  vs.  $\text{Al}_2\text{O}_3/\text{TiO}_2$  of **granites** can be used as first-order constraints on **protolith composition** and **melting temperatures**

## Magmatism in (continental) collision zones

Origin and types of granites - trace-element compositions

REE patterns of **A-** and **S-** type granites from the Damara orogen, Namibia.

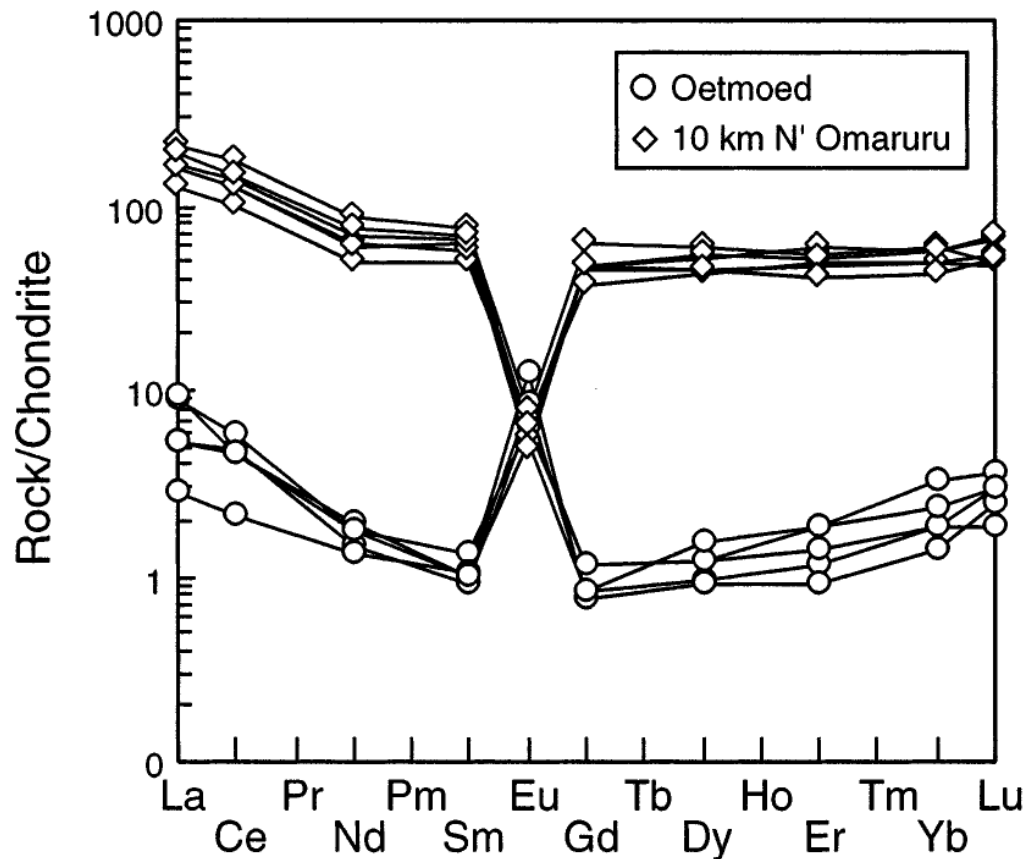
Comparison of unfractionated, moderately and strongly fractionated patterns. Note that **positive Eu-anomalies** require *Kfs* accumulation (wall-rock assimilation)



Jung et al., 2000

## Magmatism in (continental) collision zones

Origin and types of granites – trace-element compositions



REE patterns of **S-type granites** from the Damara orogen, Namibia.

Flat REE patterns (upper) indicate fractionation of predominantly monazite/apatite.

Low abundances and flat patterns (lower) indicate **strong fractionation of accessory phases** (monazite, apatite, zircon, xenotime, allanite) along with feldspar accumulation.

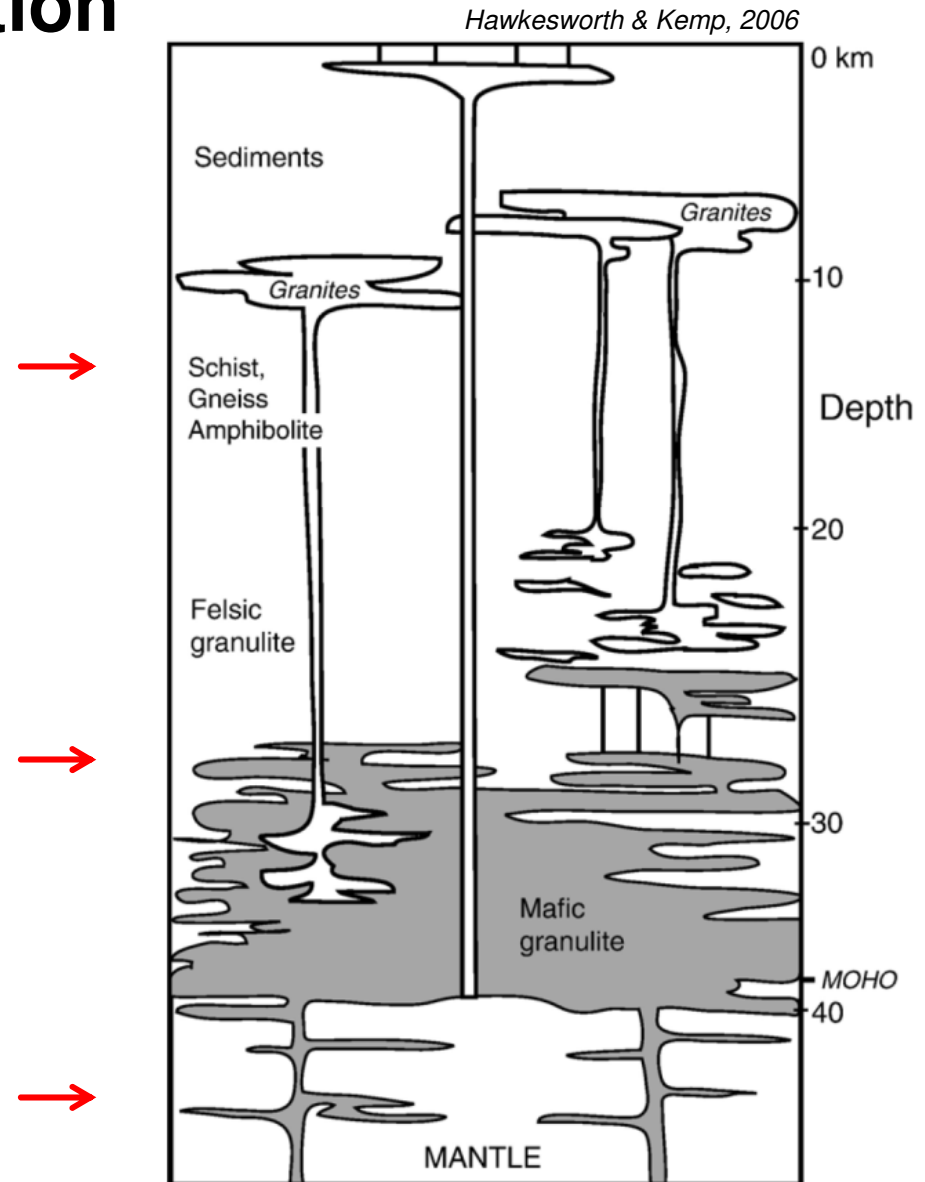
## Intracrustal differentiation

### Summary

Low- to medium-grade **metamorphism**, no anatexis partial melting, but some mafic melt injection from depth. **Intrusion of anatectic magmas** (granitic 'batholiths'), wall-rock assimilation and fractional crystallisation (AFC), contact metamorphism.

**High-grade metamorphism** due to **crustal thickening** and **partial melting** due to (1) **internal heating**, and (2) **heat addition** by magma injection.

Contribution from the **mantle** (continental arcs or mafic underplating): basaltic (alkaline, calc-alkaline, tholeiitic) magma plus **HEAT**.



**E N D**