



Penrose Conference

Plume IV:

Beyond the Plume Hypothesis

Testing the plume paradigm and alternatives

The Hotspot Handbook



August 25th – 29th, 2003
Hveragerdi, Iceland

<http://www.mantleplumes.org/>

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THE PLUME PARADIGM

The plume and the plate hypotheses are two of the most elegant ideas of global Earth science. There is a remarkable simplicity and symmetry between them and both are involved in the Standard Model of global geodynamics. Plumes were devised to explain features, such as age-progressive volcanic chains and continental breakup, which did not seem to be a part of rigid plate tectonics. Plate tectonics is the result of cooling of the Earth's surface and plumes are the result of heat transfer from the core to the base of the mantle. Pots on stoves have thermal boundary layers (TBLs) at the top and the bottom. Bottom heating and top cooling play comparable roles. Accidental perturbations in the thickness of either TBL will organize the flow in the fluid in the pot. The two modes of mantle geodynamics are usually treated separately. Ideal plumes are independent of plate tectonics and mantle convection.

The mantle is not a pot on a stove, however. Sphericity, pressure and continents break the symmetry. Material properties depend on temperature and pressure. The mantle is heated from within and contains, and loses, fossil heat. Stress plays a dominant role in plate tectonics. The concepts of rigid or elastic plates are fine for certain problems in global geodesy, plate kinematics and local bending but cannot apply as a general rule. The concept of strength has limited validity for objects as large as plates. Plate tectonics, as often described, is a rigid plate and kinematic theory. More generally, plate theory involves recycling, insulation, slab cooling and a template for mantle convection. Plates and slabs introduce chemical and thermal heterogeneity and structure into the mantle. Plates and plate boundaries are ephemeral. Long linear or arcuate volcanic features are related to stress and relative motions between plates. In the plume paradigm these types of features are attributed to high temperatures and relative motion of the plates over the mantle, rather than to stress or to incipient plate boundaries.

Plume models assume, as the normal condition, an isothermal subsolidus and homogeneous upper mantle which is either static or vigorously convecting (the "convecting mantle"). Variations in bathymetry and melt volume are attributed to core heat. Plume hypotheses are primarily fluid dynamic and thermal theories. Focusing, small-scale convection, fertility, ponding and passive upwellings associated with lithospheric architecture and extension can also create melting anomalies. These are athermal mechanisms. Can they be distinguished? Pressure is an essential parameter in convection and plume simulations but is not involved in laboratory and most computer modeling. Fluid dynamic modeling has not duplicated plate or plume tectonics. Plates are shaped and driven by Mother Nature and plumes are put in as initial singularities, or injections. Neither forms naturally. What is missing?

A good scientific hypothesis gets stronger as it is probed, questioned and tested. In particular, the assumptions behind paradigms must be constantly challenged. Paradoxes must be identified, for therein lie new ideas. Assumptions must be made to get any hypothesis started, but sometimes continued progress can only be made by looking for and dropping unfruitful assumptions. Rigidity, fixity, parallelism, homogeneity, steady-state, uniformitarianism, symmetry and incompressibility are some of the assumptions underlying current models.

Penrose meetings have been important in the development of both the plume and plate paradigms and offer an ideal forum for identifying the strengths and weaknesses of conventional wisdom. Perhaps a unifying, or at least, a self-consistent, theory will emerge.

The conveners, *Plume IV: Beyond the Plume hypothesis*

	Monday	Tuesday	Wednesday	Thursday	Friday
8:30 - 8:40	Opening address Foulger	Announcements Foulger	Announcements Foulger	Announcements Foulger	Announcements Foulger
8:40 - 9:25	I. What is a plume? What is a hotspot? Anderson Foulger	V. Opening ocean Foulger McHone	VIII. Petrology & temperature I Presnall Green		XIII. Hotspots, kinematics & dynamics Doglioni Winterer
9:25-10:10					
10:10-10:30	coffee	coffee	coffee		coffee
10:30-11:15	II. Seismology & hot spots Julian Montagne	VI. Continental Chalot-Prat Christiansen	IX. Petrology & temperature II Green Presnall		XIV. Cracks & tracks Smith Finn
11:15-12:00					
12:00 - 1:00	lunch	lunch	lunch	Fieldtrip: Hengill triple junction, Thingvellir, Geysir & Gullfoss	lunch
1:00-1:30	posters		posters		posters
1:30-2:15	III. The big, deep picture King Anderson		X. Geochem & Petrology I Natland Dick		XV. Large igneous provinces Sager Sheth
2:15-3:00					
3:00-3:30	tea	Fieldtrip: Reykjanes Peninsula	tea		tea
3:30-4:15	IV. Temperature, heat & magma Stein Gudfinnsson		XI. Geochem & Petrology II Dick Natland		XVI. Synthesis Anderson Wilson
4:15-5:00					
6:00-7:00	informal discussion		informal discussion		informal discussion
7:00-8:00	dinner	dinner	dinner	dinner	dinner
8:00 - 9:00	Lecture: Iceland Bjornsson	VII. Extraterrestrial Hamilton Jurdy		XII. Plate Tectonic End Games Flower Elston	

Session/Moderators/Key Questions

10-15 min keynote talks

Primary potential 2-3 min talkers

I. Overview: What is a plume? What is a hotspot?

Definitions, rules of the game, options, active vs. passive, lithosphere vs. asthenosphere, near-surface or deep, point sources of pollution vs. distributed, stress vs. temperature, cracks vs. tracks, global/regional vs. local, pressure effects, self-organization, central limit theorem, paradoxes, problems, predictions.

Moderators: Don L. Anderson, Gillian R. Foulger

- Rules of the game
- What is a plume?
- Initial & boundary conditions
- Assumptions
- What is plate tectonics (& implications)?
- Roles of T, P and composition.
- Sampling vs. reservoirs
- What drives what?

1. Don L. Anderson: Introduction

Introductions & overview; Where are we? What do we hope to learn? What are the issues, questions, paradoxes and bottom lines? By:

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| <ul style="list-style-type: none"> • <i>Enrico Bonatti</i> • <i>Henry Dick</i> • <i>Carlo Doglioni</i> • <i>Adam Dziewonski</i> • <i>Carol Finn</i> • <i>Martin Flower</i> • <i>David Green</i> • <i>Warren Hamilton</i> • <i>Anne Hofmeister</i> | <ul style="list-style-type: none"> • <i>Pino Guzzetta</i> • <i>Scott King</i> • <i>Greg McHone</i> • <i>Jim Natland</i> • <i>Dean Presnall</i> • <i>Carol Stein</i> • <i>Richard Walker</i> • <i>Marge Wilson</i> • <i>Jerry Winterer</i> |
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II. What does seismology say about hot spots?

Seismology is the highest-resolution technique for studying the deep structure of the Earth. This session reviews the current state of seismic imaging of the mantle and the question of what constraints seismology can place on the physical and chemical structures and processes beneath hot spots

Moderators: Bruce Julian, Jean-Paul Montagner

- What can seismology resolve? How small? How deep?
- What do seismic-wave speeds mean? Is red really hot and blue cold?
- What can we expect in the future? Bananas? Doughnuts? Flow fields?

1. **Adam Dziewonski:** Global seismic tomography: What we really can say and what we make up
2. **Jean-Paul Montagner:** Plume-lithosphere interactions: Cases of Afar (Africa), and Pacific hotspots

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| <ul style="list-style-type: none"> • <i>Don Anderson</i> • <i>Axel Bjornsson</i> • <i>Gillian Foulger</i> • <i>Anne Hofmeister</i> | <ul style="list-style-type: none"> • <i>Phillip Ilinger:</i> Whole Earth convection models • <i>Scott King</i> • <i>Seth Stein</i> |
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III. The big, deep picture

Much of our physical intuition regarding convection has come from Rayleigh-Benard type convection (constant physical properties, simplistic equation of state) or tank experiments (e.g., Griffiths and Campbell). In these experiments, there are no cratons, no plates, no mid-ocean ridges, no phase changes, no layering, no depth-dependent properties and no plate-scale flow. Have these simplifications led us down the wrong path? How does the near-surface (cratons, plates, ridges, edges, melting) impact the deep mantle? What do we learn from tomography? Why are there global plate reorganizations, and how do they work?

Moderators: Scott King, Don L. Anderson

- Does pressure reverse our intuition about convection?
- Is the mantle active or does it do what the plates tell it to?
- Is tomography a temperature or a petrology (composition, anisotropy, phase changes, flow) tool?
- What do we learn from global plate reorganizations?

1. **Scott King:** Plume Convection: What happens when you add all that icky stuff?

2. **Don Anderson**

3. **Anne Hofmeister**

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| <ul style="list-style-type: none"> • <i>Carlo Doglioni</i> • <i>Adam Dziewonski</i> • <i>Pino Guzzetta</i> • <i>Phillip Ilinger:</i> Whole Earth convection models | <ul style="list-style-type: none"> • <i>Jean-Paul Montagner</i> • <i>Alan Smith</i> • <i>Phil Wannamaker</i> |
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IV. Temperature, heat & magma

Geophysical constraints on models of hotspot/swell formation including heat flow, topography, and the assumptions about the volcanic age along island/seamount chains.

Moderators: Carol Stein, Gudmundur Gudfinnsson

- What does heat flow tell us about hot spots/plumes?
- What do depths tell us about hot spots/plumes?
- What does volcanic age progression along island/seamount chains tell us about the interaction of the upwelling hot mantle material and the lithosphere?

1. **Carol Stein:** Spots yes, hot barely or not
2. **John M. O'Connor:** What can long-lived seamount chains reveal about the origin of hotspots?
3. **Enrico Bonatti:** Mantle thermal structure below ridges: space and temporal variations

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| <ul style="list-style-type: none"> • <i>Don Anderson</i> • <i>Françoise Chalot-Prat</i> • <i>Giuseppe Guzzetta</i> • <i>Warren Hamilton</i> • <i>Anne M. Hofmeister</i> • <i>Phillip Ithinger</i> • <i>Vlad Manea</i> • <i>John O'Connor</i> | <ul style="list-style-type: none"> • <i>Hannah L. Redmond</i> • <i>Suzanne Smrekar</i> • <i>Seth Stein</i> • <i>Ellen Stofan</i> • <i>Peter Vogt</i> • <i>Phil Wannamaker</i> • <i>Dayanthie Weeraratne</i> |
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V. Opening of an ocean

The opening of the Atlantic ocean was accompanied by intense volcanism along much of the new seaboard. This persisted at several locations giving rise to a chain of volcanic anomalies along the mid-Atlantic ridge. These are traditionally attributed to localised high temperatures, but what evidence is there for this, and can an athermal model stand?

Moderators: Gillian R. Foulger, Greg McHone

- Can volcanism in the Atlantic ocean be explained by variations in mantle fertility and deformation?
- What are the implications of this view?
- What are the main problems with this theory?
- How can they be addressed?

1. **Gillian R. Foulger:** A shallow model for north Atlantic volcanism
2. **Greg McHone:** Volcanic features of the central Atlantic ocean: tectonic and magmatic models
3. **Marge Wilson:** Understanding the 135 Myr record of magmatism in the South Atlantic: Plumes, plate tectonics and propagating fractures

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| <ul style="list-style-type: none"> • <i>Enrico Bonatti</i> • <i>Axel Bjornsson</i> • <i>Don DePaolo</i> • <i>Henry Dick</i> • <i>Godfrey Fitton</i> • <i>Martin Flower</i> • <i>Bjarni Gautason</i> • <i>Warren Hamilton</i> | <ul style="list-style-type: none"> • <i>Sveinn P. Jakobsson</i> • <i>Jose Mangas</i> • <i>Hetu Sheth</i> • <i>Olgeir Sigmarsson</i> • <i>Alan Smith</i> • <i>Reidar Trønnes</i> • <i>Peter Vogt</i> |
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VI. Continental volcanism & lithospheric tectonics

The compositions and tectonic associations of continental intraplate magmatism are highly varied, and these factors are not always clearly linked. Both continental flood basalts and anorogenic magmatic suites with major silicic components can reflect significant crustal interactions but are nevertheless generated and sustained by melting in the upper mantle. The alkaline mafic magmas common in intraplate settings are often considered to be "hotspot-related", but they may erupt synchronously with calc-alkaline magmas – either syn-subduction or post-collisional. Rift-related continental mafic magmas may be either calc-alkaline or alkaline, depending at least in part on whether rifting occurred a few thousands to millions of years or hundreds of millions of years after continental-plate collision.

Are these and similar problems better explained by the mantle-plume models commonly invoked for them? Alternatively, do many of these magmas reflect varied source compositions, whereas causes of partial melting and magma emplacement depend on the direct interaction of lithospheric tectonics with deeper upper-mantle processes?

Moderators: Françoise Chalot-Prat, Bob Christiansen

- Are continental mafic magmas generated in residual or metasomatized mantle?
- Does mantle fertility reflect the recycling of oceanic or continental crust, or even of previously metasomatized mantle? When do such changes occur relative to magma genesis (importance of the lithospheric plate story before eruptions)?
- What are the causes of subcontinental mantle melting? Are there distinct roles for deep-mantle plumes, convective systems restricted to the upper-mantle, and lithospheric tectonics to both promote melting of fertile mantle and provide magma conduits to the surface?

1. **Angelo Pecerrillo:** Ultrapotassic magmatism: Shallow mantle or plume-related process? The case of central Italy
2. **Hetu Sheth:** The Deccan beyond the plume hypothesis
3. **Bob Christiansen:** Structural control and plate-tectonic origin of the Yellowstone melting anomaly

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| <ul style="list-style-type: none"> • <i>Richard Chamberlin</i> • <i>Corrado Cigolini</i> • <i>Wolf Elston</i> • <i>Don DePaolo</i> • <i>Zuzana Fekiacova</i> • <i>Carol Finn:</i> Cenozoic alkaline magmatism in west Antarctica, east Australia and New Zealand. • <i>Martin Flower</i> • <i>Warren Hamilton</i> | <ul style="list-style-type: none"> • <i>Greg Huffman</i> • <i>Vlad Manea:</i> Mantle wedge flow and thermal models for the central Mexican subduction zone • <i>Greg McHone</i> • <i>Alan Smith</i> • <i>Phil Wannamaker</i> • <i>Marge Wilson</i> • <i>Don Wright</i> |
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VII. Extraterrestrial

Evidence for plumes on other planets will be considered: Venus and Mars. Venus hosts numerous "coronae", circular features, as well as uplifted regions that may or may not result from plumes on that planet. The huge Tharsis uplift - the largest known - on Mars may have been caused by a single plume active through much of the planet's history. Some features attributed to plumes on Earth could be the result of impacts.

Moderators: Warren Hamilton, Donna Jurdy

- What is the evidence for plumes on Venus and how does it differ from that on Earth?
- What is the evidence for plumes, or possibly a single plume, on Mars?
- Did impacts on Earth cause plumes?

5-min talks:

1. **Hannah Redmond:** Tharsis Rise, Mars, result of a long-lived plume
2. **Suzanne Smrekar:** Upwelling at different scales on Venus
3. **Donna Jurdy:** Coronae as evidence of active upwelling on Venus
4. **Warren Hamilton:** An alternative Venus: Plume-free planet
5. **Wolfgang Elston:** Impacts as a cause for plumes, Bushveld, as an example

- *Ellen Stofan*

- *Don Wright*

VIII & IX What does petrology tell us about potential temperatures?

Hot plumes imply elevated potential temperatures. For about the last 15 years, it has been commonly thought that petrologists were finally beginning to develop an ability to determine this difficult parameter. However, in the last two years, major differences of opinion have developed that have reopened this issue at a very fundamental level. This session will examine the current status of this subject and discuss future directions.

Moderators: Dean Presnall, David Green

- High vs. low vs. strongly variable potential temperatures
- Picrites vs. picrites
- MORB vs. "hot spot" basalt chemistry
- Can major-elements of basalts constrain potential temperatures?
- Mantle heterogeneity vs. variable potential temperature
- Mantle heterogeneity vs. basalt chemistry
- Volatiles and melting curves
- What are the "primary" magmas at "hot spots" vs ridges?
- Where is the bottom of the seismic low-velocity zone?
- Long, short, or variable melting columns?

1. **Dean C. Presnall:** Phase equilibrium/seismic constraints on potential temperatures
2. **David H. Green:** Potential temperatures and primary magmas in MOR setting and comparison with Hawaii
3. **Gudmundur Gudfinnsson:** Contrasting origins of the most magnesian glasses from Iceland and Hawaii

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| <ul style="list-style-type: none"> • <i>Don Anderson</i> • <i>Enrico Bonatti</i> • <i>Corrado Cigolini</i> • <i>Marc Davies</i> • <i>Henry Dick</i> • <i>Adam Dziewonski</i> • <i>Godfrey Fitton</i> • <i>Dennis Geist</i> | <ul style="list-style-type: none"> • <i>Karen Harpp</i> • <i>Kevin Johnson</i> • <i>Jose Mangas</i> • <i>Jim Natland</i> • <i>D. Gopala Rao</i> • <i>Hetu Sheth</i> • <i>Reidar Trønnes</i> |
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X & XI Geochemistry & Petrology

The plate-tectonic cycle imparts heterogeneity to the mantle. Plume theory depends strongly on how melt processes sample a heterogeneous mantle. Geochemistry now suggests that most ocean island basalt (OIB) heterogeneity results from sampling of material once in the Earth's crust but that is now in the mantle, having entered there via subduction. The main question posed by OIB geochemistry is how far into the mantle these materials were carried before becoming involved again in volcanism at ridges, LIPs, islands and seamounts? Did they reach the core-mantle boundary, and arise again in plumes, or did they become trapped in the upper mantle for long periods of time before being tapped? How can petrology and geochemistry tell?

Moderators: Jim Natland, Henry Dick

- What are OIBs?
- Is any magma primary?
- Is the mantle a plum pudding? Distribution of enriched components – the statistical upper mantle assemblage
- What are the possible effects of bulk heterogeneity of the mantle on the compositions of basalt?
- Can melt-extraction processes by themselves produce heterogeneity?
- Are any mantle reservoirs truly well mixed? Fertile versus barren peridotite
- What are the roles of “recycled” ocean crust, eclogite and pyroxenite in mantle sources of basalt?
- Can we move beyond alphabet soup? The relationship of trace-elements and isotopes to possible bulk heterogeneity of the mantle
- What are indicators of a very deep mantle source?
- What is the significance of spatial and temporal geochemical variability on islands and island chains?
- Do komatiites indicate plume heads, or something else?
- What is helium trying to tell us?

1. **Jim Natland:** Opening gambit: A perspective on mantle heterogeneity
2. **Henry Dick:** Abyssal peridotites and tholeiites
3. **Don DePaolo:** Geochemical structure of the Hawaiian plume: Results from the Hawaii Scientific Drilling Project

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| <ul style="list-style-type: none"> • <i>Don Anderson</i> • <i>Enrico Bonatti</i> • <i>Françoise Chalot-Prat:</i> The link between magma genesis and lithospheric tectonics during ocean spreading • <i>Corrado Cigolini</i> • <i>Mark Davies</i> • <i>Don DePaolo</i> • <i>Godfrey Fitton</i> • <i>Gillian Foulger</i> • <i>Dennis Geist</i> | <ul style="list-style-type: none"> • <i>David Green</i> • <i>Martin Flower</i> • <i>Warren Hamilton</i> • <i>Phillip Ihinger</i> • <i>Sveinn P. Jakobsson</i> • <i>Dean Presnall</i> • <i>Hetu Sheth</i> • <i>Olgeir Sigmarsson</i> • <i>Alan Smith</i> • <i>Richard Walker</i> • <i>Marjorie Wilson</i> |
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XII. Plate Tectonic End Games

While most scientists assume plate mobility is driven by mantle flow, "slab pull", and "lithosphere push" forces, a paradigm shift may be needed given the failure of "jelly sandwich" lithosphere models and increasing indications of plate-induced mantle flow. Plate tectonic "end games" may bear critically on several phenomena - Large Igneous Provinces, back-arc basins, forearcs and ophiolites, and oceanic hotspot trails. Collision-related seafloor spreading changes, accompanied or preceded by new subduction events, are often followed by arc-trench rollback. When subduction initiation precedes a collision, rollback is probably mantle-driven rather than triggered by plate kinematics. Global-scale responses, e.g. to the Africa-Eurasian collision, may also involve far-field mantle flow perturbations. These observations highlight the paradoxes of hot plume models. Links to propagating cracks, mantle thermal and compositional heterogeneities, and shallow perturbations suggest mantle flow is both the cause and effect of plate motions. Upper-mantle anisotropy and variations in asthenospheric T_p support numerical models that can explain continental volcanism, escape tectonics, marginal basin opening, and the genesis of ophiolites as responses to small-scale, plate-induced convection. Thus while global synchronism is an important aspect of plate tectonics, resisting or dissipative stresses, which control the spatial-temporal distribution of volcanic arcs, marginal basins, and mountain belts are subject to rapid changes.

Moderators: Martin Flower, Wolf Elston

- The formation and propagation of back-arc basins - "slab pull", mantle flow, or escape tectonics?
- Large Igneous Provinces - are these linked to cratons and mobile belts?
- Ophiolite genesis at new, hot, plate boundaries - do these reflect lithosphere heterogeneities or collision-related mantle flow fronts?
- Fertile mantle at collision sutures - does this portend large-fraction melting when the Wilson cycle resumes?

1. **Martin Flower:** Mantle melting, stress dissipation, and the Wilson cycle
2. **Wolfgang Elston:** The unique 2.06 Ga Bushveld Complex, South Africa: Result of an impact-induced plume?

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| • Don Anderson | • Warren Hamilton |
| • Tiffany Barry | • Vlad Manea |
| • Franciose Chalot-Prat | • Jose Mangas |
| • Richard Chamberlain | • Greg McHone |
| • Bob Christiansen | • Jean-Paul Montagner |
| • Corrado Cigolini | • Jim Natland |
| • Marc Davies | • Angelo Peccerillo |
| • Henry Dick | • Hetu Sheth |
| • Carlo Doglioni | • Alan Smith |
| • Zuzana Fekiacova | • Marge Wilson |
| • Gillian Foulger | |
| • Fred Frey | |

XIII. Hotspots vs. Plate Kinematics and Dynamics

Age progressions, reference frames, westward drift of the lithosphere, mantle and plate kinematics, lithospheric fabric and stress

Moderators: Carlo Doglioni, Jerry Winterer

- What does the age progression of Pacific hotspots tell us about motions & shears in the mantle?
- Do plates & slabs drive themselves or is an independent mantle convective source, or something else required?
- What kinematic constraints are there on depths of hotspot and MORB sources?
- Is the westward drift of the lithosphere global or is it only a mean value? What is its origin?
- Is stress, water content, or mantle temperature the dominant parameter in localizing hotspot volcanism?
- What can we really argue about mantle kinematics?
- What are rates of ridge and trench migration, and the minimum relative rates of plates? Do these differ from hotspot motions?
- Can/do plate motions and local change rapidly and often?

1. **Phillip D. Ihinger:** Plume magmatism and mantle convection: revising the standard
2. **Carlo Doglioni:** On the westward drift of the lithosphere
3. **James H. Natland:** On changing stress during Pacific plate kinematic evolution

- *Peter R. Vogt:* Sea-floor basement morphology: Distinguishing hotspot effects from plate tectonic effects - Examples from Iceland and the Azores
- *Dayanthie Weeraratne:* An alternative model for the origin of non-hotspot intraplate volcanism in the Pacific
- *Brian Pope:* Is hot spot magmatism, like Hawaii, coming from shallow mantle?
- *Gillian Foulger:* On the apparent eastward migration of the spreading ridge in Iceland
- *Martin Flower:* Collision-induced mantle flow during Tethyan closure: a link between magmatism, lithosphere 'escape', and arc-trench rollback?
- *Phillip Ihinger:* Spatial and temporal geochemical variations along alleged hot-spot tracks
- *Hetu C. Sheth:* The Deccan beyond the plume hypothesis
- *Alan Smith:* The fate of subducted oceanic crust and the sources of intraplate volcanism

XIV. Cracks & tracks

The role of lithospheric architecture (thickness and fabric - cracks, boundaries, sutures) in intra-plate magmatism Pacific, dikes, leaky & incipient plate boundaries, island chains.

Moderators: Alan Smith, Carol Finn

- What are the roles of lithospheric architecture and stress on magmatism?
- What are the links between lithospheric (architecture and stress) and upper and lower mantle processes (that is, dynamic processes, such as upper mantle hot/warm spots, small scale convection, detachment of subducting slabs in the upper and lower mantle, cooling by subduction, chemical modification by subduction, melting at ridges, and plumes, and thermal boundary layers in the lower mantle, etc.) that result in magmatism?
- What are the relations between lithospheric architecture/stress and mantle temperature and chemistry (in particular origin and location of volatiles (CO₂, H₂O + recycled crust, eclogite) that lower melting temp.) and ponding/underplating that allows or increases volumes or rates of magmatism?
- What are the orientation, magnitude and sign of stress required to open pre-existing zones of weakness or break virgin rocks that permit magmatism?
- What are the links between lithospheric fabric and stress and age-progressive volcanism?
- Are there temporal links between the onset, termination, and longevity of regional mid-plate volcanism and plate tectonic events such as regional and global plate reorganizations and conjectured slab detachments?

1. **Alan Smith:** The Regular Distribution of Intraplate Volcanism in the Pacific Basin
2. **Erin Beutel:** Lithospheric stress state responsible for hotspots at ridge-transform intersections?
3. **Carol Finn:** Definition of a Cenozoic alkaline magmatic province in the southwest Pacific mantle domain (W. Antarctica, E. Australia and New Zealand) without rift or plume origin.

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| <ul style="list-style-type: none"> • <i>Enrico Bonatti</i> • <i>Francoise Chalot-Prat</i> • <i>Karen S. Harpp</i> • <i>Phillip Ihinger</i> • <i>Jose Mangas</i> | <ul style="list-style-type: none"> • <i>James Natland</i> • <i>John O'Connor</i> • <i>Dayanthie Weeraratne</i> • <i>Jerry Winterer</i> |
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XV. Geodynamic origin of large-volume basaltic provinces & flood basalts

Over the last 10-15 years the plume head hypothesis has become the theory of choice to explain the formation of large LIPs, such as ocean plateaus and continental flood basalts. Data have now been gathered from a number of LIPs and compared with this hypothesis. In this session, we will discuss how well the plume head hypothesis has held up to this scrutiny. We will examine alternate hypotheses that may do a better job of explaining large LIPs. Finally, we will compile the matrix of hypotheses and critical observations needed to test these hypotheses as a guide to future LIP research.

Moderators: Will Sager, Hetu Sheth

- Is the evidence from LIPs consistent with the plume head model?
- How well do non-plume mechanisms work for LIPs?

1. **Fred Frey:** The Kerguelen plume: what we have learned from ~120 Myr of volcanism
2. **Godfrey Fitton:** A plume origin for the OJP?
3. **Will Sager:** Tectonic evolution of the Shatsky Rise: a plateau formed by a plume head or not?

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| <ul style="list-style-type: none"> • <i>Richard Chamberlin</i> • <i>Wolf Elston:</i> The Bushveld enigma: A catastrophe-triggered complex LIP on a tectonically stable platform. • <i>Fred Frey</i> | <ul style="list-style-type: none"> • <i>David Green</i> • <i>Warren Hamilton</i> • <i>Greg McHone</i> • <i>Dean Presnall</i> • <i>Alan Smith</i> • <i>Marge Wilson</i> |
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XVI. Synthesis

Moderators: Don L. Anderson, Marge Wilson

- *Everybody*

Organising Committee

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Post-Conference Book

The GSA have agreed to edit a post-Conference book of papers arising. All Conference participants are invited to submit papers for this book, and proposals for additional relevant contributions from non-participants are welcome and will be considered by the Editors. The Editors will be Gillian R. Foulger, James H. Natland, Dean C. Presnall and Don L. Anderson. To date over 30 conference participants have pledged papers.

The target timetable for production and processing of manuscripts is:

Deadline for submissions	January 15th 2004
Reviews returned	April 15th 2004
Revised ms returned to editors	May 30th 2004
Submission to GSA	July 30th 2004

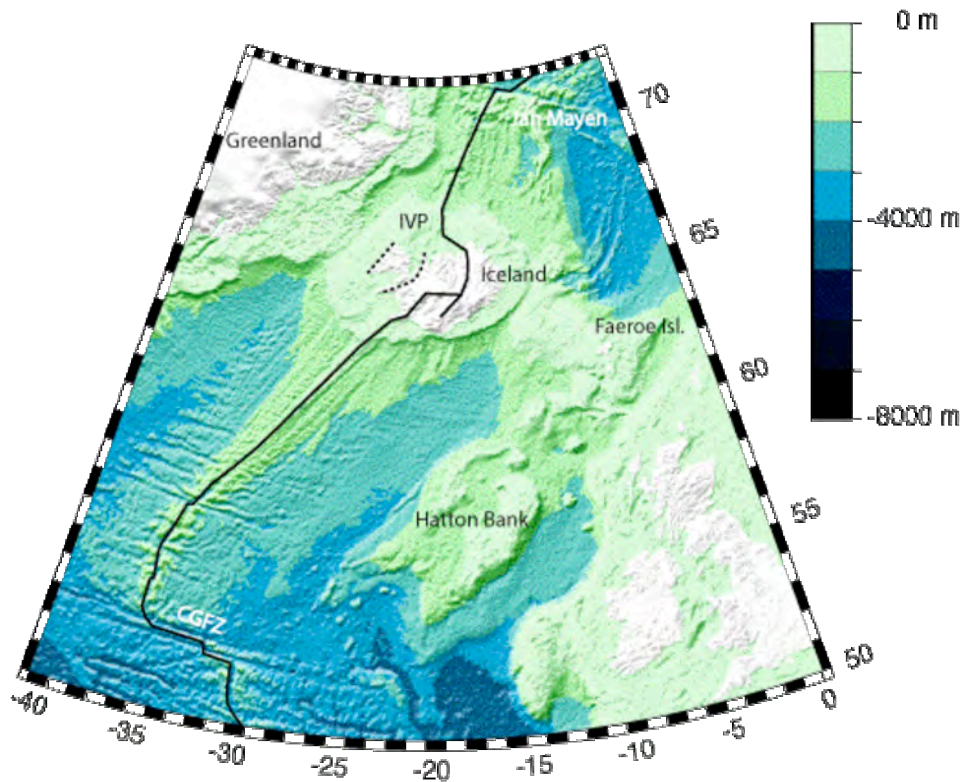
Papers will be reviewed by two reviewers, and the Editors will adjudicate in difficult cases. Authors are encouraged to submit a list of potential reviewers with their manuscripts, including candidates who are not authors of any paper in the volume.

Electronic submission to Gillian R. Foulger at g.r.foulger@durham.ac.uk is preferred. PDF files, Word documents with figures either embedded or separately, and most other formats can be handled. Submissions not in PDF will be translated to this format for distribution to reviewers.

Each paper/chapter should be at or above the standards for papers in GSA Bulletin. More information about GSA books may be found at <http://www.geosociety.org/pubs/bookguid.htm>

A brief geological tour of Iceland

Iceland lies on the mid-Atlantic ridge between $\sim 63^\circ\text{N}$ and 65°N , where the full spreading rate is ~ 1.9 cm/year. It is flanked by ridges of thick crust that extend to the adjacent continental blocks of Greenland and the Faeroe islands. These “aseismic ridges”, and Iceland itself, are underlain by crust ~ 30 km thick, which indicates that magmatism has been approximately three times that along the mid-ocean ridges to the north and south ever since the north Atlantic opened at ~ 54 Ma. This, and the ocean-island-basalt (OIB) geochemistry of the Iceland region, have traditionally been assumed to result from a mantle plume, but alternative, shallow-sourced models are currently being scrutinised [Foulger *et al.*, 2003].

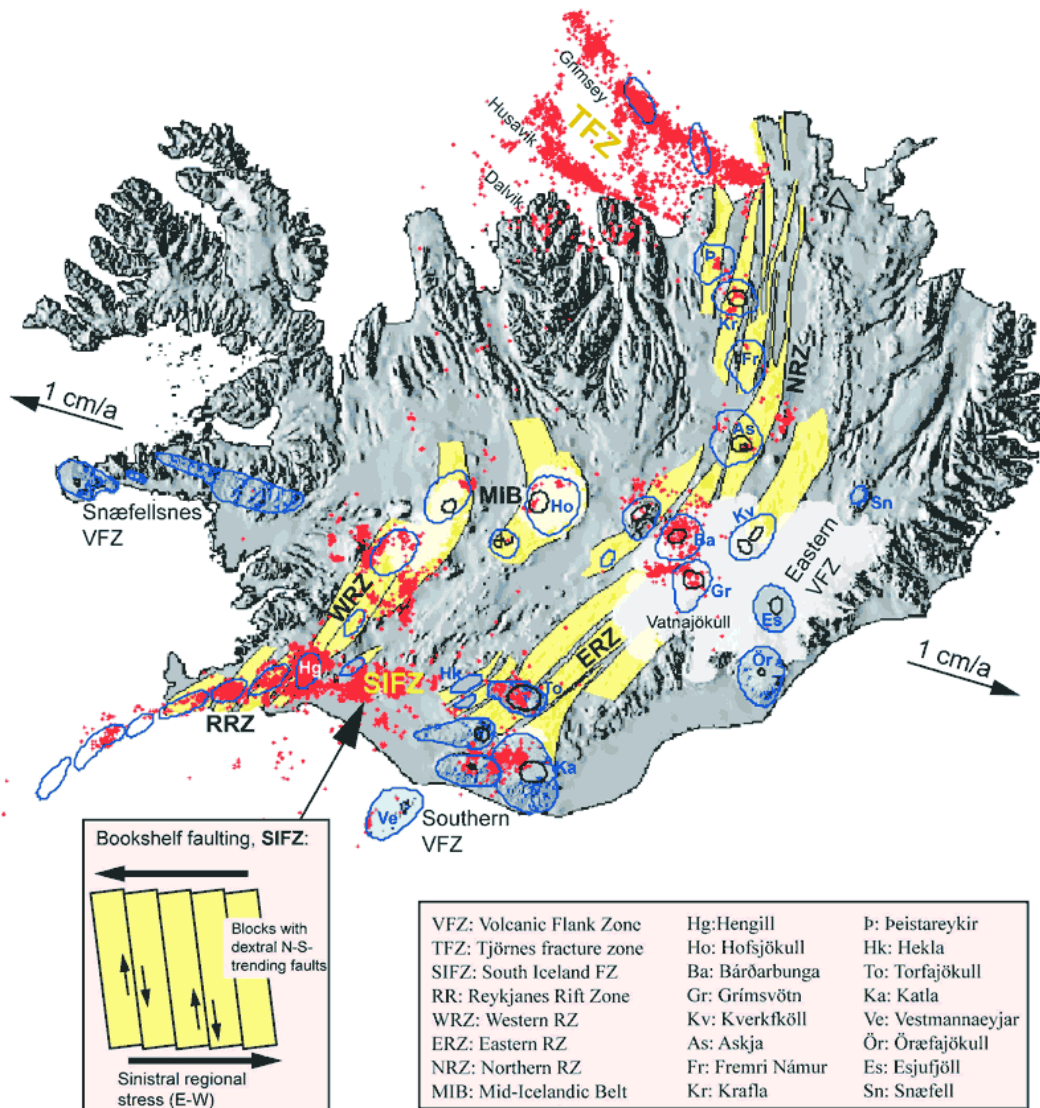


Bathymetry of the area around Iceland.

The initial opening of the north Atlantic at ~ 54 Ma was preceded by Paleocene continental magmatism at 62 – 58 Ma, which produced large volumes of basalts and associated intrusives on Baffin Island, Greenland and the British Isles. Continental breakup, plate separation and incipient ocean crust formation were accompanied by the formation of thick, seaward-dipping reflectors (SDR) along the east Greenland and northwest European margins. Plate boundary configuration in the Iceland region and to the north was subsequently complicated, and involved several migrations of the spreading ridge and spreading about a parallel pair of ridges for much of the time [Foulger, 2003]. Such a spreading style is currently ongoing in south Iceland today. The Iceland region meets the definition of a diffuse oceanic plate boundary [Zatman *et al.*, 2001]. In contrast, spreading along the Reykjanes ridge to the south has had a relatively simple history.

Iceland is currently traversed by several 40 – 50 km wide rift zones, the Reykjanes, Western, Eastern and Northern rift zones. These comprise *en-echelon* arrays of spreading segments, most containing a central volcano that may have one or more calderas, acid and intermediate rocks, a high-temperature geothermal area, and a fissure swarm 5 – 15 km wide and up to 200 km long. The Western and Eastern rift zones are subparallel in south Iceland and spreading is distributed across both. A very broad zone of rifting, comprising at least 6 subparallel spreading centers forms the Mid-Iceland Belt that traverses central Iceland. North of this spreading is taken up along only one rift zone – the Northern Rift Zone. The situation in the north may be unusual over the last 26 Myr, as two subparallel rift zones have existed there for much of this time. Extinct rift zones currently lie in north and northwest Iceland.

In addition to the active rift zones, three non-rifting volcanic flank zones (VFZ) are recognised, the Snæfellsnes, Southern and Eastern VFZs. In these, most of the volcanic centers lack well-developed fissure swarms, geothermal activity is generally lower, and volcanism is not accompanied by substantial crustal widening.



Tectonic map of Iceland. Red dots show epicenters of 25,000 earthquakes from 1994 – 2000. Fissure swarms are in yellow. Volcanic centers and calderas are outlined.

There are two complex fracture zones in Iceland that take up transform motion. One lies in south Iceland and connects the southern ends of the Western and Eastern Rift zones (i.e., the Hengill triple junction and Hekla volcano). This zone experiences earthquakes up to magnitude ~ 7.5 on short, north-south orientated faults. Seismic sequences typically occur about once per century, start in the east and propagate west, decreasing in magnitude towards the west. The zone is thought to deform in a bookshelf faulting manner on right-lateral faults, such that the whole zone deforms in a left-lateral way as required by large-scale plate motion. Two magnitude 6.6 earthquakes occurred there in the year 2000, and this zone is thus under careful surveillance. The large earthquakes in the year 2000 triggered substantial earthquake activity in the Reykjanes Peninsula. North of Iceland the Northern Rift Zone is connected to the Kolbeinsey Ridge by the offshore Tjornes Fracture Zone which takes up right-lateral transform motion on three subparallel oblique faults.

Iceland contains many remarkable central volcanoes. Hekla (lit. "Hood") erupts frequently and the SiO_2 contents of the eruptives are related to the length of repose since the previous eruption. Torfajokull in the Eastern Rift Zone is exceptionally rich in intermediate rocks and erupts mixed basalt/rhyolite lavas. Katla is remarkable for its seasonal earthquake activity, which occurs primarily in the winter. The cluster of central volcanoes beneath NW Vatnajokull, which include Bardarbunga, Grimsvotn and Kverkfjoll, represent the greatest volcanic activity in Iceland. Kverkfjoll lies on the northern margin of Vatnajokull and has melted remarkable ice caves in the glacier. Grimsvotn contains a subglacial caldera lake which continually fills as a result of melting of the icecap at its base by geothermal heat. Periodically the lake level rises high enough to lift the icecap from bedrock locally and the lake drains, with water flowing beneath the icecap to the south and forming a glacial burst ("jokulhlaup") onto the sandy plains south of Vatnajokull. The subglacial volcano Gjalp, which recently erupted and melted a hole in the icecap, causing an exceptionally large, damaging glacial burst, lies just north of Grimsvotn.

Large earthquakes in Iceland are mostly associated with the two fracture zones, but large numbers of earthquakes also occur within the rift zones [Einarsson, 1991]. On a day-to-day basis, these are of small magnitude and mostly associated with the geothermal areas. They probably result from thermal contraction cracking in the cooling intrusions that comprise the geothermal heat sources. Periodically swarms of earthquakes are associated with spreading episodes along the rift zones.

The mechanism of spreading in Iceland was revealed by an episode in the Krafla spreading segment in the Northern Rift Zone, 1975-1985 [Björnsson, 1985]. The Krafla central volcano became activated and a magma chamber beneath it inflated at a rate of $\sim 5 \text{ m}^3/\text{s}$ for a decade. Periodically the magma chamber failed and magma escaped forming dikes to the north and south along the fissure swarm. These dike injections were accompanied by swarms of earthquakes, but once the magma reached the surface and eruption began, earthquake activity greatly declined. Total lateral crustal extension of $\sim 10 \text{ m}$ occurred, but no earthquakes were larger than magnitude ~ 4.5 . This episode demonstrated that such spreading events could occur along the marine spreading plate boundary undetected by land seismic stations. Following this episode, the most remarkable post-tectonic anelastic deformation field ever observed in the world was measured with GPS. Extension across the fissure zone was several times the average plate rate for about a decade. The viscosity of the asthenosphere under Iceland is calculated to be $\sim 10^{19} \text{ Pa s}$ [Foulger et al., 1992]. This result was confirmed by study of isostatic rebound of the Vatnajokull icecap during the 20th century [Sigmundsson and Einarsson, 1992].

The volcanic systems of the rift zones mainly produce tholeiitic basalts. The major products of the flank volcanic zones are mildly alkaline and transitional (tholeiitic to alkaline) basalts [Saemundsson, 1979].

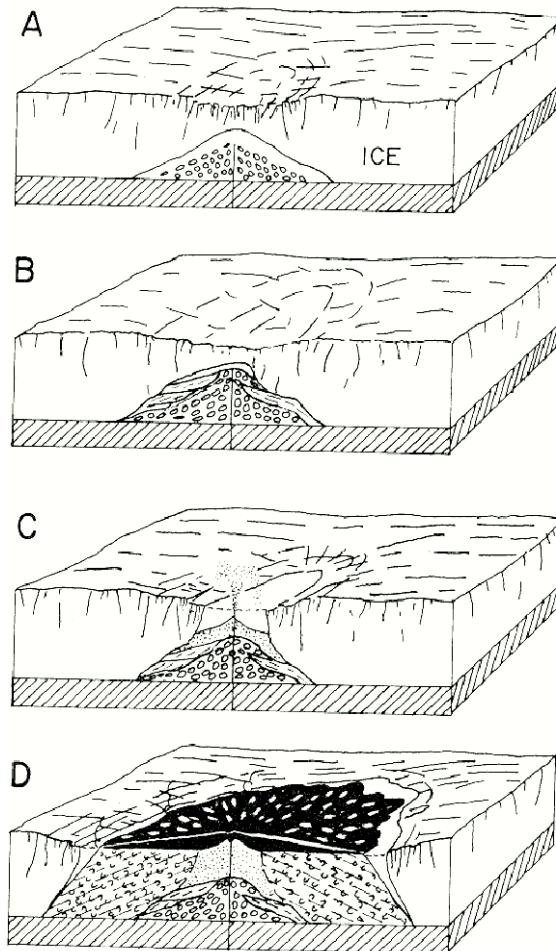


Fissure eruption in the Krafla fissure swarm

Iceland has been variably covered by ice sheets during the last 3 Ma. Heat transfer to the ice during subglacial volcanism is efficient and magma enters a subaqueous environment in the form of water-filled ice cavities or ice-dammed lakes. The character of the eruption products depends on the hydrostatic pressure at the vent and the internal volatile pressure in the magma. Decreasing external pressure leads to a transition from pillow lava via pillow breccia to hyaloclastite tuff. Most Icelandic subglacial volcanic mountains comprise cores of pillow lava, overlain by pillow breccia and hyaloclastite tuff, reflecting decreasing hydrostatic pressure as the mountain grows higher during the eruption. If the vent area becomes subaerial, the volcanism may change to lava eruptions. Icelandic hyaloclastite mountains, capped by subaerial lava flows have generally steep sides and flat tops and are called table mountains.

The landforms developed by subaerial and subglacial volcanism are very different. During subaerial conditions the predominant basaltic eruption products are lava flows from fissure eruptions or gently sloping shield volcanoes. Fissure eruption lavas tend to smooth the topography of the rift zone floor. Some of the postglacial lava flows in Iceland have traveled 50 – 100 km, and some of these flows have traveled outside the rift zones where they originated. The Eldgja (934 – 940 AD) and Laki (1783 – 1784 AD) lava flows are examples, and have volumes of 20 and 14 km³, respectively. In contrast, subglacial fissure eruptions and subglacial “shield volcanism” produce high and narrow hyaloclastite ridges and steep-sided table mountains. Subglacial volcanism therefore tends to build high topography. The high areas under the major Icelandic glaciers, and especially under Vatnajökull, thus grow more rapidly in elevation compared to the surrounding volcanic zones.

Systematic correlations between major and trace element and radiogenic isotope ratios (Sr, Nd, Hf, Pb, Os isotopes) in Icelandic lavas demonstrate that the mantle source is heterogeneous. The geochemistry of basalts from the nearby ridge segments, Vesteris seamount, the Jan Mayen area, the Early Tertiary successions of Greenland and the British Isles indicate that the upper mantle in much of the NE Atlantic is similar to that beneath Iceland. Maximum $^3\text{He}/^4\text{He}$ isotope ratios are the highest on Earth, with values of up to 42 Ra reported from Iceland, and > 50 Ra from Baffin Island [Stuart et al., 2003].

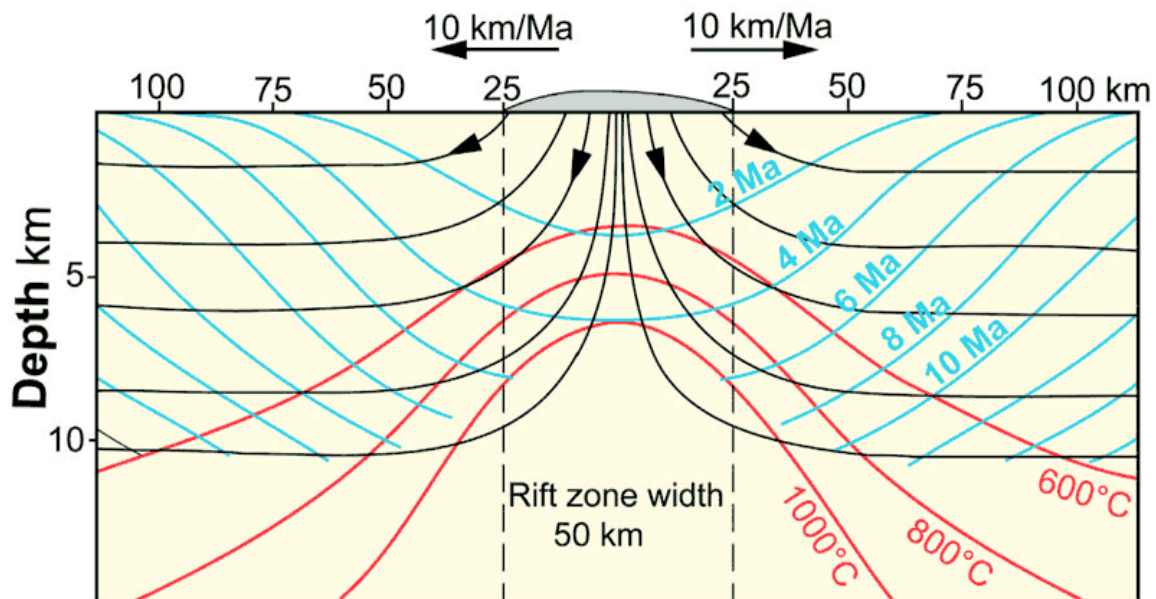


Schematic model for the evolution of Icelandic monogenetic table mountains. At high hydrostatic pressure, a core of pillow lava forms above the vent (A). Slumping on the flanks of the pillow lava pile produces pillow breccia (B). Hyaloclastite tuff is erupted when the external hydrostatic pressure is lower (C), and a lava cap progrades across its own delta of fore-set bedded breccias (D). [from Saemundsson, 1979].

The high proportion of rhyolitic extrusives in Iceland is unique in a global oceanic context. The Torfajökull central volcano is the largest rhyolite center in the present-day terrestrial oceanic environment. Rhyolites and other silicic extrusives are confined to the most evolved central volcanoes. Many of these have also erupted large-volume ash-flow deposits, associated with significant caldera collapse. The most common type of basaltic volcanism along the Icelandic rift zones are fissure eruptions fed from a basaltic magma reservoir under a volcanic center. The largest fissure eruptions of $10 - 20 \text{ km}^3$ are generally quite evolved and homogenous, suggesting extensive fractional

crystallization and assimilation of hydrothermally altered crust. The most common lava type exposed in the Tertiary volcanic successions of eastern and western Iceland is similarly evolved tholeiitic lavas [Hardarson and Fitton, 1991]. Another important basaltic volcano type is the large shield volcanoes that are scattered along the rift zones. This type appears to be unrelated to the volcanic systems and their fissure swarms. These monogenetic shield volcanoes erupted primitive olivine tholeiitic magmas fed by continuous overflow from summit lava lakes. The eruptions appear to have been nearly continuous, with the entire lifetime of the volcanoes completed within about 100 years. The volumes of some of the early post-glacial shield volcanoes range up to 20 km³. Many of the table mountains are subglacial analogs of the shield volcanoes.

The rift zones are continually covered by new lava flows and hyaloclastite mountains. The volcanic productivity of the Icelandic rift zones is anomalously high relative to the low half-spreading rate of ~ 1 cm/year. Rapid subsidence of the partially altered and hydrated volcanic pile occurs. Vertical sections through the Tertiary lava pile in glacially eroded valleys and fjords expose the uppermost 1,500 m of extrusive rocks. The lavas dip gently towards the current or extinct rift zones. This regional flexuring and tilting is a result of the continuous loading and subsidence of the rift zone crust. Whereas the loading is most pronounced under the volcanic centers, the average, time integrated (3 – 7 Ma) subsidence is highest along the rift zone axis and decreases towards the rift zone margins.



Simplified model of Icelandic rift zone dynamics [from Palmason, 1973]. The black, blue and red lines are mass trajectories, age contours (Ma) and temperature contours (°C). Partial melting of hydrated mafic lithologies starts at ~ 5 km depth beneath the central part of the rift zone.

Based on these observations and other geophysical constraints, Palmason [1973] developed a dynamic model for the crustal accretion in Iceland. The lava pile subsides and undergoes prograde metamorphism with increasing pressure and temperature to zeolite, greenschist and amphibolite facies and partial melting producing rhyolitic magmas [Oskarsson et al., 1982]. When partial melting occurs along the walls of basaltic magma chambers, the rhyolitic melt fractions mix with the basaltic liquid and promote magma evolution. In other areas rhyolitic melt fractions segregate and give rise to silicic

intrusions and extrusions. Such crustal reprocessing occurs only to a very limited extent along the midoceanic spreading ridges.

Acknowledgement: Some of the text and figures are adapted or copied from the guidebook of Trönnnes et al. [2003]. GRF thanks R. Trönnnes for kindly supplying an electronic version of this guidebook.

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FIELDTRIPS

Note: The detailed itineraries may alter if required by local conditions at the time.

1. The Reykjanes Peninsula: Tuesday 26th August, 2003

Leaders: Sveinn P. Jakobsson & Gillian R. Foulger

The excursion will focus on Recent volcanism, general petrology and hydrothermal activity.

Hveragerdi – Selvogsheidi – Eldborg – Graenavatn – Krisuvík – Sveifluhals – Ögmundarhraun – Meltunnuklif – Skalamaelifell – Hrolfsvík – Grindavík – Svartsengi – Haleyjabunga – Reykjanes – Langholl – Undirhlidar – Svinahraun – Hveragerdi

Brief description of the Reykjanes Peninsula

The Reykjanes Peninsula is surfaced by basaltic lava flows dating from the last interglacial period. The Mid-Atlantic Ridge gradually shallows towards Iceland, forming the actively spreading Reykjanes Ridge. Its structural continuation on land is the Reykjanes Peninsula, with Keflavik Airport at its NW corner, and Reykjavik at its NE corner. The Peninsula comprises *en echelon* volcanic systems and fissure swarms, with a narrow (2 to 5 km) seismic zone along the plate boundary between the North American and Eurasian plates. The fissure swarms are oblique to the actual plate boundary and thus extend a few kilometres into the plates on either side.

The least compressive stress in the Peninsula is horizontal and trends NW-SE or perpendicular to the boundary, but the maximum compressive stress and strain release varies in direction along the Peninsula. In the SW the Peninsula is characterized by normal faulting, with maximum earthquake magnitudes of 5 to 5.5. In the east the strain release is more strike-slip and maximum magnitudes are up to 6.5. The most recent seismic episodes occurred in 1929-1935 and 1967-73. The latest magmatic episodes occurred in the tenth and eleventh centuries, and the last eruption was in 1340 AD.

Volcanic activity on the Reykjanes Peninsula has been intense in Postglacial times, but has generally been given little attention, because no very recent events have occurred. The number of eruptions is not known but totals a few hundred. The volcanic activity is more or less restricted to the active fissure zones and seems to be periodic. The time lapse between periods is about 1,000 years but each period lasts for about 300 – 400 years. During each period all the fissure swarms are active and the activity starts at one end of the Peninsula and moves to the other. The last period started in the 10th century and lasted to about 1340 AD. The first eruptions took place at the eastern end (in Hellisheidi and Blafjoll) but spread to the west. Eruptions within each volcanic system are thought to behave in a similar manner to those of the Krafla Fires that occurred 1975-1984, i.e. rifting episodes which last for a few years or decades, accompanied by a few or numerous eruptions. At each time only one fissure swarm may be active. The eruptions established from historic accounts, geological mapping and radiocarbon dating are as follows:

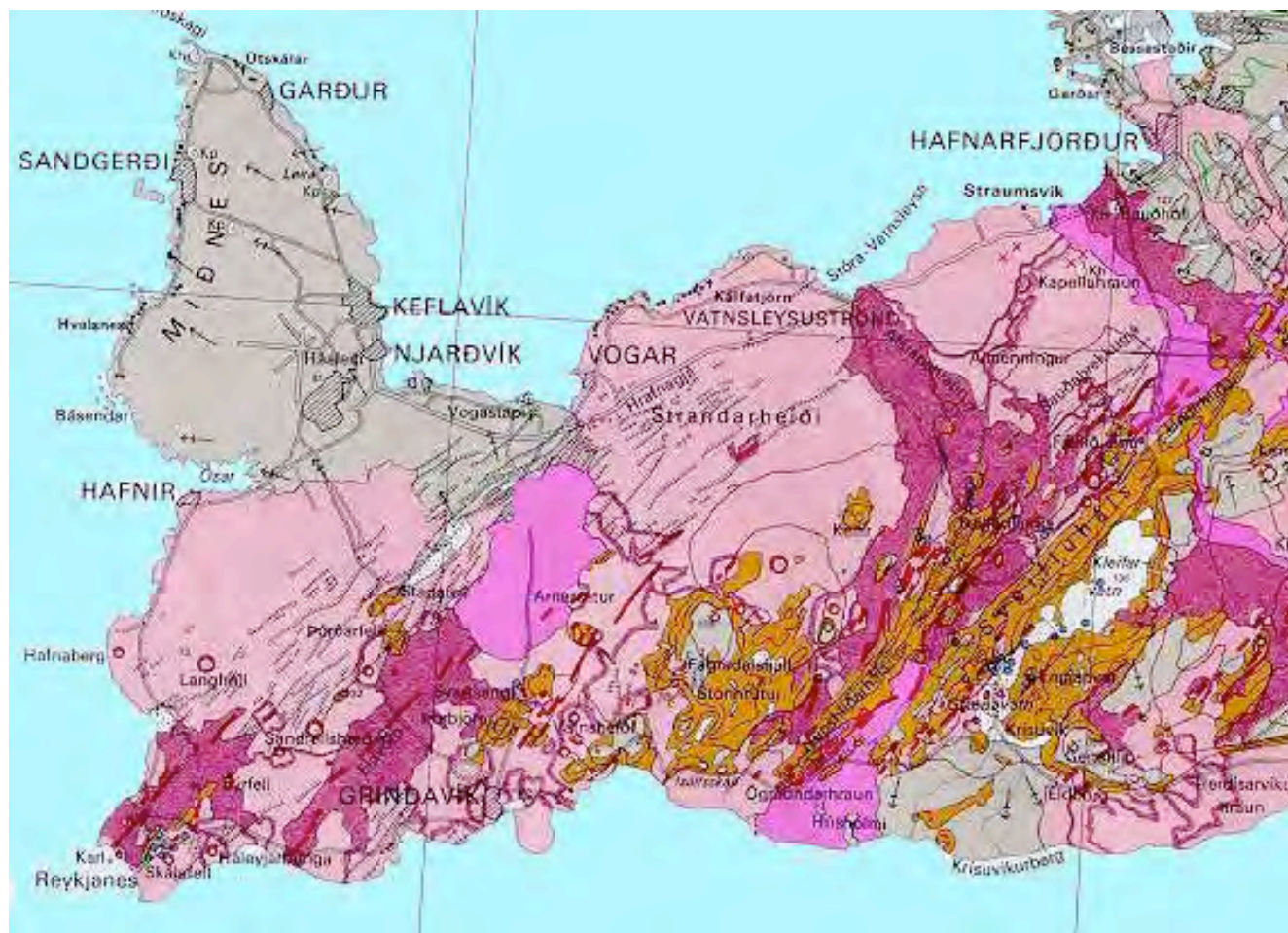
950-1000 AD in the Brennisteinsfjoll and Blafjoll swarm

1151 and 1188 AD in the Krisuvik and Trolladyngja swarm

1210-1240 AD in the Reykjanes swarm

1340 AD in the Brennisteinsfjoll swarm

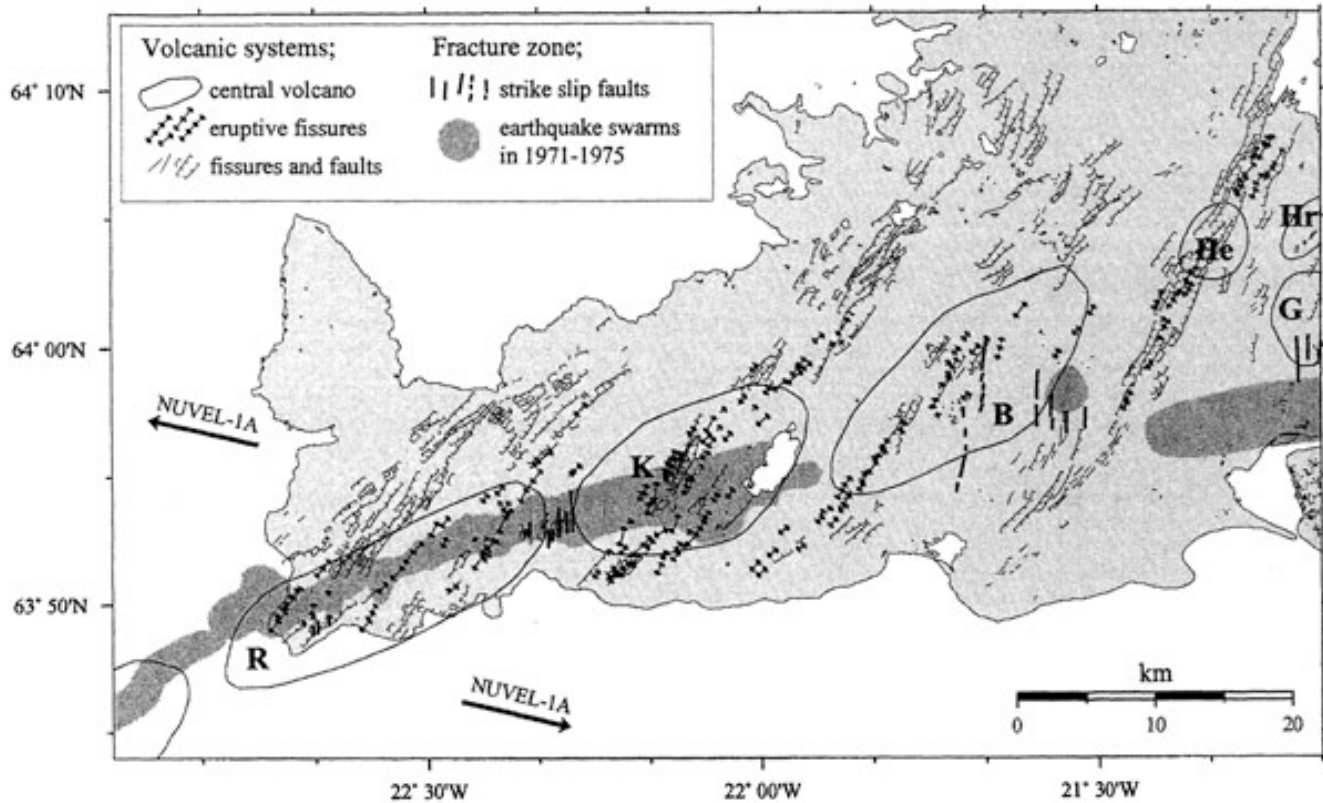
Numerous volcanic eruptions have occurred close offshore on the Reykjanes Ridge throughout the centuries, the last confirmed one in 1926.



Geological map of the Reykjanes Peninsula

The four major fissure swarms are 25 to 50 km in length, have shallow grabens, and are probably the result of repeated dike injection in the crust. There is typically a crater row in the centre of each fissure swarm. Deformation on the Peninsula is of two types: (a) seismic – involving principally brittle failure and earthquakes along a narrow zone, and (b) magmatic, where magma is introduced into the crust. In this case the magmatic fluid causes the crust to fail, and magma propagates laterally and vertically along fractures to form dikes and feed eruptions.

There are additionally N-S trending fissure systems on the Reykjanes Peninsula, which are less obvious than the NE-SW elements. They are shorter (5 to 10 km), and mainly located along the plate boundary. They are collectively arrayed EW and most of the shield volcanoes on the Peninsula also follow this trend. The N-S fractures are believed to be the continuation of the South Iceland Seismic Zone, which will be toured during the 4-day post-Conference fieldtrip.



Tectonic map of the Reykjanes Peninsula showing the en echelon volcanic systems

The majority of the Reykjanes Peninsula is covered by post-glacial basaltic lavas (< 12,000 YBP). The oldest rock formations are interglacial basaltic lavas, composed of very coarse grained basalt lavas that were mainly erupted from shield volcanoes during the last interglacial. The part of the Peninsula which includes Keflavik Airport is composed of interglacial lava flows. They are invariably glaciated, with glacial striae at the surface.

Second in age are hyaloclastites or basaltic subglacial tuffs and pillow lavas, formed by eruptions of basaltic magma beneath the ice cap during the last glacial stage (ca. 120,000 to 12,000 YBP). Because of the presence of ice cover, the eruptives were restricted in their distribution and piled up near and over the vent, as pillow lavas and hyaloclastites. All of the prominent mountains on the Peninsula are subglacial volcanoes formed in this manner, but in some the activity was sufficiently vigorous to build up an edifice above the level of the ice sheet, resulting subaerial lavas which cap these volcanoes and form table mountains.

Subglacial and subaerial (post-glacial) fissure eruptions have formed prominent NE-trending ridges and crater rows that dominate the topography of the Peninsula. Subglacial eruptions produce ridges of hyaloclastite similar to the Axial Volcanic Ridges (AVRs) on the Reykjanes Ridge. A number of table mountains and hyaloclastite cones, products of sub-glacial eruptions from isolated vents, are also present on the Peninsula and closely resemble the small seamounts that have been mapped on the MAR. Early, large-volume post-glacial basaltic and small-volume picritic shield volcanoes have also played a major role in surfacing this ridge segment with voluminous pahoehoe lava flows, which both cover and are covered by the products of fissure eruptions. Shield volcanoes and eruptive fissures have been active on the Peninsula during the Holocene.

The base of the seismogenic zone is ~ 5 - 11 km depth on the Peninsula and most seismicity occurs at depths from 1 - 5 km. A narrow zone of seismicity 2 - 5 km wide, characterized by predominantly strike-slip focal mechanisms and extending the entire length of the peninsula, is identified as the currently active plate boundary. Geodetic measurements show that left-lateral shear is currently occurring along on the Peninsula. Data from Satellite Radar Interferometry indicate that below a depth of 5 km plate motion is accommodated by continuous ductile deformation.

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http://www.volcanotours.com/iceland/fieldguide/reykjannes_Peninsula.htm

2. The Hengill triple junction, Thingvellir, Geysir, Gullfoss: Thurs. 28th August, 2003

Leaders: Gillian R. Foulger & Axel Bjornsson

The excursion will focus on the tectonics, geophysics and geothermal activity of the Hengill triple junction, with the afternoon touring Thingvellir, the site of the ancient Icelandic parliament, the Great Geysir and its geysir field, and the Gullfoss waterfall.

Hveragerdi – Kambar – Hveradalur – Pipeline road – Dyradalur – Kyrdalur – Nesjavellir power station – Thingvellir – Geysir – Gullfoss – Hveragerdi

Brief description of the areas to be visited

Geothermal activity in Hveragerdi

The hot springs of Hveragerdi (lit. “hot spring enclosure”) are fed by hot water that runs from the Grensdalur volcanic area south along fissures. The main fissure has a warm water stream at a depth of 125 m and a temperature of 180°C (356°F). At a depth of 250 m there is another stream that is 10°C cooler.

The geothermal activity is variable, and hot springs migrate about. Once a house was destroyed when a new hot spring developed beneath the living room (“Badstofahver – lit. “living-room hot spring”). The geothermal heat has been utilised, probably since the settlement of Iceland in the 9th century. Traditionally clothes were washed, food was cooked, and bread was baked in steam boxes (hverabraud – lit. “hot spring bread”). Later, pipelines were built from the springs and cisterns installed next to homes.

Some significant dates:

- 1929 The first establishment of a greenhouse.
- 1939-41 The first hot water drilling for the use of thermal energy which was used for a laundry for British soldiers during World War II.
- 1947 The volcanic eruption of Mt. Hekla caused some hot springs to disappear and new ones to be born.
- 1952 Municipal heating established by drilling eight holes in the geothermal area. Gradually private holes were added and use of the municipal wells diminished.
- 1970 Drilling by Orkustofnun (the National Energy Authority).

Today, geothermal energy heats 50,000 m² of greenhouses, private homes and the swimming pool. Half of the private homes are heated with groundwater that has been heated with steam. Other private homes, industrial sites, commercial establishments and greenhouses are heated directly with steam.

The Hengill ridge-ridge-transform triple junction

The Hengill ridge-ridge-transform triple junction in southeastern Iceland lies where the Reykjanes Peninsula Volcanic Zone, the Western Volcanic Zone, and the South Iceland Seismic Zone meet. The locus of ridge volcanism and spreading within the triple junction has migrated westward by a few kilometers over the last 1 Ma or so, from the Grensdalur system to the Hengill system. The area now comprises a tripartite complex of volcanic systems, the eastern two of which are currently inactive.

A NNE-striking fissure swarm associated with the Grensdalur volcano was the active spreading center until about 0.3 Ma. Hveragerdi lies at the southern edge of the Grensdalur central volcano and exploits geothermal heat from that system. After spreading ceased along the Grensdalur system, activity migrated west to the Hromundartindur and Hengill systems, a few kilometers to the west. The Hromundartindur system, dominated by the mountain by the same name produced mainly fissure eruptions. It is now almost extinct, having erupted last ~ 10,000 years ago. It did not develop a central volcano to the same degree of maturity as the other two systems.

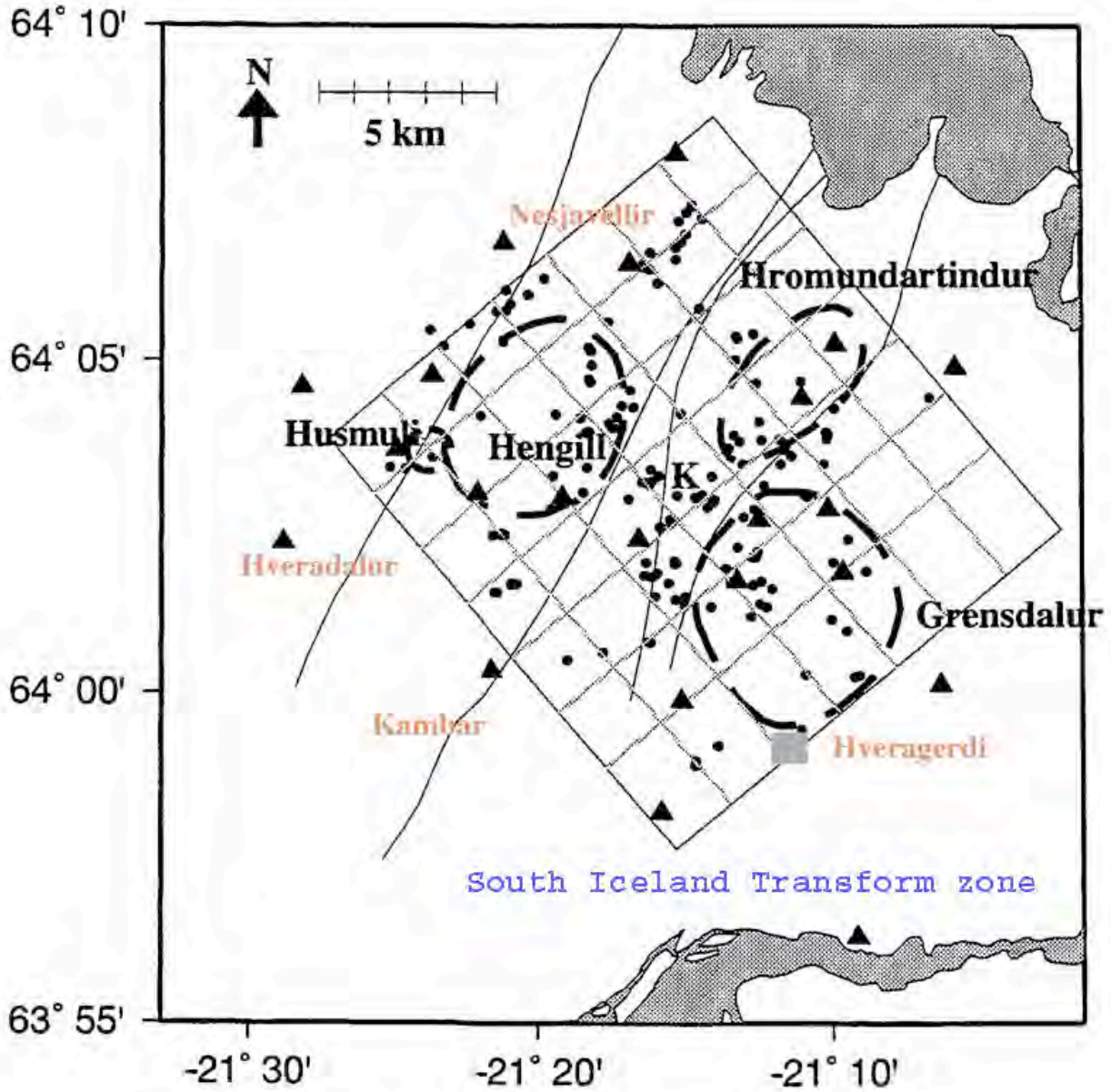
The Hengill segment is currently the locus of spreading and volcanism, and contains intermediate rocks, suggesting a long-lived magma chamber. It is dominated topographically by Mt. Hengill, which is flanked on the west by a large basalt shield, Húsmuli. The Hengill system has erupted several times since the last glaciation. 200,000 years ago, the Nesjahraun lava field north of Mt. Hengill flowed into Lake Thingvellir. The Hengill area contains widespread geothermal resources that are currently under development for electricity generation and hot water.

To the south of the Hengill volcanic complex, south of 64°N, lies the western end of the South Iceland Seismic Zone, an east-west array of north-striking faults that generate earthquakes up to about magnitude 7.

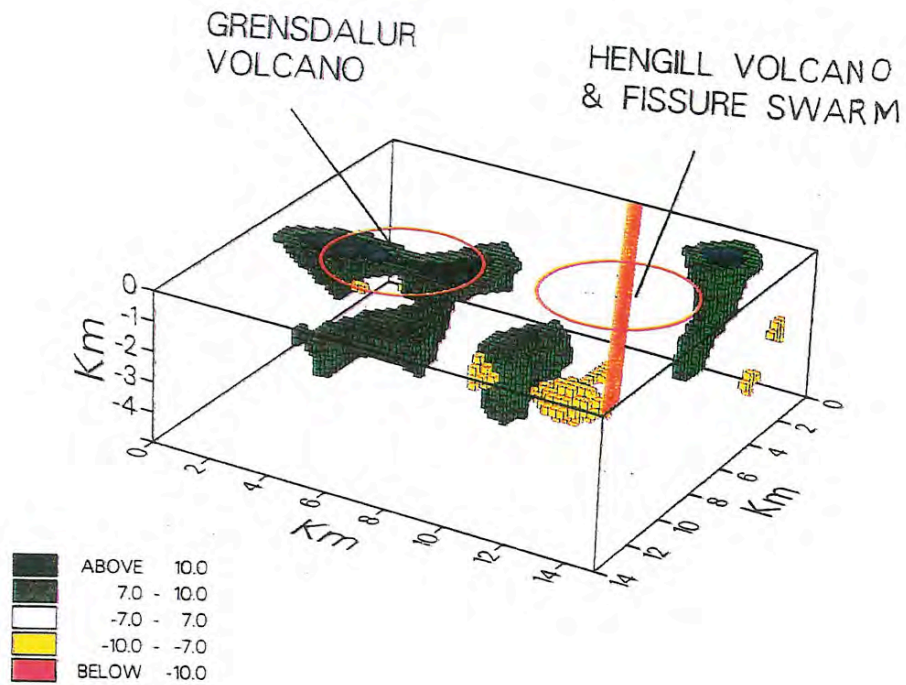
The natural surface heat loss is about 350 MW. The triple junction, in particular the Grensdalur system, is associated with small-magnitude earthquake activity that is curious and unique on a world scale. Small-magnitude seismicity is ongoing on a daily basis at the very high rate of one magnitude 1 earthquake per day. Most earthquakes associated with the volcanic systems are smaller than ~ magnitude 4. They have rare, non-shear mechanisms and are thought to be caused by cooling-contraction cracking in the geothermal heat source, as heat is mined from it and transported to the surface by geothermal fluids. Earthquake monitoring experiments have mapped the seismic volume, and thus the heat source, by locating these earthquakes in three dimensions. This situation is ideal for Local Earthquake Tomography, which has been applied to reveal the three-dimensional structure of the volcanic complex in considerable detail.

Geothermal activity is associated with all three volcanic systems in the area, and with the transform branch of the triple junction. That associated with the volcanic systems is a so-called “high-temperature” geothermal resource, and is thought to derive its heat from volcanic intrusions. That associated with the transform branch is so-called “low temperature” geothermal heat, and is thought to arise from deep circulation of groundwater in faults. The geothermal heat in Hveragerdi, which is at the southern edge of the Grensdalur central volcano, is the oldest of the high-temperature resources. The most remarkable geothermal features associated with the Hrómundartindur system are CO₂-rich springs

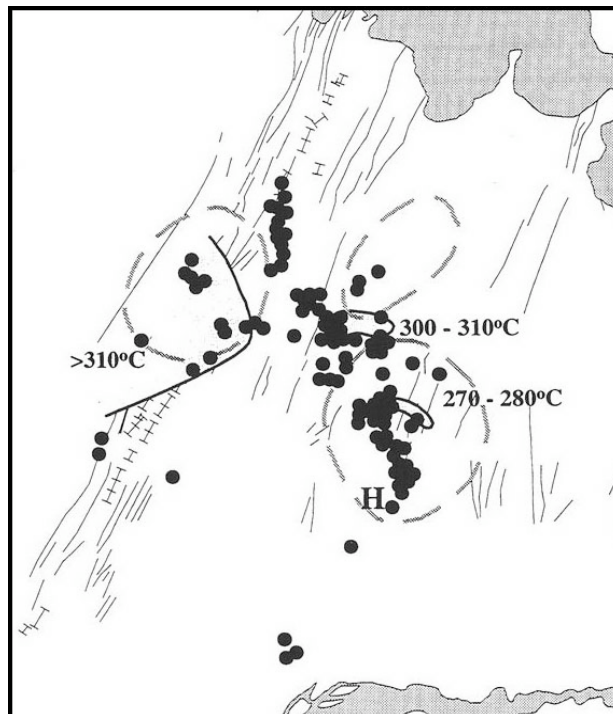
on Öldukelsháls (lit. “beer-spring ridge”) which can be reached by a long but magical hike up into the mountains. Geothermal activity associated with the Hengill system is widespread from Nesjavellir north of Mt. Hengill to Hveradalur (lit. “Hot spring valley”) south of it. The source of water is thought to be precipitation falling on the highlands north of Lake Thingvellir that seeps at a depth of 1 – 3 km and flows underground to lower areas.



Volcanic systems of the Hengill ridge-ridge-transform triple junction. Places that will be visited in the fieldtrip are labeled in red. Box shows area tomographically imaged.



Local earthquake tomography image of the Hengill triple junction. Scale indicates wave-speed anomalies in % deviation from the starting model.



The three high-temperature geothermal fields within the volcanic systems of the Hengill triple junction

Geothermal activity at Nesjavellir

Test drilling began at Nesjavellir in 1965 and continued until 1986. Nesjavellir is the focal point for heat extraction to serve Reykjavik because of land ownership issues. Hot water flows from beneath Hengill along the Kyrðalur Ridge to Nesjavellir. The water and steam extracted is used to heat local groundwater from 4 – 5°C to near-boiling. There are 22 boreholes at Nesjavellir, but five of them are permanently closed. Most of the boreholes are 1-2 km deep and the highest temperature measured is 380°C. Each borehole provides about 60 MW of thermal power of which about 30 MW are utilizable. This power is sufficient to heat a settlement of 7,500 people. Cold ground water is heated to just above 80°C and piped to Reykjavik (25 km) where it is used for central heating. The diameter of the pipe is 90 cm. The construction of the Nesjavellir Power Plant began in early 1987 and was opened by 1990 using four holes generating 100 MW with production capability of 560 l/s. At present, the capacity is 1100 l/s of hot water and 90 MW of electricity. The Nesjavellir thermal field is projected to sustain hot water production of 400 MW for at least the next 30 years.

Thingvellir

Thingvellir is at the western margin of the major fissure swarm associated with the Hengill system. This part of the fissure swarm is a large graben structure, about 6 km in length and 80 m deep, and results from extensional tectonics and postglacial subsidence. It is bounded by very recent normal faults and open fissures. To the northeast of Thingvellir is the large basaltic shield volcano Skjaldbreidur, which erupted about 10,000 years ago and produced ~ 17 km³ of lavas. The lava also flowed over Thingvellir and to the south, over land that since has subsided to form the lake. Major subsidence and crustal extension continued subsequently, forming the spectacular Almannagja (lit. “Commonwealth fissure”) fissure and normal faults. Further subsidence occurred at Almannagja in 1789, and Thingvallavatn then subsided about 2.5 m, when the lake advanced onto fields and farmland.

The most recent volcanic event in the Thingvallavatn area occurred ~ 2,000 years ago, when a fissure eruption in the Nesjavellir region formed the Nesjahraun basaltic lava flow, which advanced into the lake. At the same time a volcanic eruption broke out in the center of the lake, producing the tuff cone of Sandey, rising from about 100 m depth, and reaching 74 m above lake level.

In the Thingvellir area, the fissures occur within a basaltic lava flow of about 9,000 years old, derived from a fissure eruption in Tindfjallaheidi to the north-east. In 1938 the total extension or dilation of the 6.16 km wide graben, post-dating the lava flow cover was 75 m, or 1.25%, with almost all of the extension near the margins of the graben and no extension in the center. This amounts to an average rate of extension of 0.83 cm/yr and is broadly comparable to recent GPS measured rates of extension. This is consistent with the expectation that the NUVEL full 1.9 cm/year extension rate at Iceland is shared between the Western and Eastern Volcanic Zones in south Iceland.

The Almannagja fissure can be traced over a distance of 7.7 km, and its maximum width is 64 m. Most of the larger fissures are really normal fault features, with near vertical walls. The maximum vertical displacement in Almannagja is 40 m. The Almannagja and other fissures may be related to the intrusion of dikes at depth in the crust. Nevertheless, many fissures are reactivated and long-lived, and also exhibit vertical displacements.



Geysir

The Great Geysir of Iceland is the source of the English word “geyser”, meaning an erupting hot spring. Geysers occur where groundwater flows through rocks with exceptionally high thermal gradients and relatively low permeability, resulting from the self-sealing of rock formations due to the deposition of silica or other chemical precipitates from hot hydrothermal solutions. The vertical water column of water in the geyser is heated to near its boiling temperature. Because of the effect of water pressure, the boiling temperature increases with depth. At 5 m it is about 112°C, and at 10 m depth it is about 121°C. Any perturbation that leads to decrease of pressure of the hot water column will bring about superheating and boiling at depth. If water is ejected at the surface, deeper waters exceed boiling point, flash to steam, and bring about an eruption.

The Great Geysir ejects a jet of steam and water to height of 60 to 70 m, and the eruptions last about 10 minutes. The eruptions are preceded by rumbling sounds at depth, due to the collapse of large steam bubbles in the water column, and this is accompanied by rise and fall of the water level in the large bowl over the vent. After a few seconds the jets of steam and water begin, rapidly reaching a climax that lasts a few minutes. Then the eruption ceases suddenly, and the water level is deep within the pipe because of the large amount which has been expelled. After a few seconds a new and very powerful jet emerges, which lasts about five minutes. The vertical pipe of Geysir is about 20 m long.

The hydrothermal waters of the Geysir region carry large quantities of dissolved silica, which are precipitated as siliceous sinter around the hot springs when the waters emerge at the surface. The deposition is the result of decreased solubility of silica in the waters due to cooling at or near the surface. In the Geysir region the sinter may be 1 - 2 m thick.

In the early part of this century the eruptions of Geysir became infrequent, and essentially ceased by 1916. This was attributed to the fact that the surface area of the bowl had become very large, resulting in rapid cooling of the water at the surface, and raising the water level. New eruptions were induced by cutting a notch in the edge of the Geysir bowl, and lowering the water level. Another method to induce an eruption is to place large quantities of soap in the water. This has the effect of decreasing the surface tension and facilitates superheating of the water, leading to eruption. Most of the steam eruptions in the Geysir area today occur from the geyser Strokkur, which is an old borehole. The eruptive behaviour of the geysers at this field are very sensitive to large earthquakes in the South Iceland Seismic Zone.

Gullfoss

The waterfall Gullfoss (lit. “Golden waterfall”) is fed by the river Hvita (lit. “White river”) which arises from under the Langjokull glacier and reaches the sea after travelling 133 km. The canyon below Gullfoss extends for ~ 2.5 km and reaches a depth of 70 metres. It may have been formed in torrential

floods caused by so-called jökulhlaups (glacial outbursts) near the end of the last ice age. During a jökulhlaup the amount of water running seaward during a single 24-hour period can equal a normal flow of up to five years, but the erosive force of such sudden deluges is many times greater.

Gullfoss is actually two separate waterfalls, the upper one has a drop of 11 metres and the lower one 21 metres. The rock of the river bed was formed during an interglacial period. Water flows over the falls at an average rate of 109 m³/s. The heaviest floods have recorded a flow of 2,000 m³/s. During the summer the flow is 130 m³/s, but during the winter the waterfall is largely frozen. The gorge below the waterfall cuts into a succession of interglacial lava flows intercalated by fluvial sediments.

Proposal for dams and hydro-electric power plants on the Hvita river have been proposed, which could produce ~ 2,500 GW hrs of electricity annually and double Iceland's production. However, Gullfoss is considered a national treasure, and it is extremely unlikely that such a thing will ever be done.



Gullfoss

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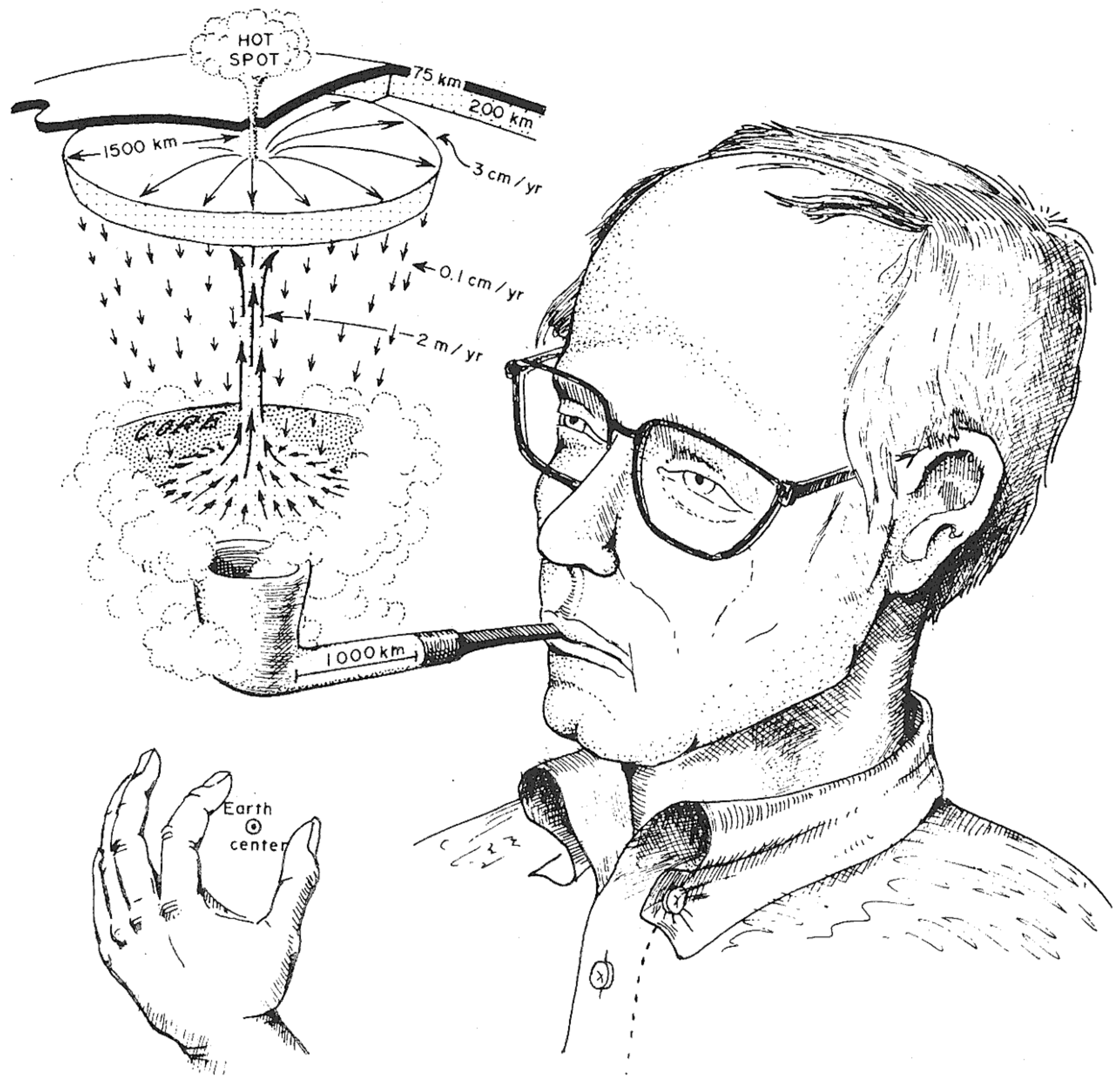


Fig. 1. Conception of the mantle plume theory, adapted liberally from W. J. Morgan (unpuffed data, 1977).

Graphic Solutions to Problems of Plumacy

John C. Holden and Peter R. Vogt

Preface

The mantle plume is just the youngest member of a big and colorful family of geological fads and fashions: diluvialisms and catastrophisms, earths expanding and contracting, global tectonics new and old. Some of these fads have become bandwagons rolling from theory to fact. Others are intellectual white elephants gathering library dust. We do not know yet how the mantle plume will fare; certainly

it has not quite attracted the bandwagon that the new global tectonics did.

Since plumes are better hidden from observation than plates, it may take years to prove or disprove their existence. This is just as well—one cannot write a research proposal to prove that the earth is round or that the continents drift.

In this paper we hope to cut through the hullabaloo surrounding mantle plumes by offering a graphic commentary on the 'hot' topic. Any resemblance between persons or deities

depicted here and those living or dead may or may not be coincidental. If kings and statesmen fall to the cartoonist's pen, why not scientists, their students, instruments, and Mother Earth herself?

Introduction

It was *Wilson* [1963] who first suggested that hot spots were fixed in the upper mantle and created aseismic ridges as crustal plates moved over them. This idea was expanded to include a mechanism that would also

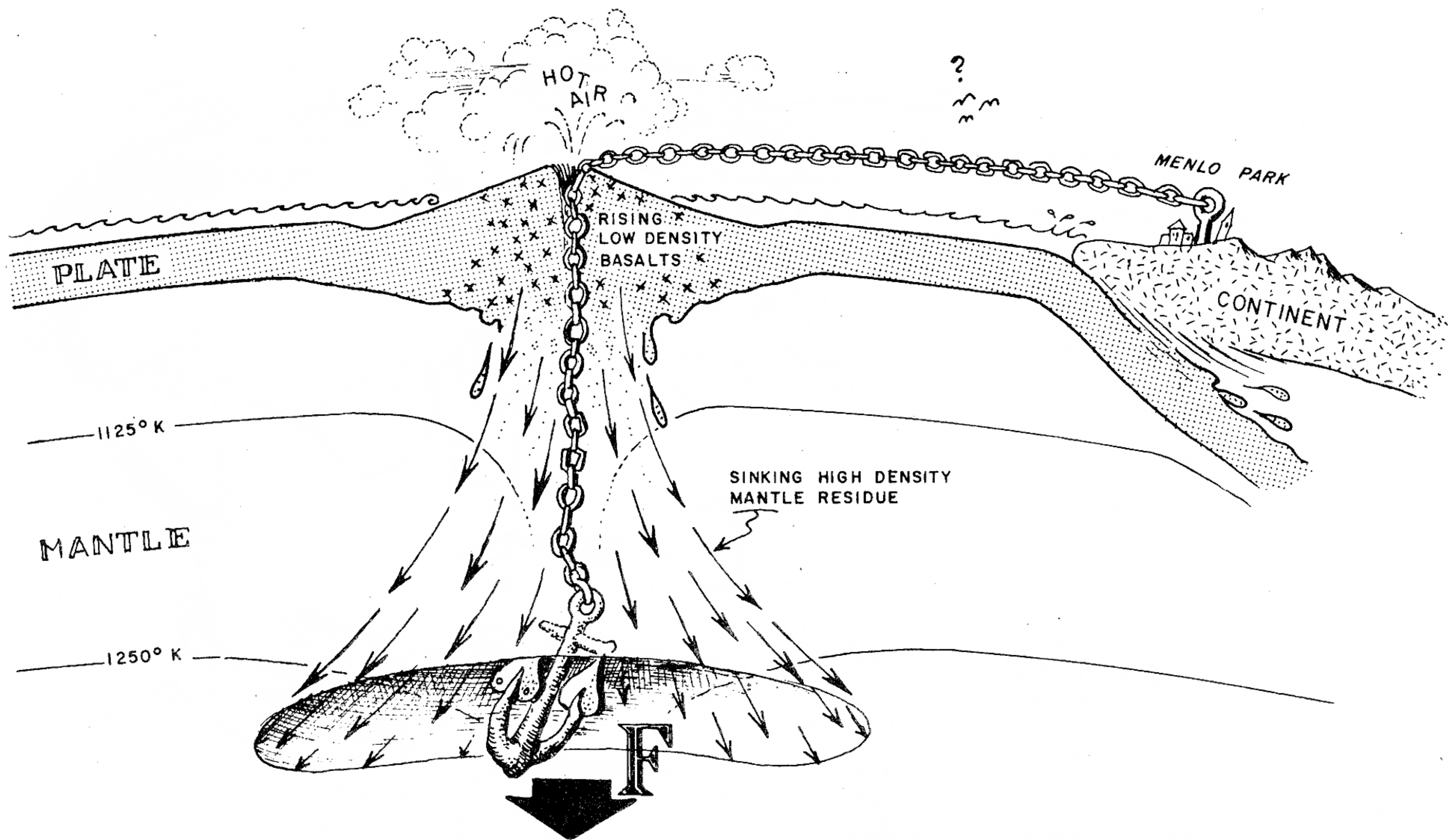


Fig. 2. Gravitational anchor theory, showing the origin of Hawaii [Shaw and Jackson, 1973]. According to this hypothesis, even the motion of California can be explained (although the motions of Californians remain as mysterious as ever).

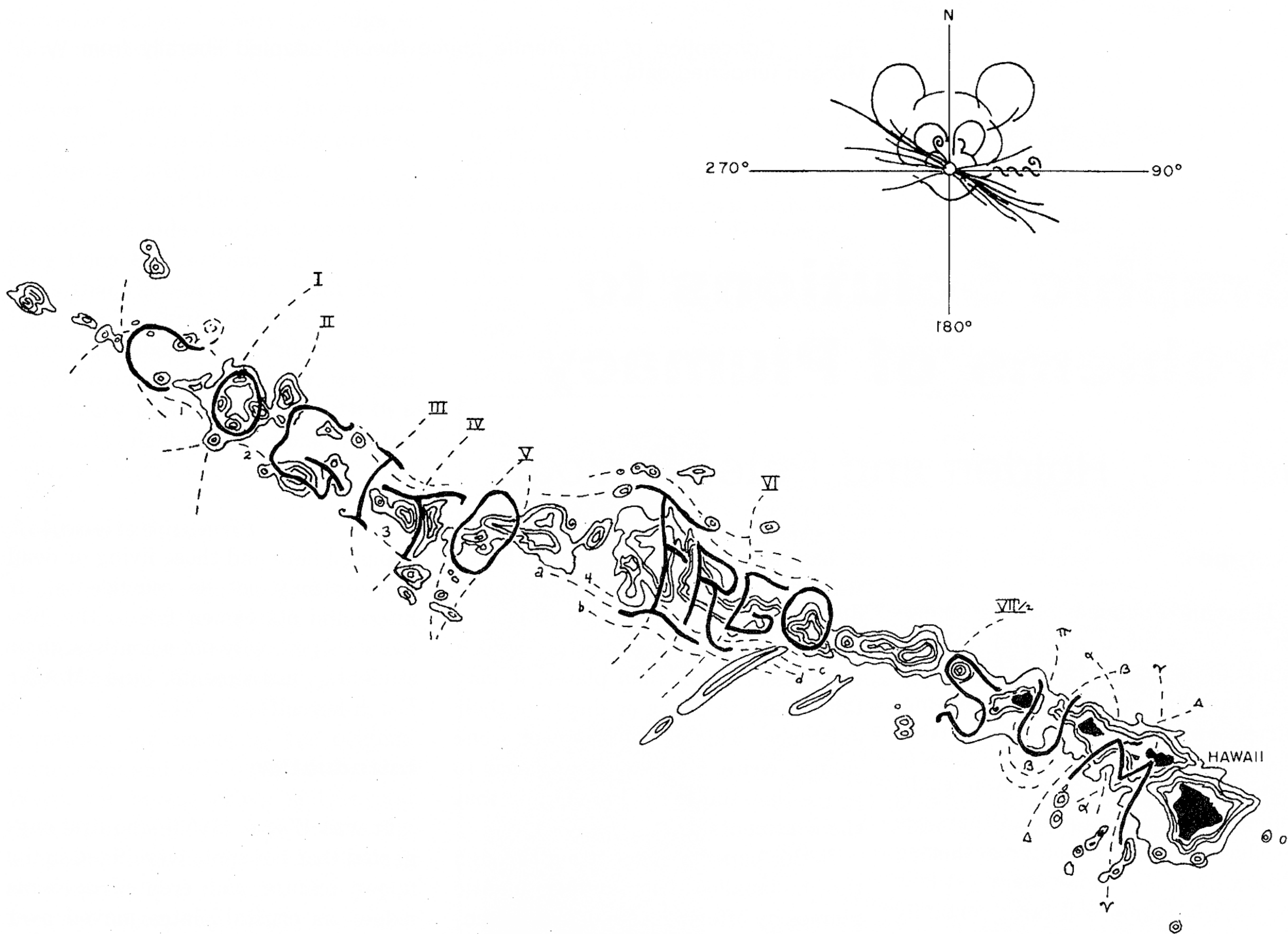


Fig. 3. Stress field lineations plotted from volcano distribution along the Hawaiian chain. The geophysical relevance of the message spelled out is a matter of intense debate.

account not only for hot spots but also for the causal forces of plate tectonics and continental drift. In short, it was proposed that ascending convection plumes exist in the deep mantle below active hot spot volcanism [Morgan, 1971, 1972]. Geometrically, these features are toroidal cells with narrow (200-300 km wide) vertical axes through which hot material is transferred from the lower to the higher regions in the mantle.

An entire science has developed around the study of plumes, albeit perhaps no more scientific than social, spiritual, or political science. If the concept is valid, then most first-order features of the earth's surface may be attributed directly or indirectly to plumes. We term this discipline 'plumacy,' its practitioners 'plumatics' (also 'plume freaks'). Recognizing the parallels between religious and geological faiths, we also propose the term 'aplumatics' for those who do not believe that mantle plumes exist; we carry over the term 'agnostic' to apply to all those fainthearted earth scientists who refuse to debate the issue on grounds that the existence or nonexistence of mantle plumes cannot be proved since the terms are not defined. This position is untenable, since the definition and etymology of the word plume has been published in the geological literature [Anderson, 1975]. Because he is a little-known author (just one of the numerous Andersons in earth science) publishing in an obscure journal, it is worthwhile to reproduce his definition here.

Tozer [1973] has objected to the use of the word 'plume' because of prior usage and connotations. However, the various definitions of 'plume' and its antecedents in French, Latin, and German seem to provide enough flexibility to describe the phenomenon, its implications, and its raison d'être on the one hand and its inventors, supporters, and detractors on the other: *Plume* (English, from French and Latin, pluma)—a feather, a long handsome feature, a token of honor or prowess; a prize, to pride or congratulate; to preen. *Plombe* (Germanic)—a plug. *Plombe* (Old English from West Germanic)—something especially desirable, as a good position. *panache* (French)—trail, stripe, swagger; *fumée* (French)—smoke, fumes, steam; *fumer* (verb)—to fume, to dung, to manure; *plumitif* (familiar)—scribbler, pen-pusher.

We leave the definition of hot spots (also called melting spots and melting

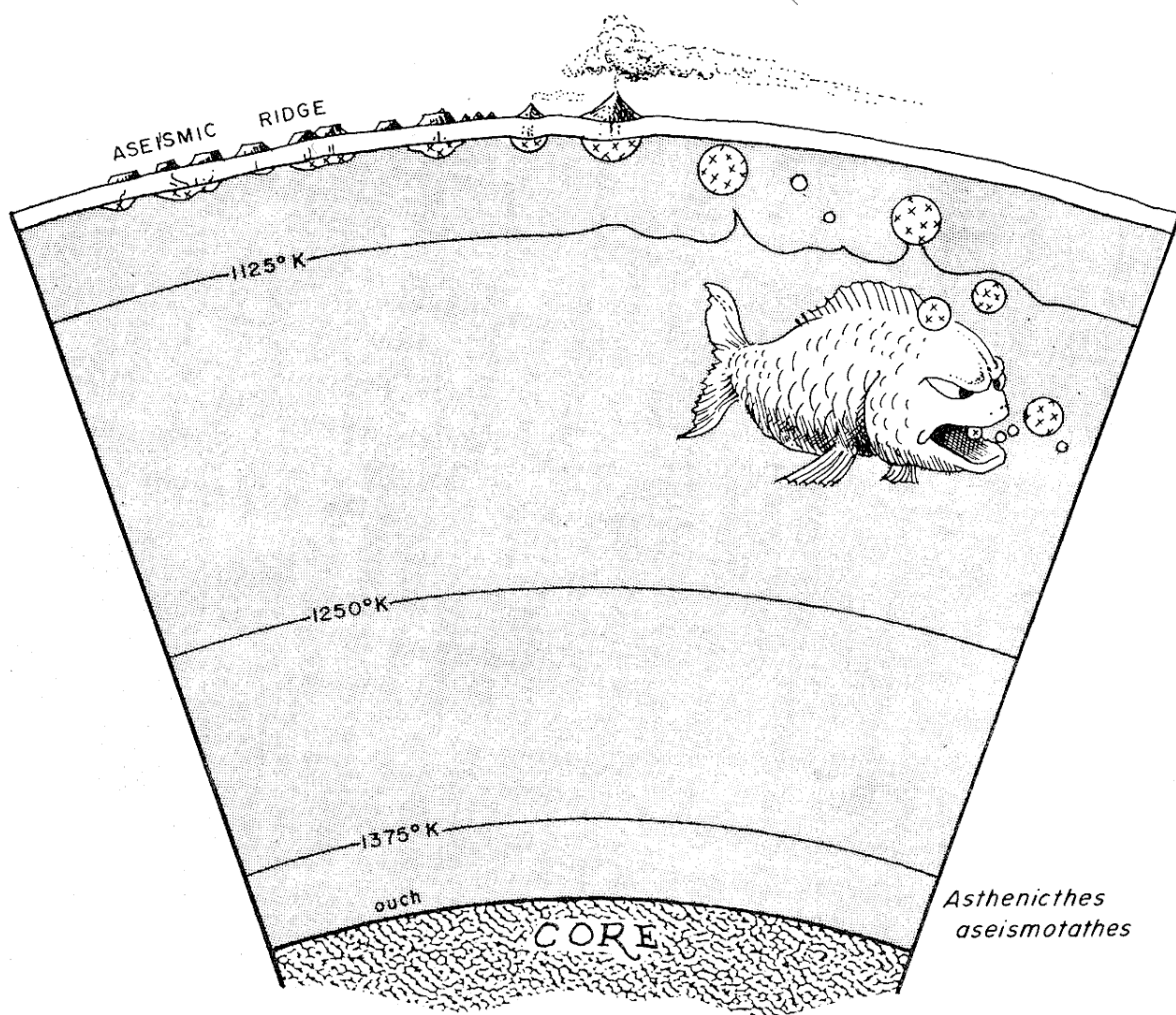


Fig. 4. Alternative to the mantle plume theory (based on an ancient Japanese legend).

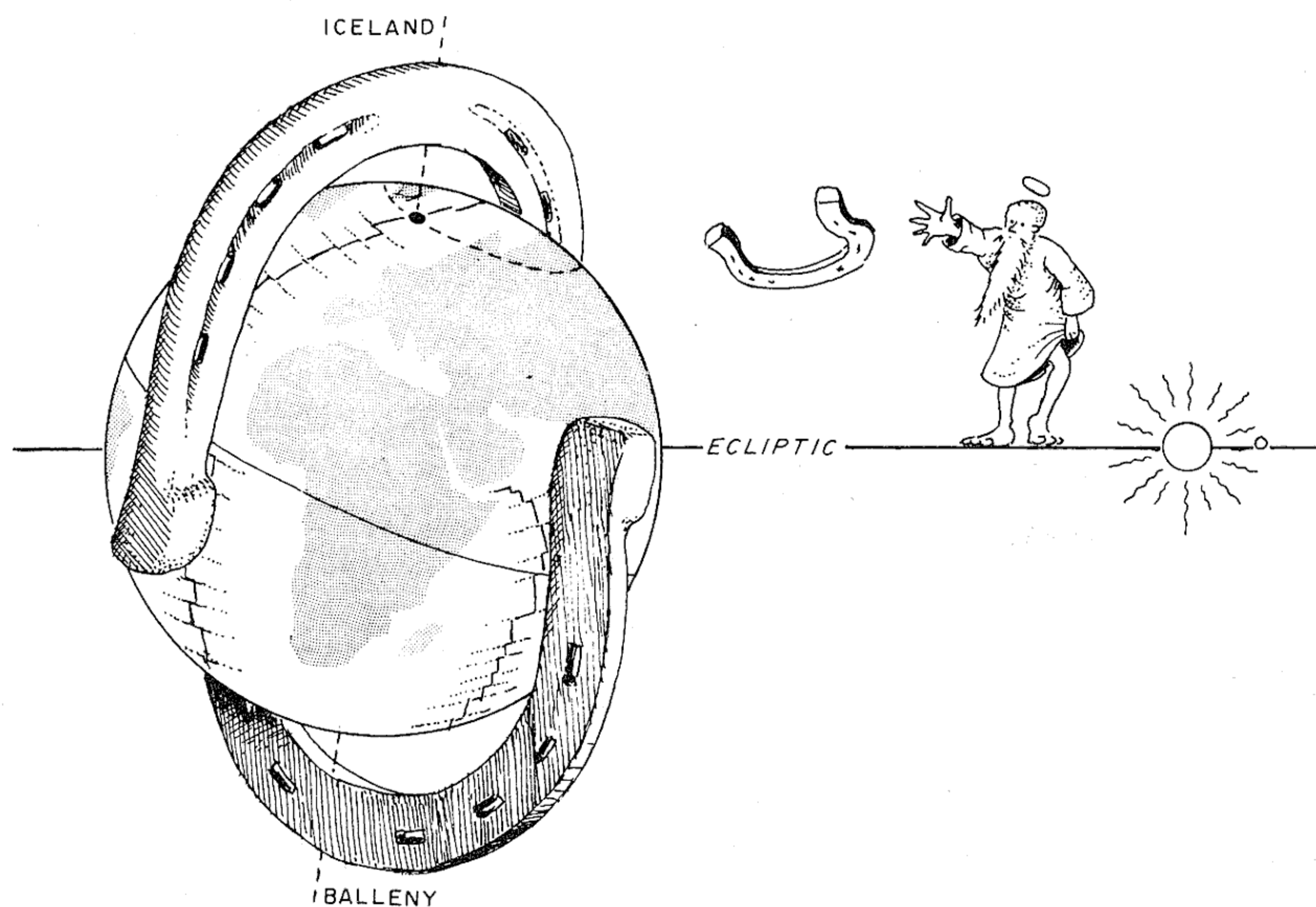


Fig. 5. Schematic of a theory for the origin of midoceanic rifts, hot spots, the solar system, creation, and God.

anomalies) as an exercise for the reader.

Plumes, Antiplumes, and Other Explanations for Hot Spots

Plumes are often invoked to explain the source of the force causing sea floor spreading and plate motion. Unfortunately, no airtight case has been made for the cause of the plumes themselves, and the original problem of where it all begins is not solved but merely pushed deeper into the earth's interior. Therefore we must in all fairness discuss not only mantle plumes but alternate theories put forth to explain hot spots.

One wonders what was being smoked when the plume concept was formed (Figure 1). It is a well-known

dictum in geopolitics that for popular acceptance the terms one chooses are equally as important as the ideas themselves. Therefore when the plume concept was under construction, the stem of the plume was cleverly dubbed a pipe, which immediately brings to mind kimberlite pipes of mantle origin. . . Bingo, a winner! After all, it is much more propitious for material to move 'up the pipes' than to go 'down the tubes.' However, to suggest that the term pipe was picked only for its public relations value is not entirely fair. After all, when geophysicists conceptualize, they habitually use the stems of their pipes as a scale for relative proportions. So it is only logical that the stem of a plume should be called a pipe; Q.E.D.

A few years after the plume idea

was out, *Shaw and Jackson* [1973] proposed a slight modification of it in their gravitational anchor theory (Figure 2). According to this scheme, dense crustal residues are sinking beneath Pacific hot spots at the eastern ends of the Hawaiian, Tuamoto, and Austral chains as low-density basaltic magmas are distilled out and up to form the spot on the hot spot. The geometry of this specialized convection cell is that of an upside-down plume, or antiplume as it were. Shaw and Jackson are absolutely correct in assuming that these anchors do not drive the plates but rather act as pinning points. We presume that the chain (not the volcanic chain) attached to the anchor (and nowhere discussed by the original authors) extends out of the orifice of the volcano. Where it goes after that is uncertain; perhaps it

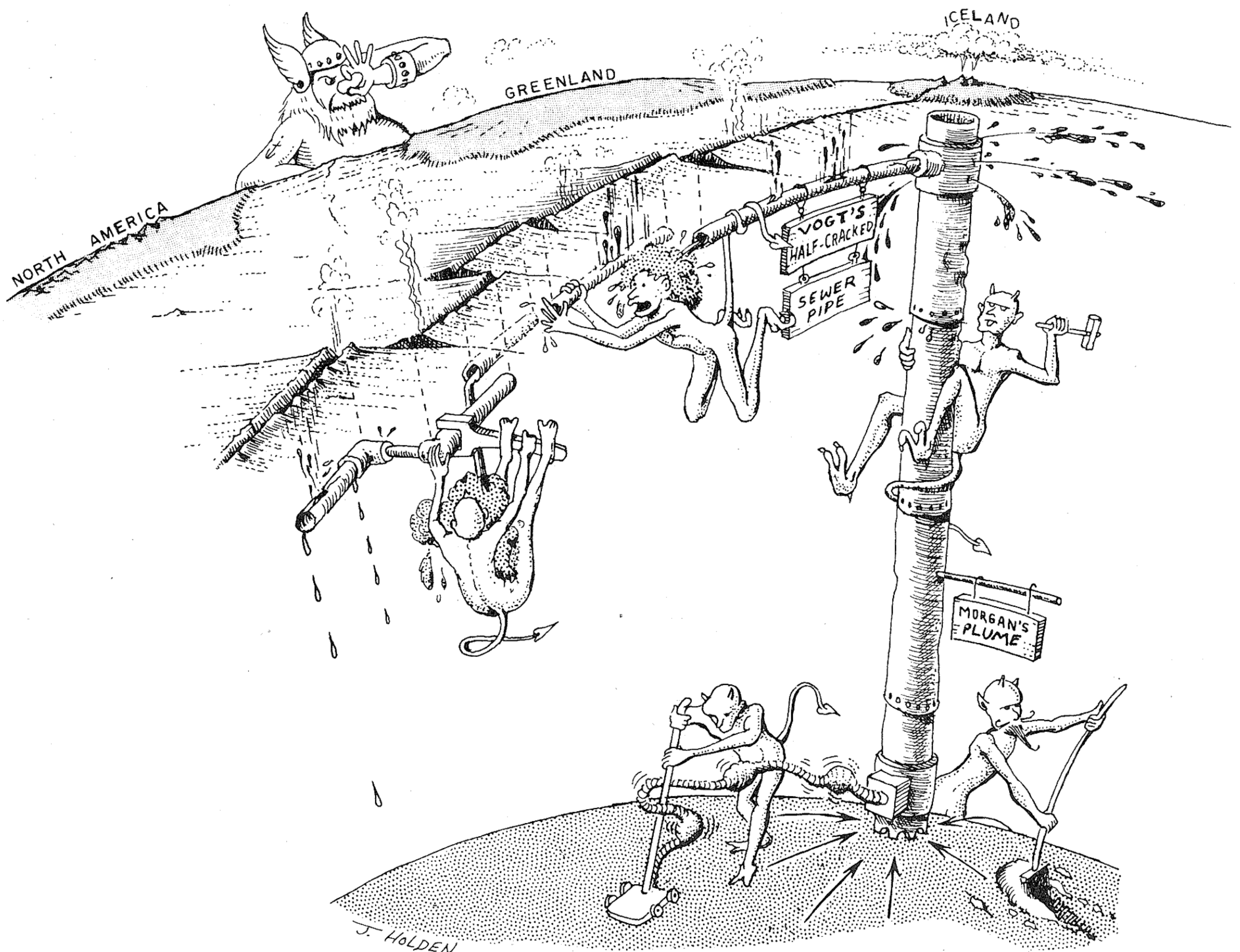


Fig. 6. Mantle plume materials transported by faulty plumbing system from the lower regions to the midoceanic ridge. (Devils not to scale.)

is attached to the nearest continent, and if it does not drive the oceanic plate, it pulls the continent over it. Because this theory accounts for the large shield volcanoes at Hawaii and elsewhere, gravitational anchors are associated with an abundance of tephra-laden hot air that often tends to obfuscate a clear understanding of these features.

On occasions when the clouds do clear somewhat, some authors can identify stress fields in the Hawaiian chain (not the anchor chain) indicating the stress of the anchor [Jackson and Shaw, 1975]. They define two stress fields, a Polynesian and a Hawaiian field. We see that a third pattern and a reassessment of all three fields yields a remarkable phenomenon (Figure 3). Carefully plotting the three lineation sets actually spells out this fact, and in Latin, no less. We misinterpret these data to read, 'We think, therefore they exist.' The mouse diagram showing trend bearings about an imaginary center marks the overall trend of the Hawaiian chain (the volcanic chain, not the anchor chain).

Plumes, antiplumes, and propagating fractures are certainly not the only possible explanations for features such as the Hawaiian chain. Ancient legend has it that the island of Japan is situated atop a giant carp and that every time the beast shifts position Mount Fuji erupts. As most legends have some basis in fact, we propose that this creature also finds its home throughout the viscous asthenosphere (Figure 4). Could it be that as it swims, at a rate of only a few centimeters per year within the mantle, it leaves behind a buoyant trail of tholeiitic bubbles which rise ponderously to create aseismic ridges? We name this fish *Asthenichthes aseismotathes*, or the asthenospheric fish that makes aseismic ridges. No doubt there are some who would question our taxonomy and would want to call the species a form of crappie. Readers may find something fishy about this theory, but then there is at least something fishy about the other theories as well.

How plumes are oriented, orifice up or orifice down, awaits future adjudication. On the other hand, we can make some definite statements

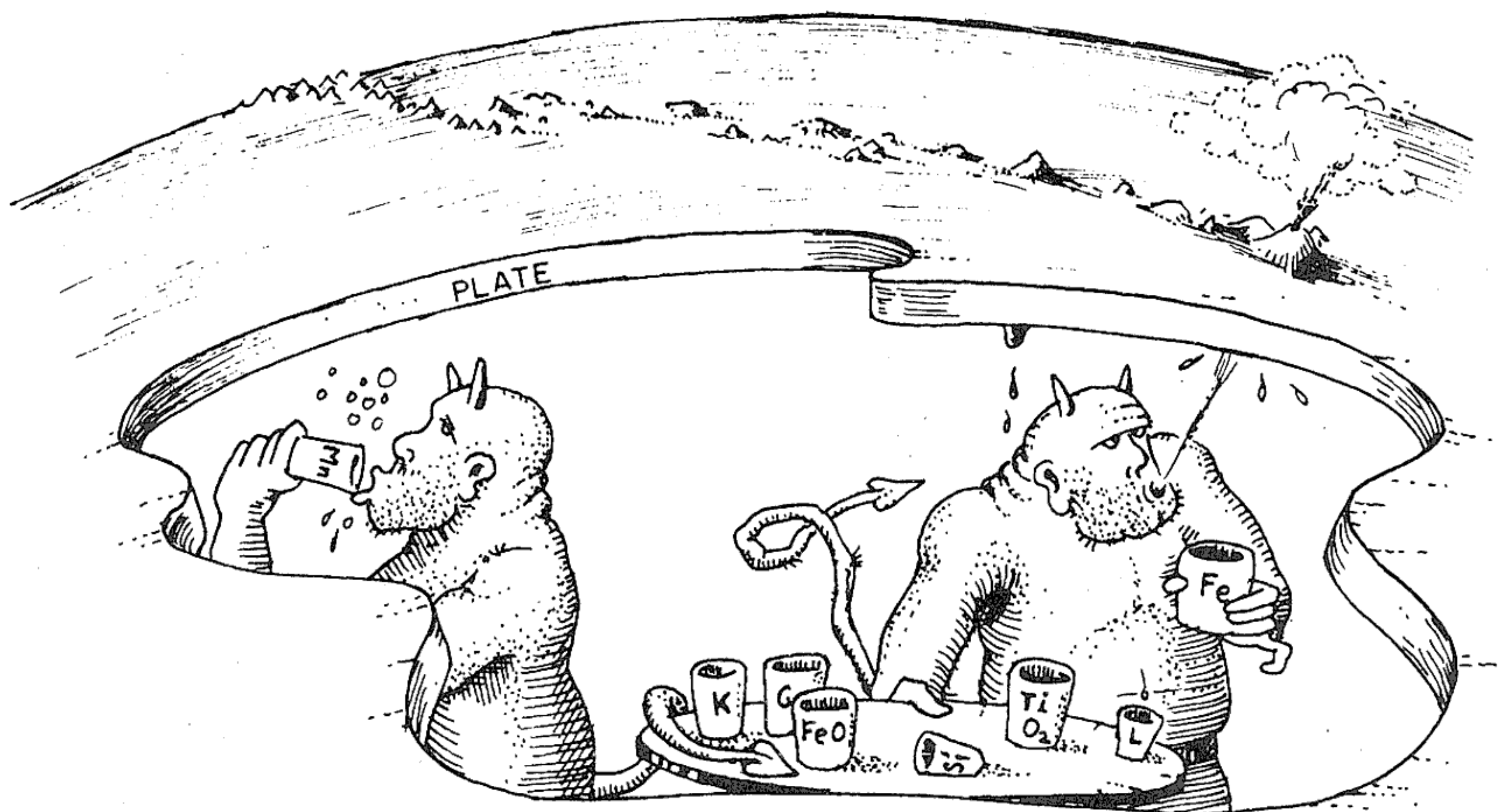


Fig. 7. Strange chemistry of ocean island tholeiites, hawaiites, balonites, and similar hot spot generated rocks, attributed to the culinary habits of X. Vulcan et al. (regurgitated material, in preparation, 1977).

about the distributions of plumes. There are, for example, two plumes on the Arctic and Antarctic circles, namely, Iceland and Balleny Islands, respectively. If Balleny Islands are, in fact, a plume, as Morgan [1972] predicted, it is nearly perfectly antipodal to the Iceland plume [Holden, 1976a]. If these two hot spots are taken as midpoints of the two world rift systems, the rifts have an interesting relationship to each other, as shown in Figure 5. One supposes that this proves the athletic excellence of the Creator, for the game is certainly horseshoes, and He has scored two perfect ringers. It would seem that this particular game has been going on for at least 200 m.y. if the rift margins of Antarctica are any indication. This continent has been located at the south pole since Pangaea broke apart, and its rift margins are all close to the Antarctic Circle [Holden, 1976b]. Unfortunately, this only accounts for two plumes; we leave it up to the reader to devise explanations for the distribution of the remaining 120.

How Many Plumes?

Mantle plumes are in the midst of a population explosion that threatens to engulf the earth in a volcanic catastrophe [Vogt, 1972]. The facts speak eloquently for themselves—no need to consult the Club of Rome. In 1971 there were 20 plumes [Morgan, 1971, 1972]; in a mere half decade the

population has risen to no less than 122 [Burke and Wilson, 1976]. Our extrapolations from these data show that there will be 1,000,000 hot spots by the year 2000. We hope someone proves that hot spots do not exist, before it is too late.

Cracked Sewer Pipes

If mantle material rises below Iceland, it must somehow spread out from the top (head) of the plume. Presumably, the flow occurs primarily in the asthenosphere below the plate. Vogt [1971, 1974] and Schilling [1973, 1975] have suggested that this flow is concentrated in the bidirectional pipe formed below the spreading axis as a result of extensive partial melting and lowered viscosity there (Figure 6). This slightly tilted 'cracked sewer pipe' flushes away the material rising in the stem of the vertical pipe.

The sewer pipe is cracked (or, more exactly, half-cracked like its author) in the sense that tholeiitic magmas must rise vertically to feed the constantly accreting oceanic crust. To further complicate matters, transform faults offsetting the spreading axis also offset the subaxial conduit, thus creating transform dams [Vogt and Johnson, 1975]. Despite all this faulty plumbing the conduit can do it, be Holden though it is to numerous mantle demons (Figure 6). Fluctuations in plume discharge down the cracked sewer pipes are thought to

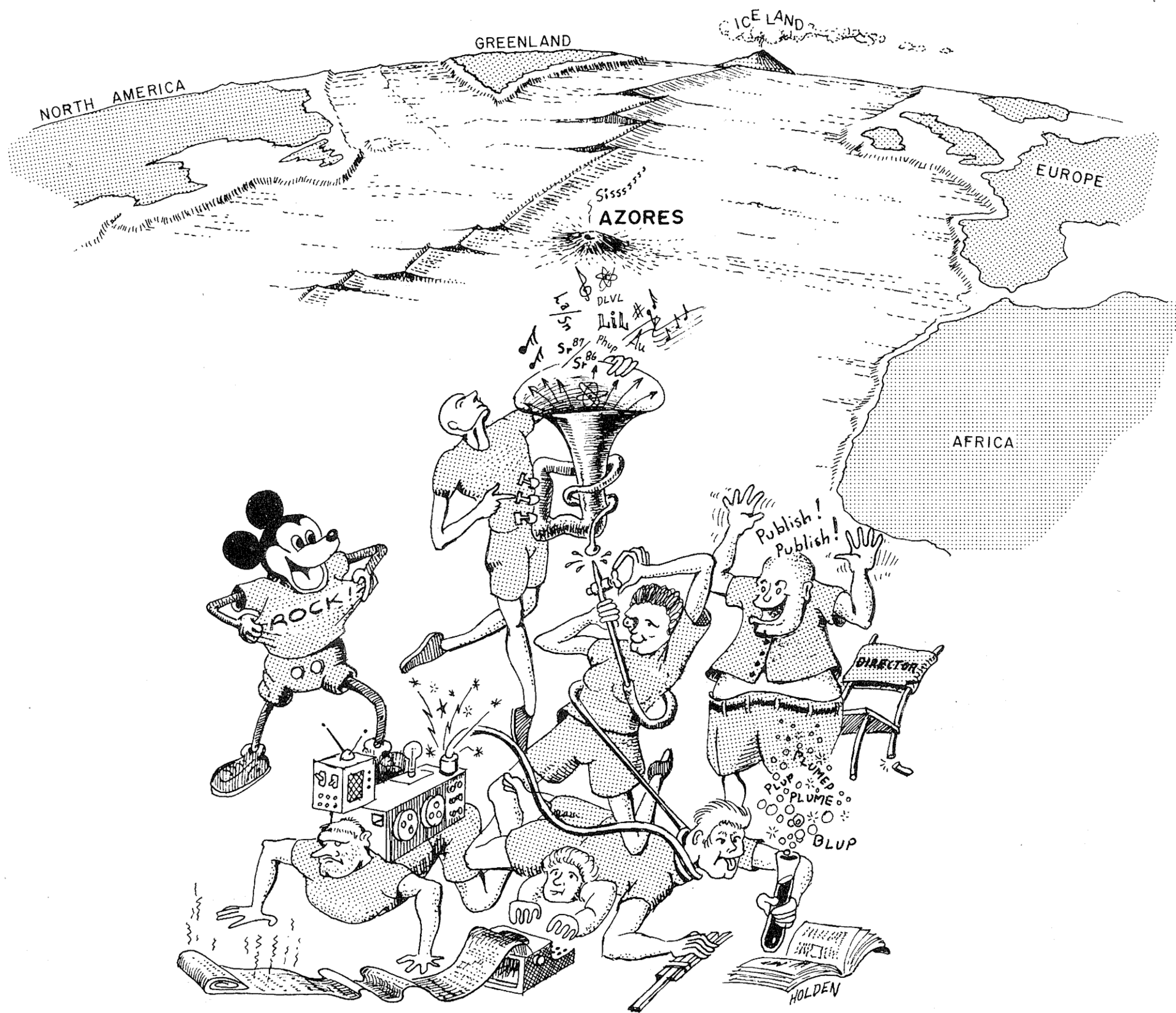


Fig. 8. Geochemical research on mantle plumes, as conducted at a modern university. The significance of the Azores, Iceland, and Mickey Mouse in this scheme is still under discussion.

leave V-shaped ridges imprinted on the ocean crust [Vogt, 1971]. Vogt is currently searching bathymetric charts to see whether there is a good spot for a P on it.

The extreme case of a fluctuating plume is the mantle blob [Schilling, 1975], a concept possibly suggested to Schilling by the behavior of a lava lamp. We expect future hot spot authors to embellish this already descriptive terminology with, for example, mantle drops and dabs and drips, mantle splats and burps and belches and spurts and...

Before leaving the intriguing though semiscatological subject of tectonic toiletometry we offer an interesting historical note. The famous early geophysicist Jules Verne

had his team descend the vertical pipe beneath Iceland in their Journey to the Center of the Earth and ascend through a volcano in the Mediterranean. This path no doubt accounts for the great difficulties they encountered, since mantle material rises at Iceland and sinks in the Mediterranean subduction zone. They were fighting against strong currents all the way. (Soon after this ordeal, Verne spent 80 days in a balloon floating around the world to recuperate.)

Plume Chemistry

It now seems that the final answers concerning the earth's largest features will be found by those dedicated research teams studying the

rarest elements and smallest subdivisions of matter.

Hot spot basalts have long been known to differ from ocean floor basalts. Even beginning Geology 1 students could take some Icelandic basalts for granite, since some of them are. Sunken continental crust was once thought to exist at depth but has since dissolved into mantle. More recently [Schilling, 1973, 1975], systematic variations in some isotope ratios and LIL (large ionic lithophile) elements along the midocean ridge away from hot spots have suggested unique PHMP (primary hot mantle plume) chemistry distinct from the DLVL (depleted low-velocity layer). Although the various PHMP's differ among themselves in LIL's, such as

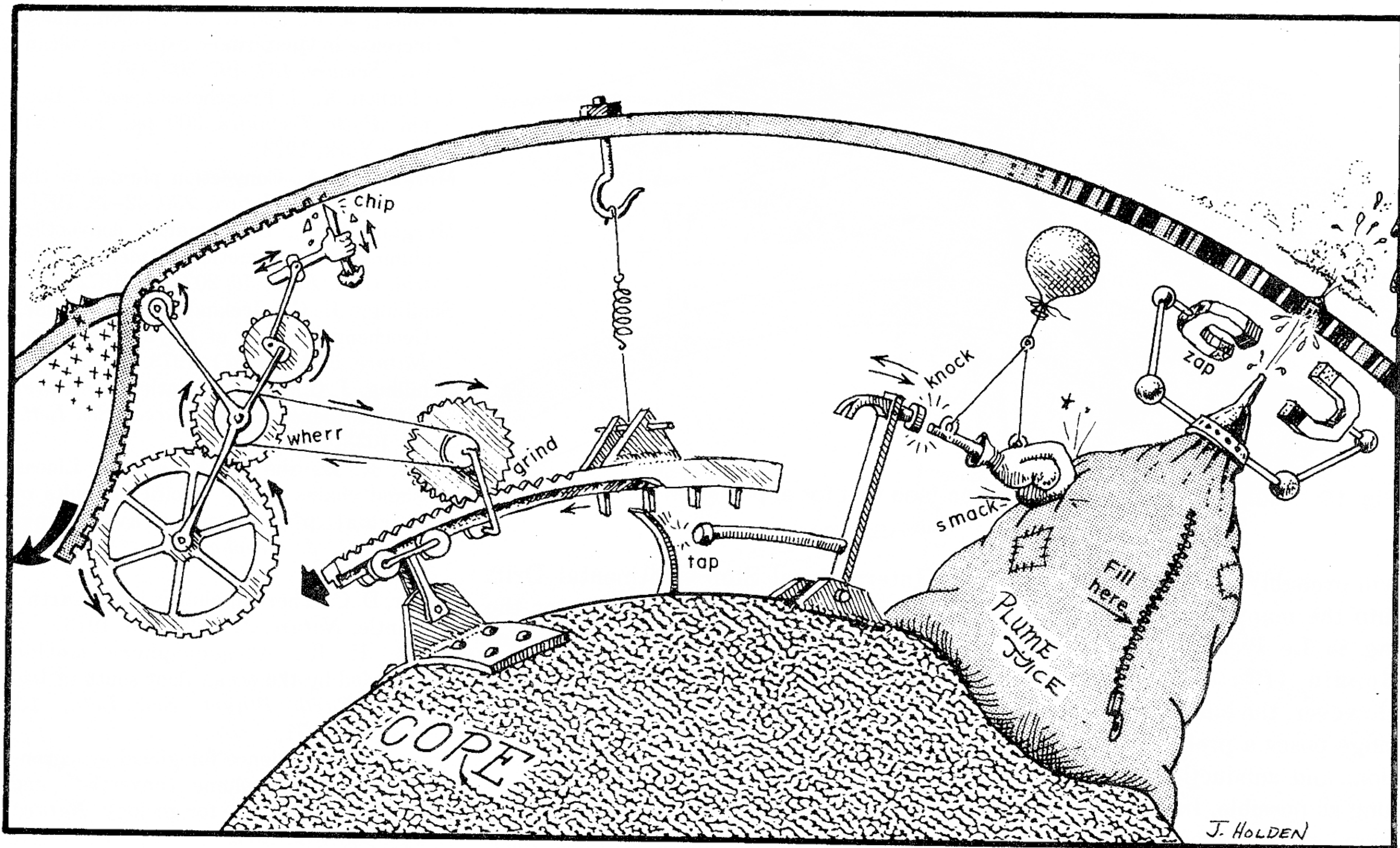


Fig. 9. Diagram hypothesizing the correlation between discharge of island arc and hot spot volcanoes. Mantle viscosity is assumed to be zero as a first approximation (with apologies to S. K. Runcorn).

La, K, P, and Ti, the DLVL always has fewer LIL's than the PHMP.

We see that hot spot basalts are really a GAS (geochemical alphameric soup) that seems diabolically difficult to understand, probably because it is the work of the UMD's (upper mantle demons; see Figure 7). The decoding of this GAS clearly requires elaborate technologies, such as mass spectrometers, neutron activators, eutecticized equilibrating quasi-partial melting fractionators, and multi-phased polyglazing computers. To operate this fancy gadgetry and produce significant results, a LAGS (large army of graduate students) is also required (Figure 8).

To make chemistry of ocean crust more accessible to the simpleminded (and penniless) layman, Vogt and Johnson [1973] have invented 'magnetic telechemistry.' In this modern version of water-witching, all that is needed is a \$2,000,000 aircraft or a \$15,000,000 research ship outfitted with a magnetometer. Iron-rich basalts associated with mantle plumes can be charted once the measurements are corrected for sunken submarines

and schools of magnetic fish. The concentrations of other elements (gold!?) can be predicted accurately, at least to within a few orders of magnitude, over at least 0.1% of the ocean floors.

Global Synchronism

For many years now, dashed correlation lines have connected short, fat magnetic anomalies with tall, skinny magnetic anomalies and even with featureless magnetic plains. But such correlation is simply a correlary of the Geologic Correlation Axiom. According to this rule, any variable can be made to correlate with any other variable once the vertical and horizontal scales are suitably adjusted, the eyeball suitably trained, and the correlation lines suitably slanted to the author's prejudices.

Vogt [1972] has applied this axiom to show that hot spot volcanoes, magnetic reversals, dinosaur populations, the stock market, and the annual number of Norman Watkins papers all follow the same global rhythm. Although further proofs of global syn-

chronism are really unnecessary, Kennett and Thunell [1975] have shown that volcanic activity along the world's island arcs has indeed fluctuated in sympathy with the hot spot discharge curve inferred by Vogt [1972]. But why should this be so? Why should volcanoes in the Marianas or the Caribbean 'know' what the Hawaiian volcanoes are doing? One answer is that mantle plumes speed up plate motions and thereby increase subduction rates and related volcanic activity. However, since many scientists still do not accept mantle plumes, we offer a yet more plausible mechanism in Figure 9. (To the reader who might question who would fill the plume juice bag, we would answer, 'Nitpicker!')

Logjam Tectonics

Having dwelt too long on the generation of aseismic ridges, we close by considering their destruction. Basalts deposited on continental crust will tend to be eroded off, along with some of the underlying intrusives. Aseismic ridges formed on ocean crust

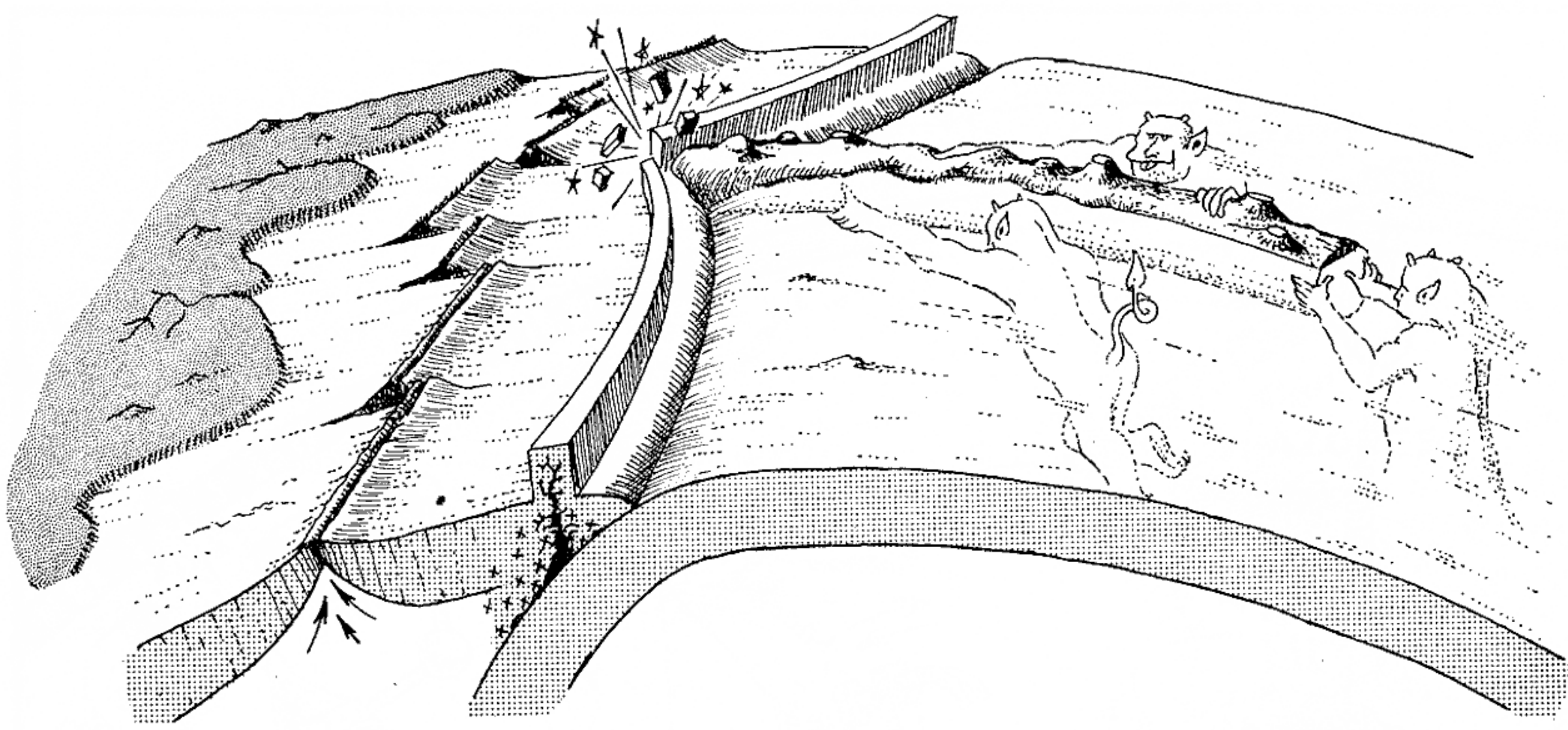


Fig. 10. Illustration of logjam tectonics (see text for explanation of symbols).

will probably eventually be subducted into the mantle (The Gospel According to Le Pichon, Francheteau, and Bonnin (*Plate Tectonics*): 7-3). However, the bulky, buoyant aseismic ridge poses a problem to plate digestion, and subduction is postponed as long as possible. In the meantime, the island arc and trench migrate toward the downgoing plate as a result of back-arc spreading. This migration is slowed or stopped where the ridge is being subducted, and (Eureka!) a cusp is formed [Vogt, 1973]. Isn't that cleaver? Figure 10 shows the battering ramifications of the pushy process postponing plate plunging.

The only other theory for island arc formation besides logjam tectonics is Ping-Pong ball tectonics. This theory holds that the earth is a giant Ping-Pong ball, its surface dented inward in numerous places to produce island arcs [Frank, 1968]. Frankly, we find any theory that likens the earth to a wet, dirty, dented Ping-Pong ball too ludicrous to debate.

Acknowledgments

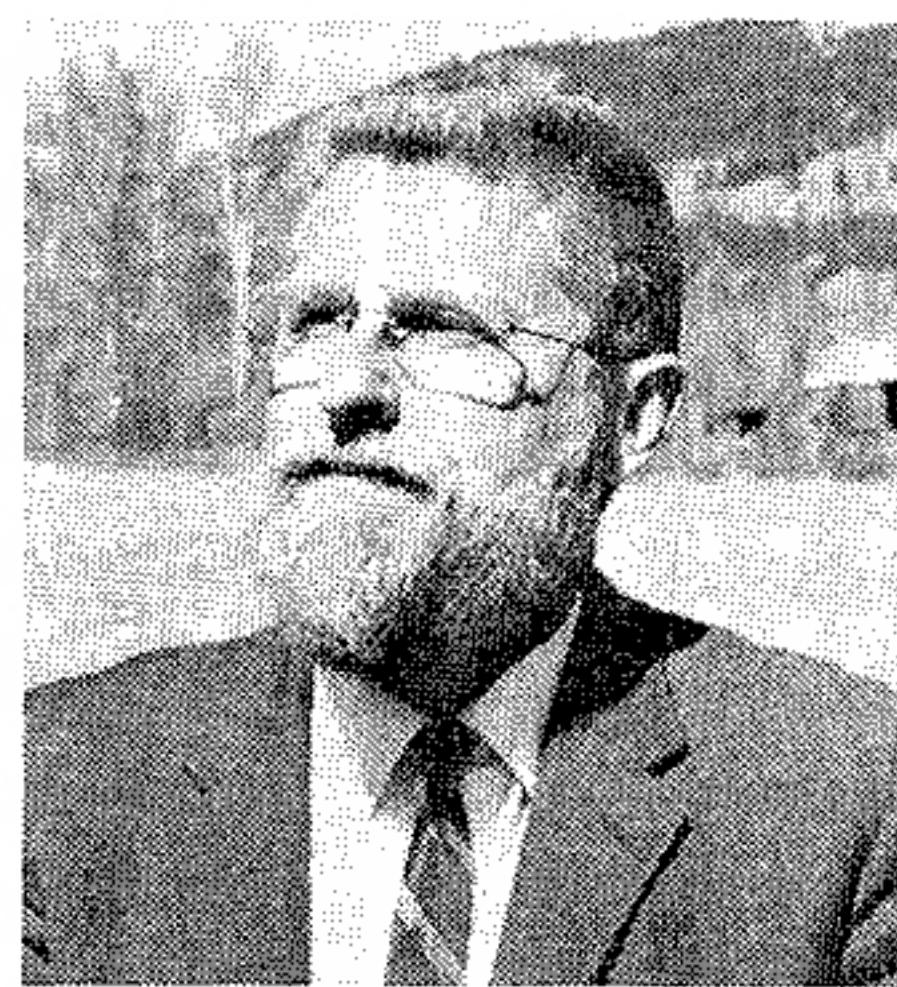
The authors completed this paper while on sabbatical at the Plumatic Asylum, Lower Slobbovian Institute of Advanced Geoquackery. P.R.V. wishes to thank the anonymous reviewers for censoring the original manuscript and AGU copy editor Joan Welsh for humor above and beyond the call of duty. He also thanks the lampooned individuals and institutions not to press libel charges. This paper does not necessarily reflect the views of the Plumatic Asylum, the Institute of Geotectonic Redundancies, the

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Peter Vogt holds all the usual academic credentials, including a valid Birth Certificate and a Third Degree, from recognized, well-meaning academic, mental, and corrective institutions. When not engaged in lively battles with the reviewers of his scientific manuscripts, he works at correlating various kinds of anomalies at the Naval Research Laboratory, Washington, D. C. Vogt is a Fellow of the Geological Society of America and a member of the ISCDS.



Words of wisdom



"It is a standing vice of geophysics not to argue against unpalatable facts and arguments but simply to ignore them and carry on as if they did not exist."

Prof. Peter Fellgett, FRS, Astronomy & Geophysics, 2003

"Of experiments intended to illustrate a preconceived truth and convince people of its validity: a most venomous thing in the making of sciences; for whoever has fixed on his cause, before he has Experimented, can hardly avoid fitting his Experiment to his cause, rather than the cause to the truth of the Experiment itself."

Thomas Spratt, "History of the Royal Society", 1667

"The traditional method of confronting the student not with the problem but with the finished solution means depriving him of all excitement, to shut off the creative impulse, to reduce the adventure of mankind to a dusty heap of theorems."

Arthur Koestler

"I cannot give any scientist of any age better advice than this: the intensity of the conviction that a hypothesis is true has no bearing on whether it is true or not. The importance of the strength of our conviction is only to provide a proportionally strong incentive to find out if the hypothesis will stand up to critical examination."

Sir Peter Medawar, "Advice to a Young Scientist", 1979

"It is all too easy to derive endless strings of interesting-looking but untrue or irrelevant formulae instead of checking the validity of the initial premises."

John Ziman, "Reliable Knowledge", 1978, p. 14

"...highly speculative, or boldly generalized theories are easily formulated, and take hold of the imagination of scientist and layman alike. Such theories may acquire widespread authority, not because they are well founded and reliable but because they have no competition from other less consensual sources of knowledge or insight. Whether or not it is eventually validated by overwhelmingly convincing evidence the 'scientific picture' presented by this sort of theory is inevitably schematic and oversimplified. The danger is that its limitations will not be adequately recognized, and that it will be extrapolated recklessly into an all-embracing dogma."

John Ziman, "Reliable Knowledge", 1978, pp. 91-92

"The voluminous literature on hypothetical plumes is notable for its ingenuity in the near-total absence of constraints."

Warren Hamilton, Precamb. Res., 1998

"When anybody contradicted Einstein he thought it over, and if he was found wrong he was delighted, because he felt that he had escaped from an error, and that now he knew better than before."

Otto Robert Frisch, on Einstein

"It was a reaction from the old idea of protoplasm, a name which was a mere repository of ignorance."

J.B.S. Haldane, "Perspectives in Biochemistry", 1938

"What is known for certain is dull. I rarely plan my research; it plans me."

Max Perutz

"It takes many years of training to ignore the obvious."

The Economist on "Theories of Economic Growth"

"Whether true or false, others must judge; for the firmest conviction of the truth of a doctrine by its author, seems, alas, not to be the slightest guarantee of truth."

Charles Darwin, letter to Lyell, 25th June, 1858

"In fact, no opinion should be held with fervour. No-one holds with fervour that $7 \times 8 = 56$, because it is known that this is the case. Fervour is necessary only in commending an opinion which is doubtful or demonstrably false."

Voltaire, quoted by Bertrand Russell

"Great God, how can we possibly be always right and the others always wrong?"

Montesquieu, Cahiers

"We see that many assumptions used in previous hypotheses can be discarded as unnecessary. ...there is no need to locate the source of plumes in the lower mantle."

Richter & Parsons, 1975

"Finding the world would not accommodate to his theory, he wisely determined to accommodate the theory to the world."

Washington Irving

"Every dogma must have its day."

H.G.Wells

"Convictions are more dangerous enemies of truth than lies."

Nietzsche

"As soon as I hear 'everybody knows' I start asking 'does everybody know this, and how do they know it?'"

Dave Jackson, from J. Fischman, "Falling into the gap", Discover, 58-63, October, 1992

"There is something fascinating about science. One gets such wholesale returns of conjecture out of such trifling investment of fact."

Mark Twain, "Life on the Mississippi", 1883

"Words, as is well known, are the foes of reality."

Joseph Conrad, "Under Western Eyes", 1911

List of Abstracts

Citing these abstracts: The approved citation style for these abstracts is illustrated by this example:

Anderson, D.L., Plate Tectonics: The General Theory, abstract in The Hotspot Handbook, Proceedings of Penrose Conference Plume IV: Beyond the Plume Hypothesis, Hveragerdi, Iceland, August 2003.

Don L. Anderson, California Institute of Technology, USA

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Plate Tectonics: The General Theory

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The Emperor and Hawaiian volcanic chains

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Lithospheric control on silicic magma generation associated with the Ethiopian flood basalt province

WHAT IS A PLUME ?

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A plume is often invoked, as an explanation of magmatism, because of some perceived difference with “normal” volcanism. This difference can be volume, chemistry or tradition. In discussing the origin of “melting anomalies” it is necessary to have precise definitions. Workers in different fields have different ideas of what constitutes a plume so it is important to agree on the concept.

normal volcanism

Spreading ridges involve pressure-release melting. As ridges spread the space is filled with adiabatically upwelling mantle which is close to or above the melting point. The release of pressure causes melting or an increase in the melt content.

Island arc volcanism is attributed to reduction of the melting point by the release of volatiles from the downgoing slab. Melting is not entirely due to the increase of temperature or the release of pressure but by reduction of the melting point.

Hotspot volcanism, by contrast, is attributed to locally high temperatures.

melting anomaly

Excess or long-lived volcanism. This can be due to wet or fertile mantle (compared to normal mantle), focusing, small-scale convection, or high-temperatures. The plume hypothesis focuses on the high-temperature explanation. Some of the other mechanisms are athermal - that is, melting anomalies can be generated from normal temperature mantle.

plume

A narrow thermal feature, which can be either hot or cold, which rises or sinks because of its anomalous temperature compared to the surrounding fluid. In fluid dynamics a jet has the same meaning. A plume, or jet, arises from the instability of a thermal boundary layer, which is heated from below or cooled from the top.

mantle plume

A hot narrow buoyant upwelling rising from deep in the mantle and generally attributed to thermal instability of a thin layer near the core-mantle boundary (CMB). In Earth sciences a plume is also defined as a form of convection independent of other kinds of convection or plate tectonics. Plumes are considered to be the way the core gets rid of its heat, while plate tectonics is defined as the way the mantle gets rid of its heat.

Properties of Plumes

Plumes are hot. The primary thermal diagnostics are temperature, heat flow, uplift, and thermal erosion of the overlying lithosphere. Normal variations of potential temperature associated with plate tectonics are of order 200 degrees or more. The core is hotter than the mantle and the thermal boundary layer at the base of the mantle involves a larger temperature change than the one at the top. Deep thermal plumes are expected to have temperature excesses of more than 300 degrees compared to normal upper mantle basalts. Plumes have been predicted to cause thermal uplifts of 1 to 2 kilometers prior to volcanism. The heat flow at midplate plumes should be equivalent to very young oceanic lithosphere.

Plume heads. Injection experiments show that a large bulbous plume head is required to start a plume from deep in the mantle. There should be a one-to-one correspondence between a proposed hotspot track and a large igneous province (LIP) and this LIP should be generated at high elevation. "Plume Head" basalts should be colder than OIB.

Plume heads spread out. Tomography and heat flow should reveal slow seismic velocities and thermal anomalies over at least a 1,000 km radius in the upper 100-200 km. Of the mantle under proposed plume sites.

False Plume Proxies

Some characteristics of provinces with magmatic anomalies have been taken as proxies for plumes. These include non-MORB geochemical characteristics, high $^3\text{He}/^4\text{He}$ ratios, rapidity of extrusion, volumes of basalt and crustal thickness. Some geochemical models predict high ^3He contents and high $^3\text{He}/^4\text{He}$ ratios have been taken as a proxy for that.

Tomography

Seismic anomalies in the lower mantle are sometimes related to surface features. Continuity to the surface must be demonstrated. Statistical methods, such as Monte Carlo, should be used to confirm that the coincidences (between, say, a deep low-velocity-zone and a surface volcano) occur at a higher level than random chance.

Geochemical Diagnostics

The best chemical diagnostics would be those that have something to do with interaction with the core, at the appropriate time. Much of the mantle has probably been in equilibrium with molten iron at some point in the accretion/differentiation process (except in the extreme inhomogeneous accretion models). But if some magma was in contact with the core less than 500 myr ago this would be a good diagnostic of a deep upwelling. On the other hand if the upper mantle trace siderophiles occur in chondritic proportions this would indicate isolation of the deep mantle (the irreversible chemical differentiation model). Recycled materials in the mantle are intrinsic to plate tectonics and do not imply a deep or plume origin.

Characteristics of Melting Anomalies

The accompanying table summarizes the physical and chemical characteristics of many proposed hotspots. Many of the proposed Primary Plumes do not have the characteristics most closely associated with thermal anomalies.

Table: Summary of candidate plume-diagnostic observations. PDF viewers: Expand screen magnification to at least 200% for optimal viewing.

Assumptions & Fallacies

“ The method of postulating [assuming] what we want has many advantages. They are the same as the advantages of theft over honest toil.”

(Bertrand Russell, Introduction to Mathematical Philosophy)

Among the more critical assumptions that have been made in developing the plume hypothesis are:

- “normally” the mantle is below the melting point
- melting anomalies are due to localized high temperature (not low melting point)
- the mantle is almost isothermal (adiabatic)
- cracks will not be volcanic unless the local temperature is anomalously high
- high temperature requires importation of heat from the CMB in the form of narrow jets.
- the upper mantle is vigorously stirred and is chemically homogeneous.
- steady state - hotspots are supplied by a steady stream of deep mantle material (rather than tapping melt lens that have accumulated over long periods of time).
- steady state - plate tectonics is steady state and one does not expect more magmatism or different components at the onset of seafloor spreading.

These assumptions need to be continuously tested. The proximity of the upper mantle to the melting point and the variable fertility of the mantle due to plate tectonic processes, may call into question the validity of some of these assumptions and may make the plume hypothesis unnecessary. Evidence now used in support of plumes includes the absolute volume of erupted basalts, the rapidity of eruption, the chemistry of the magma, the elevated helium isotope ratios of some of the basalts at some hotspots, and the observation that inferred hotspot tracks cross ridges.

The most convincing arguments for a “hotspot” or a plume would be high magma temperature, uplift, thick crust, high heat flow, thermal erosion of lithosphere, or a deep mantle tomographic signal. These are indicators of a thermal mechanism, as opposed to athermal mechanisms which have also been proposed for oceanic plateaus, swells and CFB. Athermal mechanisms include focusing, fertility, ponding, the edge and rift mechanisms, and mechanisms involving lithospheric stress and dikes, and a partially molten shallow mantle. The time element (transients, long-term ponding), the stress element (lithospheric valves) and the fertility element (recycled crust, volatiles), in many respects, serve as substitutes for high temperature.

The absolute amount of magma is often used as an argument supporting plumes but, usually, no comparisons with other mechanisms are made. For example, ridges also produce large quantities of basalt and do so for much longer periods of time. Focusing, ponding and edge-driven effects can increase rates, for short periods of time. The current eruption rates at Hawaii are certainly impressive but prior to 6 myr ago, the output of the Emperor-Hawaii chain were not impressive. Most “hotspot” tracks are only active for 15 myr or so. CFB are transients and 3D while most ridges are steady state and 2D. These factors alone increase eruption rates and volumes by large factors over ridges, with no increase in temperature. Also, some plateaus clearly have a continental base and are not entirely recent features as often assumed. Other processes that can give results similar to plumes are small-scale convection, an intrinsic part of plate tectonics, and convection induced by lithospheric architecture (corner flow).

Stress-controlled rates (the lithospheric valve) and fertility and volatile variations in the shallow mantle (all athermal mechanisms) received a boost from these calculations. Large volumes, and large eruption rates, especially if only temporary, can be caused by decrease in melting point, increase in basalt content of the shallow mantle (the recycling mechanism), increase in volatile content, edge and rift induced convection, and focusing. High temperature alone does not seem to be adequate.

Arguments & Counterarguments

What are some of the arguments (A), and counter arguments, (C) used in support of a mantle plume as the standard model for the origin of flood basalts?

A1. Exceptionally large volume of tholeiitic magma.

C1. One must compare the observed rates with something. The absolute value by itself means nothing, particularly since it is trivial, and not long-lived, compared to the output of ridges, island arcs, and backarc basins.

A2. Flood basalts are erupted in an extremely short time.

C2. The short time actually implies stress or lithosphere control, a valving action. Plume theorists have shown that in the plume model the timescale is controlled by the viscosity of the deep mantle and they get time scales of 10 myr or longer. A stress mechanism can be instantaneous. The transient nature of magmatic bursts also suggests pre-eruptive ponding. Global synchronism of volcanism seems to favor a stress explanation, one that involves a global plate reorganization.

A3. Huge volumes and high eruption rates are unique to continental flood basalt provinces. As such they appear to require a unique tectonic/magmatic event.

C3. This unique event can be a change in stress or a plate reorganization. The EDGE and rift-induced convection mechanisms are, by nature, episodic, and flux rates vary enormously so there may be no “event”, just as no causative event is responsible for a continent-continent collision or a ridge-trench annihilation or variations of eruption rates along volcanic chains (although these MAY be caused by stress variations). A mantle plume is often assumed necessary to get the volumes and rates, assuming a steady-state mechanism.

A4. *From the prospective of flood basalt provinces, no other model appears to provide for the unique volumes and eruption rates of these large magmatic provinces.*

C4. The recent literature seems to offer viable alternatives (following up on earlier suggestions of a partially molten asthenosphere, a fertile source (eclogite, piclogite), focusing, EDGE convection, continental insulation (midplate mantle is warmed up), refertilization of the shallow mantle, melt ponding, and ultimate release by stress control, diking and so on. These must be tested.

A5. *These huge eruptions can be shown to frequently occur at the beginning of a long trail of lesser eruptions which end at a currently active volcanic center. The Deccan is by far the best example of this correlation.*

C5. Fewer than half the LIPs have even a postulated tail and the most prominent examples are contentious

A6. *Hotspots often cross ridges, showing that a fixed plume underneath the plate is responsible.*

C6. Plate reconstructions based on the fixed hotspot assumption have this feature but other plate reconstructions do not show ridges crossing hotspots. The association of some linear volcanic features (e.g., 90 E. ridge, Chagos-Laccadive Ridge) with CFB has been used to assert that the LIP is now separated by a ridge from the hotspot. These associations have been disputed by other plume specialists.

A7. *A line of evidence in support of the mantle plume - hot spot model for the origin of continental flood basalt provinces lies in the composition of the magmas. Here, again, the evidence is not unambiguous and certainly does not prove the existence of mantle plumes. But it does fit the mantle plume model. Some eruptions contain those elevated helium isotope ratios that are equated with an origin deep in the mantle.*

C7. This involves the circular argument that, because Yellowstone, Hawaii and Iceland are products of a hot spot then elevated helium isotope ratios must be produced in the lower mantle. High $^3\text{He}/^4\text{He}$ ratios are found in many places but when found they are attributed to deep mantle plumes. In all case the absolute ^3He abundances are orders of magnitude less than in MORB. In other words, by definition, the elevated ratios are from the lower mantle. The only reason elevated helium ratios were associated with plumes in the first place was because Yellowstone, Iceland and Hawaii had some high ratios and they were thought to be plumes. This again is a circular argument.

A8. *The problems with the thermal plume idea can be fixed up by adopting aspects of the chemical plume idea. A more iron-rich or fertile source has long been advocated and supported by experimental evidence. An eclogite-bearing mantle plume source derived from subducted ocean basalts recycled through the deep mantle appears to satisfy this requirement and to satisfy the trace element concentrations.*

C8. The chemical plume idea, and the eclogite and recycled crustal source idea are old ones and are alternative mechanisms to deep hot plumes. Introduction of eclogite into a plume was thought to be necessary to get the observed volumes but when this is done one no longer needs the plumes or deep recycling. If the shallow mantle is close to the solidus of peridotite it will be near the liquidus of eclogite and melting anomalies can be created at “normal” mantle

temperatures. A shallow fertile source is one alternate to plumes, and may give the necessary volumes at low T, especially if combined with the ponding/stress-release idea.

A9. There is lack of geologic evidence for extension prior to eruption of CFB.

C9. There is abundance evidence for extension, but usually not uplift (the plume diagnostic), prior to volcanism. Dikes can also take up extension. Only 1 cm of extension, with magma viscosities, is all that is needed to provide the volumes and rates from a fertile and partially molten mantle. Meter wide dikes can certainly provide the necessary flow rates and this can be below geologic resolution for extension.

Fallacies

Some of the common arguments in support of particular models of mantle convection or geochemical box models can be cast into the form of logical arguments and analyzed for their validity. Some well known fallacies are categorized below, with examples from the recent literature.

Circulus in demonstrando

Mid-ocean ridges are able to migrate over hotspots, which implies that the hotspot source is deeper than about 200 km.

(plate reconstructions not using the fixed hotspot reference frame do not demand that ridges cross hotspots)

argumentum ad populum

“For many geoscientists, the mantle plume model is as well established as plate tectonics”.

False Dilemma and Affirming the Consequent, plus rhetoric and Bifurcation

“The apparent controversy can be broken down into two questions. Is there evidence that deep mantle plumes exist? And do all volcanoes not associated with plate boundaries require a deep mantle plume? The answers seem most likely to be “yes” and “no” respectively.”

(A limited number of options (usually two) is given, while in reality there are more options (or perhaps even one). A false dilemma is an illegitimate use of the “or” operator. Putting issues or opinions into “black or white” terms is a common instance of this fallacy.)

The actual question is, “Is there evidence that any volcano requires a deep mantle plume?” Deep mantle plumes, in the sense of thermal instabilities of a thermal boundary layer, certainly exist but do they rise to the surface and are they narrow? Pressure (and chemical layering) at the CMB causes them to be broad, sluggish, long-lived and slow to form, apparently consistent with the large features seen by tomography. The probable existence of a deep mantle TBL is not the same as the assumption that these must be the source of OIB. The discovery of deep mantle low-

velocity zones is not evidence for connection to the surface; even a chemically stratified mantle will have variable temperature (and composition?) in each layer.

Bifurcation

Also referred to as the “black and white” fallacy and “false dichotomy”, bifurcation occurs if someone presents a situation as having only two alternatives, where in fact other alternatives exist or can exist.

Red herring and fallacy of Irrelevant Conclusion

The upwelling mantle under Hawaii must also be 200-300 K hotter than the surrounding mantle to achieve the required large melt fractions at depths below the 80-km-thick lithosphere. Such hot rock material **must** come from a thermal boundary layer. The CMB is the **most likely** source, unless there is another interface within the mantle between compositionally distinct layers.

(these are requirements of the steady-state thermal plume hypothesis, not general requirements).

The Ratio Fallacy and the Slippery Slope Fallacy

The chemistry and isotopic composition of many hotspot lavas, especially the high $^3\text{He}/^4\text{He}$ ratios, indicate that the hotspots sample a part of the mantle distinct from that sampled by mid-ocean ridge basalts. High $^3\text{He}/^4\text{He}$ ratios imply high ^3He contents and therefore an ancient undegassed reservoir and therefore the deep mantle.

Fallacy of Irrelevant Conclusion

Numerical simulations of plumes reproduce many of the geophysical observations, such as the rate of magma production and the topography and gravity anomalies produced by plume material as it spreads beneath the lithosphere. Therefore, plumes exist.

Modus Moron

Mid-ocean ridge basalts come from the upper mantle.
Therefore, ocean island basalts come from the lower mantle.
Plumes come from the lower mantle.

Fallacy of Irrelevant Conclusion, Affirming the Consequent and Permissivity

Theoretical and laboratory studies of fluids predict that plumes should form in the deep Earth because the core is much hotter than the mantle. Therefore hotspots are caused by plumes from the CMB.

(confusion of “should” with “do” or “must”.)

Ignoratio Elenchi and Circulus in Demonstrando

The persistence of flow through the plume tail for 100 myr or more (several times the number of years required for plume heads to rise through the mantle) implies that the plume is much less viscous than the surrounding mantle.

(this has nothing to do with whether plumes exist or the characteristics and requirements of other models).

Continental flood basalts erupt a million cubic kilometers of basalt or more in 1 myr or less. Therefore plumes erupt a million cubic kilometers of basalt or more in 1 myr or less.

(this characteristic is now used to prove that continental flood basalts are caused by plumes).

The above two conclusions are contradictory. The rate of plume magmatism is controlled by lower mantle viscosity while in the plate theory it is controlled by lithospheric stress (the valve).

Table: Summary of candidate plume-diagnostic observations. PDF viewers: Expand screen magnification to at least 200% for optimal viewing.

Plate Tectonics: The General Theory

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OVERVIEW

The standard model of mantle dynamics and chemistry involves a complex interaction between plates and plumes, and the upper and lower mantles. This model requires many assumptions and produces many paradoxes. The problems and complexities can be traced to a series of unnecessary and unfruitful assumptions. A simpler and more general hypothesis is described which is based on convective systems which are cooled from the top and relies on a theory of plate tectonics that is not constrained by assumptions about absolute plate rigidity, hotspot fixity, mantle homogeneity and steady-state conditions. Plate tectonics causes thermal and fertility variations in the mantle and stress variations in the plates, thus obviating the need for extraneous assumptions about the deep mantle and a central role for core heat. The general theory of plate tectonics is more powerful than the current geodynamic models which, by assumption, exclude incipient plate boundaries (volcanic chains) and athermal explanations (e.g. fertility variations, focusing) of melting anomalies. Important concepts in the general theory are plates, cooling from the top, the surface thermal boundary layer, self-organization, stress, pressure, the central limit theorem, ponding, the end games, recycling and fertility variations. In contrast, the plume hypothesis is a strictly thermal theory and core heat is the essential element. A brief overview of the general theory of plate tectonics is provided. The bottom line is that, if the average temperature of the upper mantle is close to the solidus, as seems probable, then normal variations in mantle temperature, fertility and volatile content (consequences of plate tectonics) will generate “melting anomalies” without the need for deep mantle plumes. The localization and rates of these melting anomalies are controlled by the lithosphere. The total volumes are controlled by the fertility of the mantle.

BACKGROUND

Plate tectonics has been one of the most successful theories in the history of the natural sciences and has revolutionized thinking in all of the Earth sciences. Yet it remains a largely descriptive and kinematic theory, albeit a remarkably predictive one. It is, however, much more powerful than generally acknowledged. Plate tectonics causes thermal and chemical inhomogeneity in the mantle, and the forces that drive the plates also cause them to deform and break. Overlooked aspects of plate tectonics are the necessity for ridges and trenches to migrate, for triple junctions and boundary conditions to evolve and for plates to interact and to reconfigure when the boundary conditions change. Second order features of plates and plate boundaries (fracture zones, accreted terranes, transform faults, broad diffuse zones, swells, sutures and microplates) and boundary reorganizations are actually intrinsic and provide the key to a more general view.

In systems cooled from above, the instability of the surface layer drives the motions of both the surface and the interior and this is the kind of convection involved in plate tectonics and the thermal evolution of the Earth. Yet it is motions of the mantle, and temperature variations in the deep mantle, independent of the surface conditions that is often assumed to drive the plates and create volcanic chains. In one theory the plates control their own fate; in the other many surface features are controlled by deep convective motions, core heat and the surface passively responds or at most is just the surface boundary layer of a system where the bottom boundary layer is as important as the top boundary layer, in spite of the effects of pressure and sphericity.

PLATES

The word 'plate' implies strength, brittleness and permanence and often the adjectives rigid and elastic are appended to it. But plates are collages, held together by stress and adjacent portions rather than by intrinsic strength. Plates break at suture zones (former plate boundaries), fracture zones and subplate boundaries, usually generating volcanic chains in the process. The plume hypothesis assumes that continents break up along lines that have been pre-weakened by underlying plumes. In the general theory of plate tectonics, plate architecture, fabric and stress replace high temperatures, and the igneous activity at new plate boundaries (or reactivated old ones) is a result of breakup, not the cause.

MANTLE TEMPERATURE

The first thing to realize about hotspots is that they are not hot. It is important to realize that they are not, strictly speaking, spots either, but it is easiest if you try and realize that a little later, after you've realized that everything you've realized up to that moment is not true.

A series of recent petrology and geophysics papers is elaborating on the theme that hotspot mantle is not particularly hot (see www.mantleplumes.org for references). The idea that the shallow mantle may have variable contents of eclogite or piclogite, and volatiles is also undergoing a revival. Variable fertility of the mantle is an alternate to high temperatures in the explanation of melting anomalies. When viewed in concert with recent papers by the Morgans, the Steins, John Tarduno, Marcia McNutt and Paul Wessel, and their colleagues, on the non-fixity, non-age progression, no heat flow anomaly and non-parallelism of hotspot tracks (actually, volcanic chains) this means that the last evidence for the thermal plume hypothesis has now been removed.

Numerous ad hoc amendments have been made to the plume hypothesis (polar wander, lithospheric drift, tilted conduits, multiple plumes along a track, large radius of influence, long distance lateral flow, group motions, numerous changes in Euler poles etc.) and it may be time to consider alternatives such as propagating cracks, leaky transforms, self-propagating volcanic chains, dikes, small-scale convection etc. which may explain the data without constant ad hoc adjustments. It can be recalled that the original crack and stress hypotheses of Jackson, Shaw, Oxburgh, Walcott etc. were abandoned because they did not explain fixity, parallelism, high temperatures or excess magmatism (see the review in the USGS Professional Paper on Hawaii,

Vol. 1). Many plume hypotheses (e.g., McKenzie, Campbell) assume an isothermal and subsolidus mantle, and the more successful ones require a deep origin (CMB) and high temperatures (1600 - 1650 C). The mean temperature of the mantle and its variability are obviously key issues.

The assumptions that the upper mantle is cold, subsolidus, dry and homogeneous (unless invaded by a plume) can no longer be supported. Nor can the assumption that 'hotspots' are substantially hotter ($> 200^{\circ}\text{C}$) than other parts of the mantle, or that 'excess' magmatism uniquely requires excess temperature. Proposed plume temperatures generate too much melting. The corner flow calculations of Keen, Korenaga, King etc. generate large volumes (LIPs, Iceland...) and rates, even from normal temperature or isothermal mantle. Thus, fertility and geometry can both generate large volumes of melt with not thermal anomaly.

The volume of melting depends on source fertility, composition, focusing, history and EDGE effects (dynamic convection and fluxing of upper mantle through the melting zone by horizontal temperature gradients) as well as temperature. Lithospheric architecture and stress may be more important than localized high temperature areas. No hot thermal boundary layer is required for surface volcanism but may participate in lower mantle dynamics.

The ideas that OIB may be H_2O and FeO (K_2O , eclogite...)-rich compared to MORB are not new but it is still generally perceived that 'hotspots' are hot and have something to do with deep convective instabilities in thermal boundary layers (rather than due to dikes in extending lithosphere). Usually, crustal thickness has been used as a proxy for mantle temperatures but there are now more direct estimates. The plume hypothesis is an extraordinary one, using core heat as a geologic agent. Feynman said 'extraordinary hypotheses require extraordinary evidence'. Is there any remaining 'extraordinary evidence' that favor deep mantle plumes over more mundane plate tectonic explanations?

THE OPENING AND END GAMES OF PLATE TECTONICS

Uniform spreading ridges and subduction of old dense lithosphere are aspects of steady-state plate tectonics that may not be appropriate for the opening and closure stages of oceans. The opening stage is often associated with bursts of magmatism (Large Igneous Provinces, Seaward Dipping Reflectors) that are generally attributed to plume heads. But lateral temperature gradients, tensile stress, EDGE (Edge Driven Gyres and Eddies) convection, dynamic convection and focusing can be important at this stage. In addition, break-up at sutures may involve particularly fertile and low melting point mantle, the remnants of young trapped oceanic plates.

Fragments and isotopic signatures of crust, lithosphere and sediments have been identified in oceanic samples and there is no requirement or evidence that these were embedded in dense oceanic plates or deeply recycled and returned to the shallow mantle from the core-mantle boundary. It is the assumption that the upper mantle is uniform and homogeneous that has led to these kinds of models. The homogenization of magma from diverse lithologies in magma chambers beneath the axis of the midocean ridges provides an alternate explanation for the

apparent homogeneity of midocean ridge basalts. This is a consequence of sampling theory, in particular, the central limit theorem, a well known result in statistics.

The endgames of plate tectonics are particularly interesting. When continents collide, and oceans close, young buoyant lithosphere and crust is inserted into the shallow mantle and some is caught in the suture. When continents break up, usually along old plate boundaries, the accessible material is different than is available in the middle game of plate tectonics, the more steady phase of the supercontinent cycle.

STRESS

Plate boundaries change their configurations very slowly except at times of global plate reorganizations. Plates themselves are even more constrained to change their motions – velocities and directions – slowly. The surface stress reference system therefore changes slowly. Since regions of extensile stress control the locations of magma ascent – volcanoes and dikes – then a nearly fixed reference frame is predicted without anchoring volcanoes to a deep immobile layer. Plate boundaries form as the result of the stress system which involves boundary and body forces. When the boundary conditions change the stress system changes and the plates must reorganize. This involves creating new plate boundaries (volcanic chains, mountain belts) and closing down old ones.

The interesting thing about this view of plate tectonics is that a few simple rules control the evolution of the system. Self-organization does not require templates or fine tuning; it takes care of itself. The creation of new plate boundaries does not require thermal weakening (plume tracks).

THE DEEP SLAB DIVERSION

I suggest that most of the chemical heterogeneity of the upper mantle is due to subduction of sediments, fluids, crust and plates of various ages, including young plates which either get overridden by continents, experience flat-subduction or are quickly warmed up to a condition of neutral buoyancy. Only thick old oceanic plates achieve enough negative buoyancy to sink rapidly through the upper mantle but even these may contribute their fluids, and even parts of their crusts, to shallow mantle heterogeneity. I suggest, however, that much of this recycling is shallow and involves oceanic crust that was young upon subduction. This buoyant material can be isolated for long periods of time in the shallow mantle. Shallow chemical heterogeneities comprised of oceanic crust or lithosphere can also cause topographic anomalies (of both signs) and fertility and melting point variations, without thermal anomalies. The materials introduced into the mantle at subduction zones and through cracks provide part of the material subsequently reprocessed at ridges and island chains.

WRAP-UP

Plate tectonics is far from being just a kinematic theory that requires extraneous theories to explain its existence and so-called ‘midplate’ phenomena. It is a self-contained and complete theory of geodynamics, mantle geochemistry, and mantle thermal and chemical structure and evolution. It implies the creation and migration of plate boundaries, the growth and shrinkage of plates, and global reorganizations. Mantle convection is a result of topside cooling, plate tectonics and of lithospheric architecture and motions. Lateral variations in mantle temperature, melt volumes, chemistry and “age” are an inevitable result of plate tectonics.

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References and further background are to be found at www.mantleplumes.org

Evidence for intermediate composition in bimodal basalt-rhyolite large igneous province

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ABSTRACT

The rhyolitic ignimbrites from Were Ilu area are unlike most of the ignimbrites from Ethiopian CFB province, being contained phenocrysts and microphenocrysts of plagioclase (An_{23-39}), augite ($En_{30-34}Wo_{36-38}Fs_{29-35}$), pigeonite ($En_{34-37}Wo_{9-10}Fs_{54-56}$) and titanomagnetite, typically occurring as glomerophyric clusters. Inclusion relationships suggest that the sequence of crystallization of minerals should be pigeonite < augite < Ti-magnetite < plagioclase. These phenocrysts show embayed and rounded margins indicating resorption. The glass inclusions within plagioclase phenocrysts are compositionally akin to the matrix glass. Mineral compositions are inconsistent with the host rhyolites. Textural and mineralogical characteristics suggest that the phenocrysts were not in equilibrium with (i.e. crystallized before) the host rhyolite and that they must have crystallized from an iron-rich intermediate magma. This provides strong evidence for the coexistence of intermediate magma with rhyolite. The absence of intermediate products is believed to be related to the crystal load which lessen their probability of eruption.

Tristan volcano complex: oceanic end-point of a major African lineament.

Ken Bailey and Gill Foulger

No direct evidence for plumes is yet available: seismic tomography, currently the best hope, so far lacks the resolving power for plume detection (Grand et al., 1997). Hence, an assumed plume can be tested only by ground observations, or from volcanic materials. To evaluate, or to use, such circumstantial evidence it is necessary first to identify the key attributes inherent in the plume concept.

By definition, any postulated mantle plume must be independent of the overlying lithosphere, this independence (in terms of plate motions, for instance) being the foundation of the hotspot reference frame (Morgan, 1971). The plume should also be independent of lithosphere structure, and the timing of activity at other igneous complexes.

Lithosphere independence thus emerges as the only primary attribute of plumes. While it does not prove the existence of a plume, it is a necessary condition that should be checked. Where magmatism is demonstrably localised by some pre-existing structure in the lithosphere, for instance, there can be no direct connection to a plume (unless by extraordinary coincidence).

The first check should be for hot spot tracks, with age progression along their lengths, consistent with the plate moving over a fixed source. These have been sought in Africa since the early days of the plume concept, with varied results. Burke (1996) reviewed this question in detail, concluding that the only plausible candidate is the Tristan-Walvis Ridge (Figure 1) on the ocean floor sector of the African plate, pointing out that even here the Tristan hot spot has been stationary for the last 30 Ma. At the continental end is the Etendeka volcanic province (130 Ma), which he takes to be the initiation of the Tristan hot spot track. Alignment of the Walvis Ridge he attributes to northward movement of the plate (with its accreting ocean floor) over a fixed plume between 130 and 30 Ma (~ Anomaly 10), from which time sea floor spreading continued, but with the African plate stationary. By implication the hot spot was near the plate margin until Anomaly 10: a fixed plume model would then require a unique stationary point on the south Atlantic spreading ridge for 100 Ma (see Bailey, 1977).

If there is a volcanic chain along the Walvis Ridge, with a completely regular age progression, then it would be a track (130 Ma old), but its mode of formation can be assessed only against the full geological context, not in isolation. An immediate question arises with the Etendeka volcanism, located at the continental margin, and erupted at the time of complete lithosphere opening. Other major structures also converge around this place, where the continental margin is intersected by:

1. Major transform fracture zones of the SE Atlantic;
2. The southern boundary of the Angola craton and the Lufilian belt;
3. The northern boundary of the African Superswell.

Hence, any magmatism initiated at this site, at the time of plate separation, was clearly not independent of pre-existing lithosphere structure. In fact, this is such a remarkable focus of pre-existing and contemporaneous structure lines that a sub-lithosphere plume is not merely inapt, but superlatively so. When the discussion is widened the need for an alternative explanation becomes even more apparent.

Basalt-quartz latite volcanism at Etendeka is not akin to the strongly alkaline eruptions that characterise Tristan, and as if to underline this, activity at Etendeka did not cease at 130 Ma. A subsequent carbonatite complex was erupted through the 130 Ma lavas. From here, through Namibia/Angola/Congo Republic there is a belt of alkaline igneous/ carbonatite/kimberlite complexes stretching 1,900 km along the extension of the Walvis Ridge. These are of various ages from 795-70 Ma with no regular age progression (although many are contemporaneous with Etendeka). Some of the young complexes are at the distal end of the belt, and one at least may be Tertiary, which would make this belt almost a continental mirror image of the Walvis Ridge.

The great circle extension of the Walvis Ridge (Figure 1) is also close to the NW margin of the African Superswell. Although this boundary verges where it rounds the SE sector of the Congo Basin the topographic break reappears on the far side, being dramatically marked by the collinear NW margin of the Ethiopian Highlands, until it reaches the Red Sea (south of Mecca). In this sector, there are clusters of alkaline igneous/carbonatite complexes in the Sudan, along a belt parallel to the highland front. Hence this great circle follows a long lived lithosphere structure, marked in the geology, and moreover by lateral density variations in the deep mantle tomography, to which the Superswell has been related. Perhaps most remarkable in the topographic re-constructions of the Superswell is that the Walvis Ridge forms its NW margin where the Swell extends into the SE Atlantic. Thus the Walvis Ridge appears as an oceanic prolongation of this major step in the continental crust/lithosphere/mantle. It follows that the most straightforward explanation of the Walvis Ridge is that it formed by continuous propagation of a profound step in the African sub-structure as new oceanic lithosphere formed during spreading. Invocation of a fixed plume below the moving plate is therefore unnecessary.

This conclusion is consistent with the absence of tracks in the rest of Africa, where in stark contrast, magmatism is found repeated at the same sites in Cretaceous and Tertiary (Bailey, 1992), not at the ends of tracks. All the Cenozoic alkaline igneous/carbonatite/kimberlite activity across the whole African plate (including oceanic extensions) may thus be seen to be lithosphere dependent. Repetition of this low volume, volatile rich magmatism suggests that lithosphere structure not only localises the sites of eruption, but provides the required conditions for melting. Fluxing of volatiles from deep mantle through lithosphere penetrating lesions, must play a key role ("pie funnel effect": Bailey, 1983). The triggering events are regional expressions of global tectonic changes (Bailey and Woolley, 1999).

Other recognised linear zones in the African plate follow similar trends to the Tristan-Red Sea great circle (W, Figure 1), imparting a distinct NE-SW grain to the continent. One such NE-SW belt (marked by the mid-Zambesi Luangwa rifts) crosses Africa from Indian to Atlantic Oceans (Z, Figure 1), with rifts and magmatism along its length

(Bailey, 1961). This, and the Tristan zone, bracket the East African highlands, where the Superswell has been said to “jog NE” from its southern portion. Hence these ancient structural trends effectively frame the modern East African plateau, and its Tertiary volcanism.

The continental end of the Walvis Ridge, where major geological and topographic features converge, marks a crucial node in lithosphere structure, and the Ridge may be the result of its propagation across newly accreting ocean floor, in a manner similar to that suggested for the Hawaiian chain (Shaw and Jackson, 1973).

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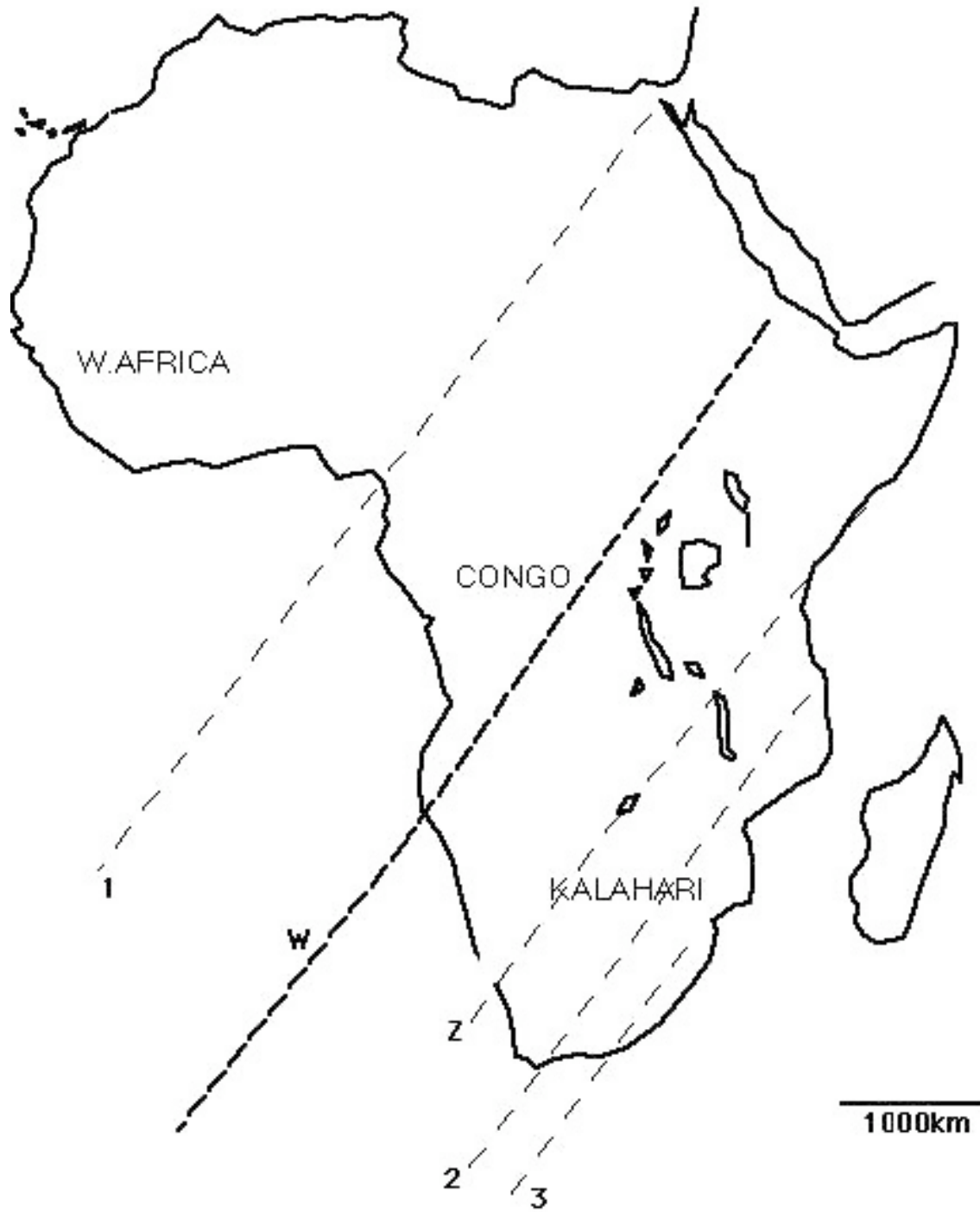


Figure 1. Linear zones across African plate: W, Walvis Ridge; Z, Zambesi-Luangwa; 1, St. Helena-Cameroon; 2, Cape Rise; 3, Shona Rise-Drakensburg.

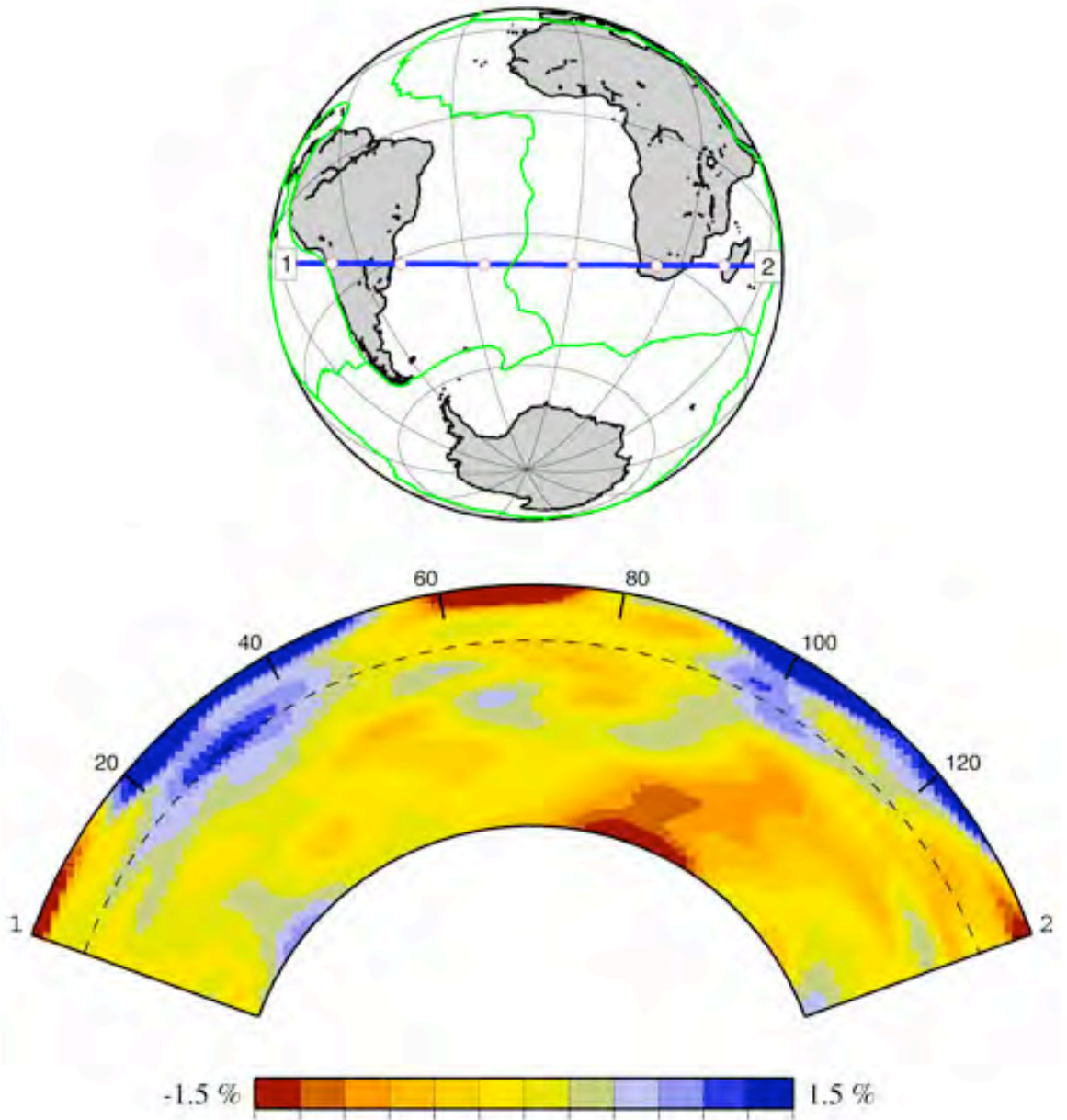


Figure 2. Whole-mantle tomography cross section through the Tristan volcano complex, using the tomographic model of Risema et al. (1999).

Extended abstract for Penrose conference. Plume IV: beyond the hypothesis
(poster presentation)

**Cenozoic intraplate volcanism in Mongolia; if not a mantle plume then
what?**

T. L. Barry

Diffuse, small-volume basaltic volcanism has occurred intermittently throughout central Mongolia for the past 30 My. Trying to determine causes of long-lived intraplate volcanism is difficult, particularly where there is uncertainty over the timing of related extension, uplift, and of course magmatism. The problem is most acute in central and east Asia, one of the largest and least known areas of intraplate igneous activity. In this region, Cenozoic basalt fields are scattered across an area approximately 2000 km from east to west, and 1500 km from north to south (Whitford-Stark, 1987). This region provides an excellent opportunity to study intraplate continental volcanism that is far removed from the effects of subduction-related processes or continental rifting. The crust within central Mongolia is estimated to be 45 km thick on the basis of P-T studies of crustal xenoliths (Stosch *et al.*, 1995). Although magma has erupted through 45 km thick crust, the basalt rocks appear to show very little crustal contamination (Barry *et al.*, 2003). The volcanism provides an important link between Cenozoic basaltic volcanism to the north around the Baikal rift zone and similar aged basaltic volcanism to the east in NE China which infills extensional grabens. Volcanism and rifting in the Baikal region has been explained by a range of processes including (1) partial melting of small asthenospheric diapirs (Ionov *et al.*, 1998) or a mantle plume (e.g., Logatchev, 1984; Windley and Allen, 1993), (2) a crustal weakness (Yarmolyuk *et al.*, 1991) or (3) the combined effects of the India-Asia collision with secondary input from a mantle plume (Khain, 1990). Whereas lavas erupted in NE China have been attributed to passive upwelling of ocean island basalt (OIB)- and/or mid-ocean ridge basalt (MORB)-type mantle, with subsequent modification of magmas by assimilation of continental lithospheric mantle or crust (e.g., Song *et al.*, 1990; Tatsumoto *et al.*, 1992). A close spatial relationship between older Paleogene Chinese basalts and sedimentary basins

suggests that, initially at least, Cenozoic magmatism was associated with extension and thinning of the lithosphere along the eastern margin of the Eurasian plate (Northrup *et al.*, 1995), which in turn related to the rate of convergence between the Pacific plate and Eurasia, with the Indo-Eurasian collision only a 'far-field' influence (cf. Northrup *et al.*, 1995). Younger Neogene basalts, however are not confined to basins. Despite contrasting explanations, the chemical composition of lavas in Baikal, Mongolia, and NE China are remarkably similar, especially given differences in underlying crustal and lithospheric mantle compositions and ages of eruption (Barry & Kent, 1998). This is inferred to suggest that the mantle source region beneath this vast area has played a significant role in contributing to magma genesis. Therefore, when considering a model to explain volcanism within Mongolia, which shows neither rifting-dominated processes (e.g. like Baikal) nor extension-related processes such as the NE China basins, one general model should be able to explain all the volcanism.

Trace element, REE and isotopic modeling of Mongolian basalt compositions indicates that basaltic melts were most likely sourced within the lowermost lithospheric mantle. Large-ion lithophile elements and Nb concentrations suggest recently metasomatised lithosphere. This is further supported by studies on mantle xenoliths erupted in the Mongolian and Russian basalts that show multi-enrichment and replacement textures (Ionov *et al.*, 1994). There is no evidence for high heat flow within the mantle beneath Mongolia (Khutorskoy & Yarmolyuk, 1989), but recent geophysical studies infer anomalously dense material to be present at the base of the lithospheric mantle (Petit *et al.*, 2002) which is coincident with a low velocity zone at ~200 km depth (Villaseñor *et al.*, 2001). However, there does not appear to be anomalous low velocity material within the asthenospheric mantle or extending deeper towards a mantle boundary. The geochemistry of the basalts gives no positive indication for the presence of an underlying mantle plume and can be explained by progressive enrichment of the lithospheric mantle by small degree partial melts from the asthenosphere; even helium isotopes plot within the range of values found for sub-continental upper mantle and are not elevated. Conversely, whilst localized extensional tectonics may have facilitated transport of magma to the surface, the small amount of extension cannot account for the generation of the basalts (McKenzie & Bickle, 1988) thus requiring a mantle process to

first generate melt. The area lacks evidence for a high heat flux mantle plume e.g. any age progression of volcanic eruptions, any excess lithospheric temperatures within mantle xenoliths (Ionov *et al.*, 1998), and lack of geophysical evidence for a mantle upwelling. Therefore, rather than a mantle plume it is suggested that Mongolia and adjacent regions are underlain by a thermal anomaly, i.e. elevated thermal energy within the asthenosphere but not a chemically distinct, thermally buoyant anomaly sourced from a boundary layer within the mantle. The ultimate cause of such a phenomenon remains enigmatic but, if understood, may help explain other regions of diffuse, long-lived intraplate continental volcanism.

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Lithospheric Stress State Responsible for Hotspots at Ridge-Transform-Intersections?

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Contrary to our current understanding of ridge-hotspot interactions, several papers have noted small seamounts with hotspot geochemical signatures located at or near ridge-transform-intersections (RTIs) rather than at the center of ridge segments (Graham et al, 1996; Hekinian et al 1999; Johnson et al, 2000; Klingelhoger et al, 2001). I propose that these ridge-transform-intersection seamounts are the result of lithospheric stress concentrations that propagate into the mantle. This implies some seamount locations are influenced by lithospheric stress concentrations, which suggests a new mechanism by which all “hotspot” locations may be affected.

The existence of seamounts with hotspot geochemical signatures, (hereafter simply called hotspots with no implied inference to their origin), in the center of ridges is relatively well understood. In the case of a plume origin for hotspots, ridges are believed to focus plumes: a) because hotspots are entrained in the mantle upwelling that feeds the ridges, b) because they are weak areas within the crust, and c) because there exists a physical and thermal gradient that effectively funnels the material to the ridge like a sink to a drain (Figure 1) (Kincaid et al, 1996; Kincaid et al. 1995; Sleep, 1990). Once at the ridge, hotspots add magma to the central area of upwelling, i.e. the center of the ridge segment. In the case of a heterogeneous mantle origin for hotspots, ridges are believed to intersect areas of fertile mantle, which results in rapid decompression melting and seamount formation (Anderson, 1992). Neither theory offers a clear explanation for the presence of hotspots at ridge-transform-intersections.

Any hypothesis presented regarding the presence of hotspots at ridge-transform-intersections must not only explain why hotspots exist at RTIs, but also why they would only exist there on occasion. I propose that seamounts form at RTIs because stress is concentrated at crack-tips, but that this stress is only large enough to influence the mantle and produce seamounts under certain conditions. Hotspots at ridge-transform-intersections may occur when ridge-tip propagation is not feasible or during the build-up of stress prior to propagation.

Three-dimensional finite element models of a series of ridge segments were constructed to determine if the stress field at the ridge-transform intersections changes when slip along the transforms is inhibited. The elastic finite element program Felt by Gobat and Atkinson (1996) was used to construct a box model of 3 ridge segments 1200-1400 km long separated by transforms 600-800 km long. The ridges are modeled as weak areas in a relatively strong 100 km thick lithosphere that is underlain by a 100 km weak low-velocity zone and then 400 km of mantle. Ridge-push forces were applied to the ridges and slab-pull forces were applied to the edges of the model. A variety of scenarios involving different transform strengths (i.e. ease of slip) and force ratios were run to determine what effect changing transform strength had on the stress field at the ridge-transform intersection. In all cases, increasing the resistance to slip along the

transform resulted in increased extensional stress at the RTI not only in the lithosphere, but also in the mantle (Figure 2).

Increased extensional stress at the ridge-transform intersection can result in ridge propagation, however, multiple criteria may need to be met for the ridge to propagate. Phipps Morgan and Parmentier (1985) proposed that ridge propagation could only occur when the stress at the ridge tip is: a) greater than the strength of the plate; b) greater than the energy to create a new transform (also Lachenbruch and Thompson, 1972); and c) greater than the energy dissipation from the opposing ridge (also Pollard and Aydin, 1984). Other factors affecting ridge propagation include the degree of asthenospheric flow to the ridge tip and the topographic gradient between the ridge center and ridge tip (Phipps Morgan and Parmentier, 1985; Parmentier and Forsyth, 1985; Spence and Turcotte, 1985; Wilson and Hey, 1995; West et al., 1999). If the stress at the ridge tip (RTI) is not great enough to overcome most of the above criteria, then we must assume that the ridge will not propagate and high levels of extensional stress will be concentrated at the ridge tip (RTI).

It is under this condition that I propose hotspots form at the ridge-transform-intersection (RTI). High levels of extensional stress in the lithosphere and asthenosphere (Figure 3) will likely result in localized normal faulting events and crustal thinning. Crustal thinning and extensional stress in the asthenosphere would result in decompression of the mantle.

Decompression of the mantle can produce a seamount with a hotspot geochemical signature via two very different means. In the case of a heterogeneous mantle, decompression melting of fertile mantle would result in an excess of melt with a hotspot signature. Decompression of the mantle and the formation of a melt would also a change the thermal and physical gradient at the ridge. Crust at the transforms tends to be thinner than at the center of the ridge and the presence of melt would increase the positive buoyancy of the crust. This change in the physical and possibly thermal gradient could draw plumes from the center of the ridge to the ridge-transform-intersection (Figure 3).

To conclude, the presence of seamounts with hotspot geochemical signatures at ridge-transform-intersections are potentially the result of concentrations of extensional stress at the RTI due to inhibited slip along the adjacent transform. These hotspots may be the result of either plumes or a heterogeneous mantle as both would be affected by the attendant lithospheric decompression.

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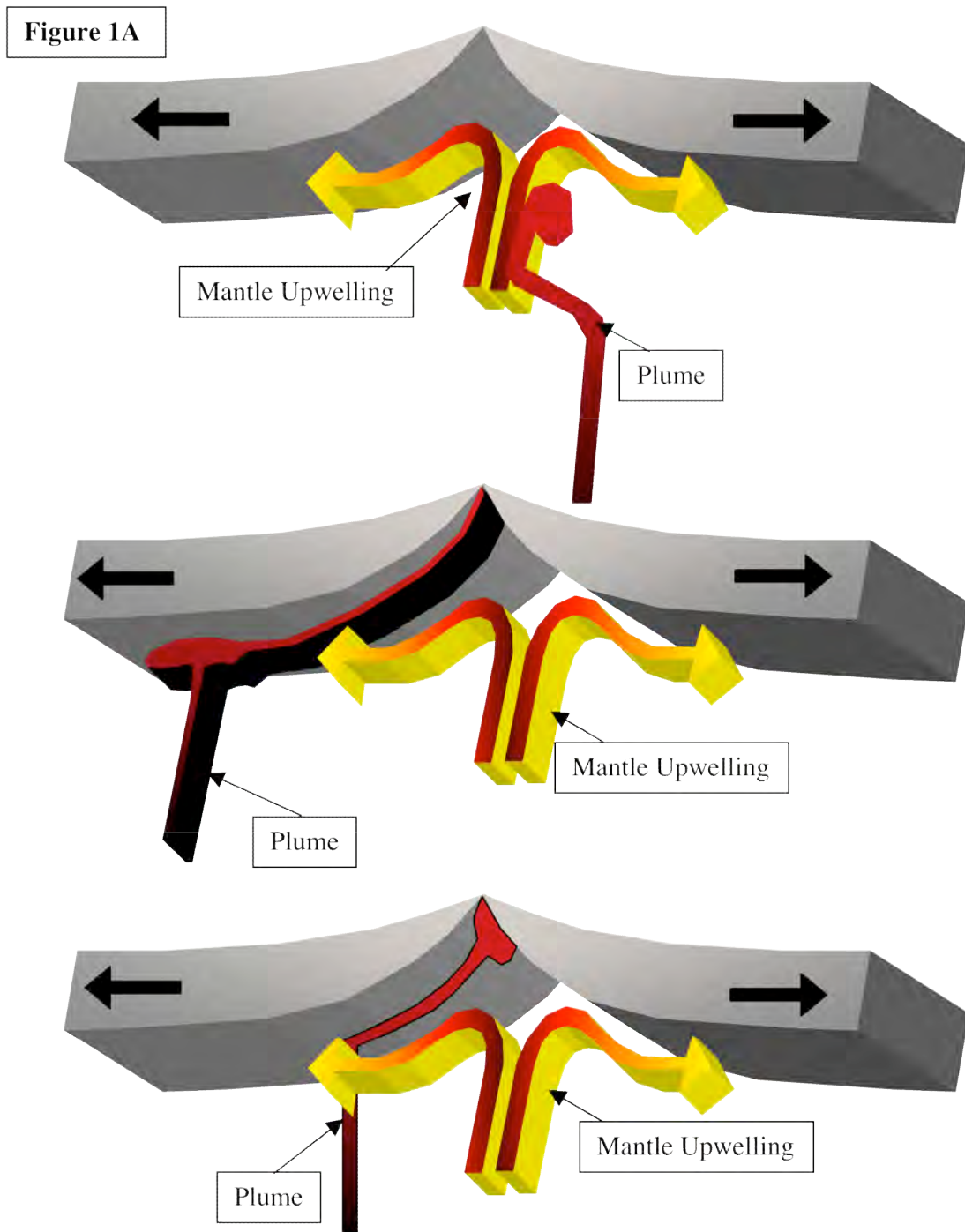


Figure 1: Possible Ridge-Hotspot Interactions (e.g. Kincaid et al., 1996; Kincaid et al., 1995; Sleep, 1990)

A: Plume is entrained in shallow upwelling beneath ridge.

B: Cartoon of a hotspot supplying material to a ridge via a conduit flowing up the topography of the ridge and overcoming the mantle flow away from the ridge.

C: Cartoon of a hotspot entrained to a ridge by flowing along the base of the lithosphere, also overcoming the mantle flow away from the ridge.

Figure 2A: Mapview Results, 33 Km, Strong Transform

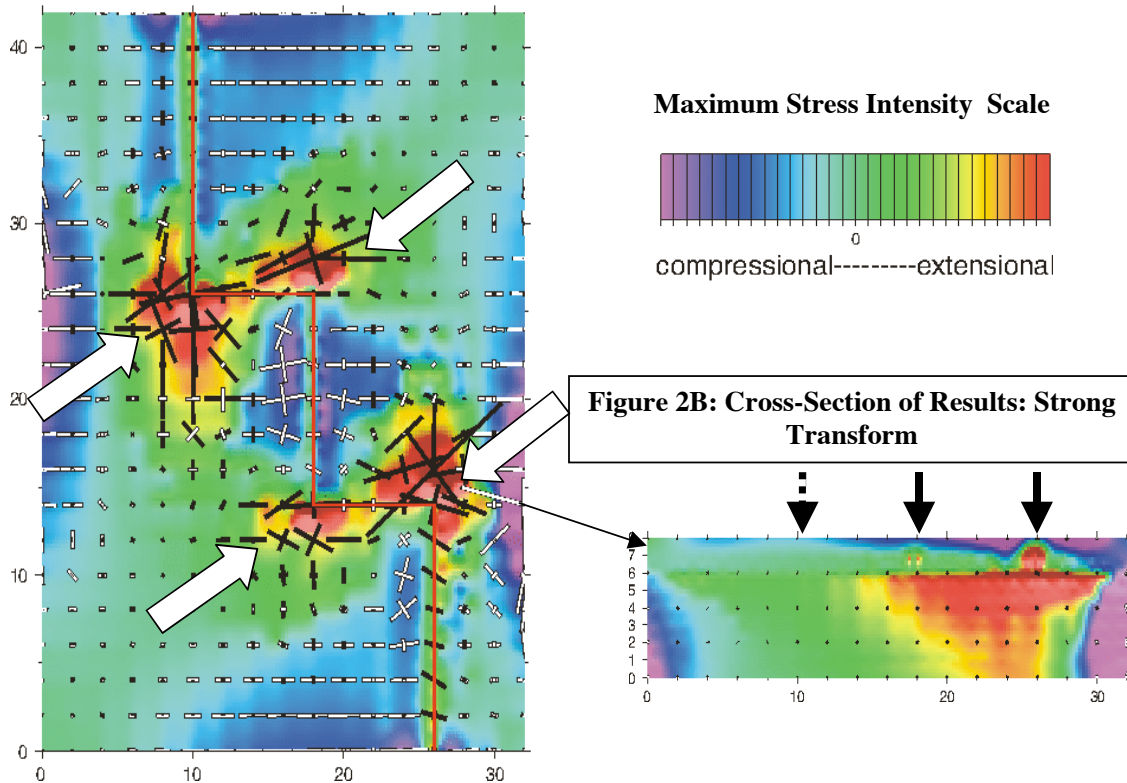


Figure 2B: Cross-Section of Results: Strong Transform

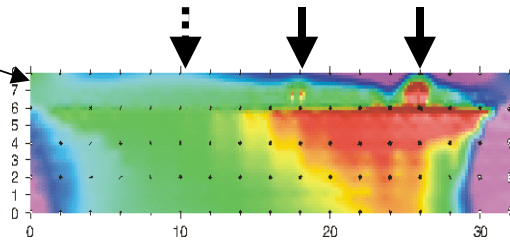


Figure 2A: Mapview Results, 33 km Depth, Strong Transform: Results of ridge-push forces exerted on a 3-D box with 3 ridges and 2 strong transforms. The model is fixed at the base of the box, 8 units (~800 km) below the surface. The background colors indicate the type and intensity of the maximum stress, blue being compressional and red being extensional. The bars are maximum and minimum stress vectors, white indicates compression and black indicates extension. White arrows indicate zones of extension at ridge-transform-intersections.

Figure 2B: Cross-Section of Results, Strong Transform: Cross-section of model described above. As indicated by the arrow, the cross-section location is parallel to the first transform (at Y=14) on the mapview results. Color scheme and bars are as described above. Solid black arrows indicate location of ridge-transform intersections, and dotted arrows indicate the where the ridge segment would exist if it were extended in length. Note that the extensional stress propagates through the lithosphere, asthenosphere and into the mantle.

Figure 3A: Along Ridge Cross-Section

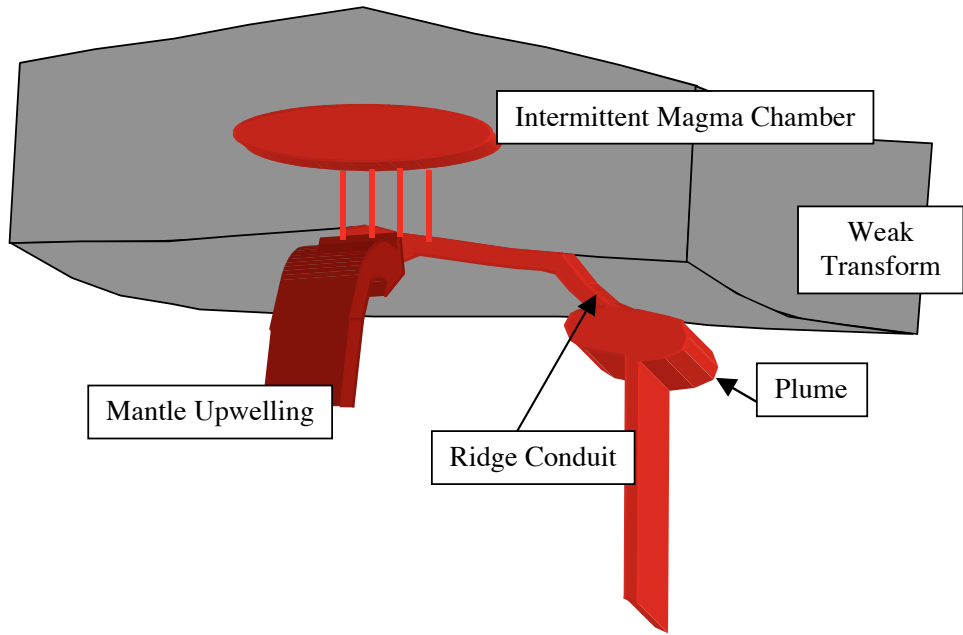


Figure 3B: Along Ridge Cross-Section

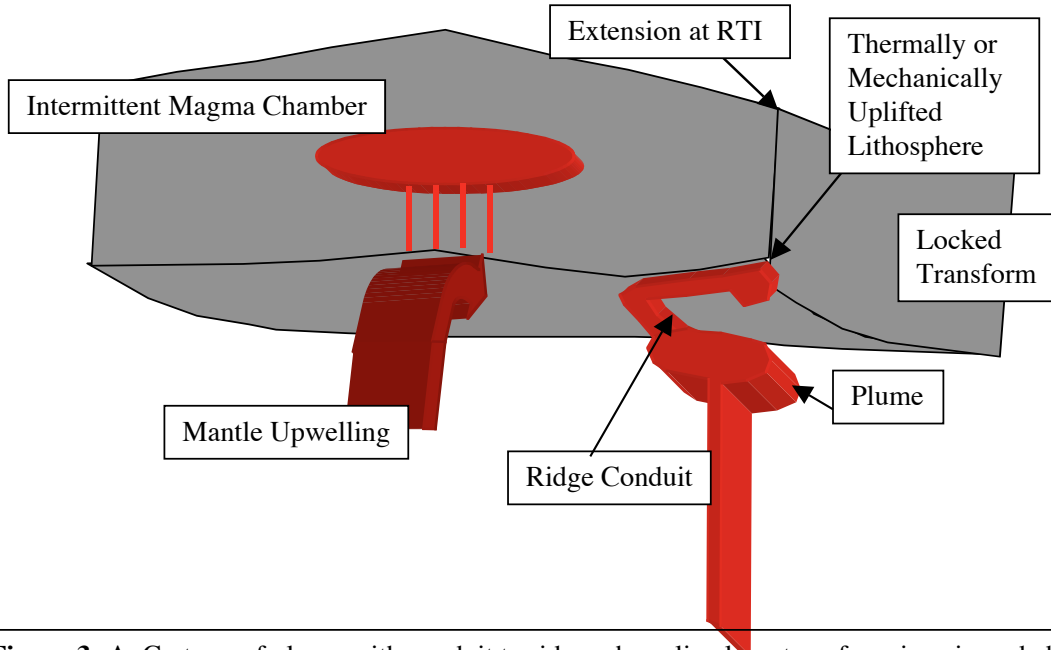


Figure 3: **A** Cartoon of plume with conduit to ridge when slip along transform is unimpeded. **B:** Cartoon of plume conduit to ridge diverted by extension at the ridge-transform-intersection. Note that both cartoons are along ridge cross-sections and show a schematic of the ridge from one transform to the next.

Is the Mid-Atlantic Ridge becoming hotter with time?

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More than 20 million years of oceanic lithosphere accretion history at a segment of the Mid-Atlantic Ridge are recorded in the Vema Lithospheric Section (VLS), a 300 km long flexured and uplifted sliver of lithosphere exposed near the Vema Fracture Zone in the Central Atlantic. Systematic sampling of the basal mantle ultramafic unit and of crustal basalts along the VLS together with geophysical surveys gave us the opportunity to study temporal changes in the processes of generation of the oceanic lithosphere at a ridge axis. The degree of melting of the mantle upwelling below the ridge axis, estimated from the chemistry of mantle-equilibrated mineral phases in the peridotites, as well as crustal thickness, inferred from shipboard and satellite gravity data, both show ~3-4 my long oscillations superimposed on long-range steady increases with time. Based on basaltic glasses elemental and isotopic chemistry, we assume the composition of the source stayed nearly constant. The steady increase with time of mantle degree of melting and of crustal thickness suggests that the mantle rising beneath the MAR became gradually hotter during the last 20 million years, even though the spreading half rate slowed significantly during this time. We offer two explanations for the increase in mantle temperature with time at the Mid-Atlantic Ridge. A first possibility, of local significance, calls for gradual lengthening of the eastern MAR segment where the VLS was created, leading to an increasing degree of melting below center of the segment as it lengthens, due to the a decreasing influence of the "cold edge effect" from the Vema transform. The second, of broader significance, calls for a gradual increase of mantle potential temperature along a significant portion of the northern MAR during the last 20 million years, resulting in an increase of melt production despite decreasing spreading rates. This second hypothesis is supported by an increase of crustal thickness towards ridge axis observed at several other locations in the northern MAR. The

chemistry of basaltic glasses, collected along the VLS above the peridotites, suggests that no deep plume source is involved in the steady heating of the Ridge.

Volcanism synchronous with mantle exhumation at the axial zone of a fossil slow spreading ocean: evidences from the Chenaillet ophiolite (Franco-Italian Alps)

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The mechanisms checking together emplacement of volcanoes, mantle and gabbro exhumation and ocean enlargement at the axial zone of a slow spreading ocean are poorly understood. In order to better assess how they could be linked, a detailed mapping of a fossil ocean-floor structure, preserved from alpine tectonic and metamorphism, was performed in the Chenaillet unit (Franco-Italian Alps)(Chalot-Prat & Coco, submit.). In parallel a geochemical sampling of basalts was carried out to constrain the observed space and time relationships between volcanoes and to decipher the periodicity of replenishment of reservoirs. The results are as follows:

From its overall dimensions (segment of 3-5 km in width on 30 km²), the morphology and topography of the volcanic complex (numerous narrow ridges, 200 m to 1 km in length on 100 to 500 m in width, 100 to 600 m above the oceanic floor with steep slopes), the global architecture of the volcanic complex (several strings of micro-segments forming composite volcanoes at their junction; pseudo-symmetry of volcanic strings), the dimensions of composite volcanoes (700-1500 in diameter on 100-300 m high, some hundreds of volcanoes), the dimensions and internal structure of single volcanoes (3-15 m in diameter on 3-30 m high), the Chenaillet unit would be a witness of an axial zone of Atlantic type, similar to a volcanic abyssal hill or a 4th order segment in the Reykjanes Ridge. On both sides of this volcanic hill, large zones of exhumed basement strewn with rare volcanoes attest for the discontinuity of eruptive zones as in any internal rifts of slow spreading oceans.

The basaltic cover of the volcanic hill is only up to 50 m thick and overlays a basement of serpentinized peridotites and gabbros. The top of this basement is convex upward below the composite volcanoes. It is underlined by a tectonic breccia horizon (10 cm to 5 m thick) attesting for the existence of a detachment fault zone responsible for the basement exhumation at the seafloor (Chalot-Prat & Manatschal, 2002). Clasts of dolerite, found within the fault zone, indicate that the basement exhumation had to be active during and even after emplacement of volcanoes.

On the volcanic hill, stair- and comb-type volcanic systems check the distribution of separate volcanoes. Both types of systems are built on rather steep slopes, and everywhere the higher the volcano, the younger it is relative to the others.

In the stair-type (up to 600 m of height difference between base and top on up to 2 km in length), each step is formed with a stacking of several pillow and tube tongues fed from fissural conduits located at the root of each step (Fig. 1). This system would form by uplift, step by step fracturation of an already exhumed basement, and magma injection along the fissures once formed.

The comb-type, the most widespread system on the volcanic hill, consists in a number of alignments of conic volcanoes on a steep slope (50 to 200 m of height difference between base and top on 100 m to 1 km in length) (Fig. 2). For each volcano (3-15 m in diameter on 3-30 m high), the tube and pillow distribution implies that it was built on a slope. The central feeder dyke, vertical or most often steeply inclined up- or down-hill, is located at the crossing

of two directions, oblique (*tooth*) and parallel (*line*) to the *branch* of the comb. Towards its base, each feeder-dyke appears frequently curved up- or down-hill, and seems uprooted above the detachment fault zone. On each tooth, the higher the volcano, the younger it is relative to the others, while along a same line, eruptions are coeval as proved by rhythmic variations of chemical compositions of basalts from one line to another. This also means that the branch of the comb superposed to a deep fracture which served as main magma conduit for the comb system.

All these observations suggest that these “in line” volcanoes formed on the ridge of a basement in uplift, at the crossing between a steeply dipping major fracture (main conduit) and a set of shallow parallel fractures (secondary conduits). Once formed, volcanoes were dragged away and down on a travelator to give place to new volcanoes and so on. This travelator consisted in a new basement (mantle or gabbro), the top of which was underlined by a detachment fault. This detachment fault formed at depth and was nothing else than the main fissural conduit which canalized the magma up to the surface. Thus the building of comb systems was synchronous with an enlargement of the basement surface. Besides, and it is an important point, most comb structures are pseudo-symmetric on both sides of narrow ridges (same dipping of feeder-dykes, but teeth with different directions and length). This evidences that the exhumation process occurred in opposite directions, synchronously but not at the same rate as observed at any mid-oceanic ridge axis.

From the major and trace element data on basalts within a same system (comb or stair), the basalts appear to be related by both partial melting and fractional crystallization trends. The fast alternation between partial melts and their differentiates suggests that the main magma conduit was rooted within a frequently fed small reservoir. Nevertheless the regular recurrence of partial melts within a same comb could also suggest that the detachment fault rooted within the asthenospheric mantle source.

Thus at the axial zone of a Mid-Oceanic Ridge, volcanic eruptions would trace the emergence of detachment faults, themselves closely associated with lithospheric matter stretching and transfer up to the sea floor.

Figure 1: sketch of a stair-type volcanic system (in Chalot-Prat et al, submit.)

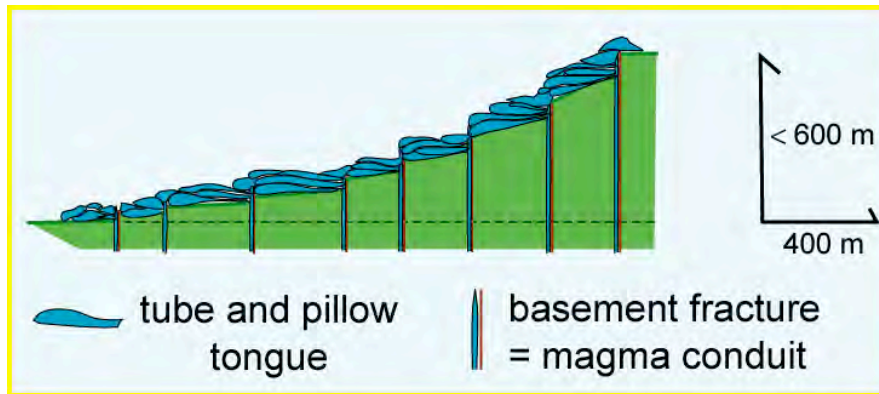
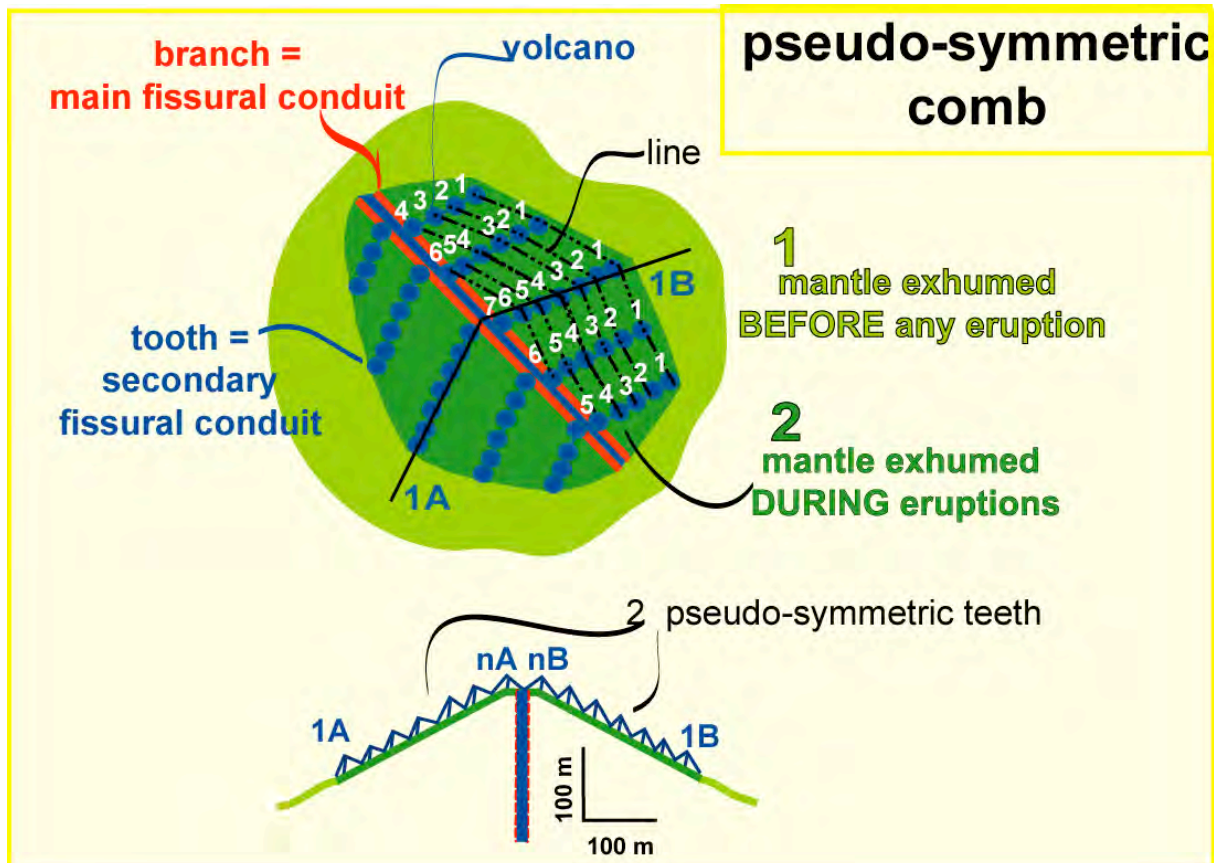


Figure 2: sketch of a pseudo-symmetric comb-type volcanic system (in Chalot-Prat et al, submit.)



Oligocene calderas, mafic lavas and radiating mafic dikes of the Socorro-Magdalena magmatic system, Rio Grande rift, New Mexico: surface expression of a miniplume?

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Common traits of mantle plumes are: 1) domal uplift prior to volcanism, 2) a definite age progression along a volcanic chain, 3) long mafic dikes that radiate from the volcanic core, 4) large basaltic plateaus or shield volcanoes, and 5) petrochemical indicators of anomalously high temperature melt zones in the upper mantle, such as high Ni/MgO ratios in picritic basalts (Campbell, 2001). We suggest the Socorro-Magdalena magmatic system of Oligocene age (Fig.1, eruptive volume 7100 km³) exhibits characteristics similar to a mantle plume, but at 1/10-1/100th the scale of a deep mantle plume, which may qualify it as a miniplume?

A cluster of five overlapping ignimbrite calderas is moderately well exposed in strongly extended, tilted fault-block mountain ranges of the central Rio Grande rift southwest of Socorro NM (Fig.1). The westward younging Socorro-Magdalena caldera cluster (SMCC) is 85 km long and 20-25 km wide. It parallels the southeastern margin of the Colorado Plateau and the WSW-trending San Agustin arm of the rift. The latter produces the appearance of an incipient triple junction within the dominantly north-trending rift system. Precise ⁴⁰Ar/³⁹Ar ages of sanidines from the rhyolite ignimbrites and detailed geologic mapping demonstrate that the distended calderas become progressively younger to the west-southwest (McIntosh et al., 1991; Chamberlin et al., in press). Large volume ignimbrite eruptions occurred at 31.9, 28.7, 27.9, 27.4 and 24.3 Ma. A large satellitic caldera, which formed at 28.4 Ma, is located 20 km southwest of the main overlapping trend. A small collapse structure, which is nested in the Socorro caldera, erupted at 30.0 Ma. The total volume of densely welded ignimbrite and moat-fill lavas erupted from the SMCC is 5500 km³.

Within 40 km of the northeastern margin of the caldera cluster, the rhyolite ignimbrites are interleaved with a 400-700 m thick plateau-like belt of basaltic andesite lavas (Fig.1). These mafic lavas are assigned to the La Jara Peak Basaltic Andesite (Osburn and Chapin, 1983a). They range from slightly alkaline trachybasalt to moderately alkaline basaltic trachyandesite and sub-alkaline basaltic andesite. Sparse small phenocrysts of olivine, commonly altered to reddish brown iddingsite, are characteristic. Phenocrystic plagioclase, indicative of differentiation at depths less than 30 km (Wilson, 1989), is typically absent. Individual basaltic andesite flows are commonly 7-10 m thick. Stacked flows between ignimbrites have an aggregate thickness of as much as 330m and locally define wedge-shaped prisms formed by domino-style extension in the early Rio Grande rift (Chamberlin, 1983; Ferguson, 1991). A 32-33 Ma flow and tephra unit near La Joya was locally fed by a short NE striking basaltic-andesite dike that appears to radiate from the 31.9-Ma Socorro caldera (Fig1). A primitive trachybasalt in the SE moat of the Socorro caldera (~31 Ma; Chamberlin et al., in press) contains 9.3 % MgO and 170 ppm Ni; this suggests a relatively hot source zone in the mantle, compared to most subduction related basalts (Campbell, 2001). The total volume of Oligocene basaltic andesite lavas peripheral to the SMCC is 1600 km³. The maximum rate of basaltic andesite eruption, ~1800 km³/Ma, was coeval with the zenith of domino-style extension and apparent caldera migration at 27.9-27.4 Ma.

Moderately alkaline to sub-alkaline basaltic andesite and trachybasalt dikes of Oligocene age (31-24 Ma, K/Ar, Aldrich et al., 1986; Laughlin et al., 1983) form a large semi-continuous radial array that is broadly focused on the SMCC (Fig.1). The Magdalena radial dike swarm (MRDS) fans through

220° of arc from Pie Town clockwise to Acoma, Chupadera, Bingham and Elephant Butte. The maximum radius of the MRDS is 125 km; subswarms of near parallel dikes are commonly 20-75 km long. The mafic dikes are commonly fine grained or diabasic, medium to dark gray, and often contain microphenocrysts of olivine, which may be altered to iddingsite. Phenocrystic plagioclase is usually absent. Longer dikes are typically basaltic andesites (53 -56 wt. % SiO₂). Within the NW-striking Pie Town subswarm, exposed dike length (75, 24 and 2 km) decreases with increasing MgO content (3.1, 5.9 and 8.0 wt. %, respectively; Baldrige et al., 1989). This is consistent with dominantly horizontal flow and lateral intrusion of magma at different levels of neutral buoyancy (Fig.2). Shorter NNE- and NNW-striking dikes near Riley are more numerous and closely spaced than in other sectors of the MRDS. Some of the shorter dikes were probably coin-shaped feeder dikes similar to the 32-Ma dike at La Joya. Cross cutting relationships of intersecting dikes near Riley and La Joya are generally consistent a westerly migrating focus of mafic magmatism. However, one NNW striking dike south of Riley exhibits bifurcations to the NNE, is cut by a NNE striking dike and terminates in a NNE bend. These dike patterns and similar orientations of early rift faults suggest a complex interplay of short-term magmatic stresses and longer-term plate stresses. New ⁴⁰Ar/³⁹Ar data for dikes near Riley (Fig.1) demonstrate that they were emplaced during the peak of regional extension, caldera migration, and mafic volcanism. Several NNE-striking dikes near Riley also contain granitic xenoliths that exhibit magmatic reaction rims. Collectively, the above observations imply that radial dikes of the Magdalena swarm are most likely linked at depth with a westerly migrating mafic sill complex in the lower crust under the caldera cluster (Fig.2). Presumably the inferred mafic sill complex also fueled rhyolite generation in the caldera cluster. Projections of mafic dikes shown on Figure 1 commonly intersect about 5-10 km north of the caldera cluster; this suggests that the inferred lower crustal sills are larger in diameter than the overlying calderas.

A sequentially rising diapir chain, associated with rollback of the Farallon slab through the upper mantle, is suggested as a reasonable mechanism to produce the observed westerly migrating calderas and surrounding mafic magmatism (Fig.3). Slab rollback may have created a westward migrating diapir source zone (Rayleigh-Taylor instability; Olson, 1990) that fed a sequentially rising chain of diapirs, thus producing the apparent westward migration of magmatism in the overlying crust. Alternatively, westward flow in the upper mantle wedge may have tilted a rising diapir and sheared it into multiple segments (Whitehead and Helfrich, 1990). Regional east-west extension in the early rift at this time also implies westward stretching at a rate greater than westward drift of the North American plate. Indications of regional doming prior to Oligocene mafic volcanism at Socorro are presently unclear; if present, the expression must be subtle. Cycles of pre-caldera tumescence and post-collapse resurgence are, however, locally evident within the eastern section of the caldera cluster (Chamberlin et al., in press). The SMCC is underlain by a broad Laramide uplift of early Tertiary age, which is generally attributed to regional crustal shortening.

The tectonic setting of the Socorro-Magdalena magmatic system is somewhat unique. The Socorro system lies at the north end of a 1500-km long magmatic arc of Eocene to Oligocene age. The arc locally terminates against relatively rigid microplates of the Colorado Plateau and the Great Plains near Socorro. The Socorro region also appears to occupy an incipient triple junction (now abandoned?) within the Rio Grande rift system. If our model is correct, then the contrast between basaltic plateau volcanism to the north, small shield volcanism to the west (Fig.1), and satellitic rhyolite volcanism to the south can be largely attributed to lateral variations in the thermal regime of the crust at that time. Regional map relationships also suggest that mafic dikes propagating through the lower crust may interconnect with deep plumbing systems of separate magmatic centers as much as 100 km distant.

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OLIGOCENE CALDERAS, MAFIC LAVAS AND RADIATING MAFIC DIKES OF THE SOCORRO-MAGDALENA MAGMATIC SYSTEM

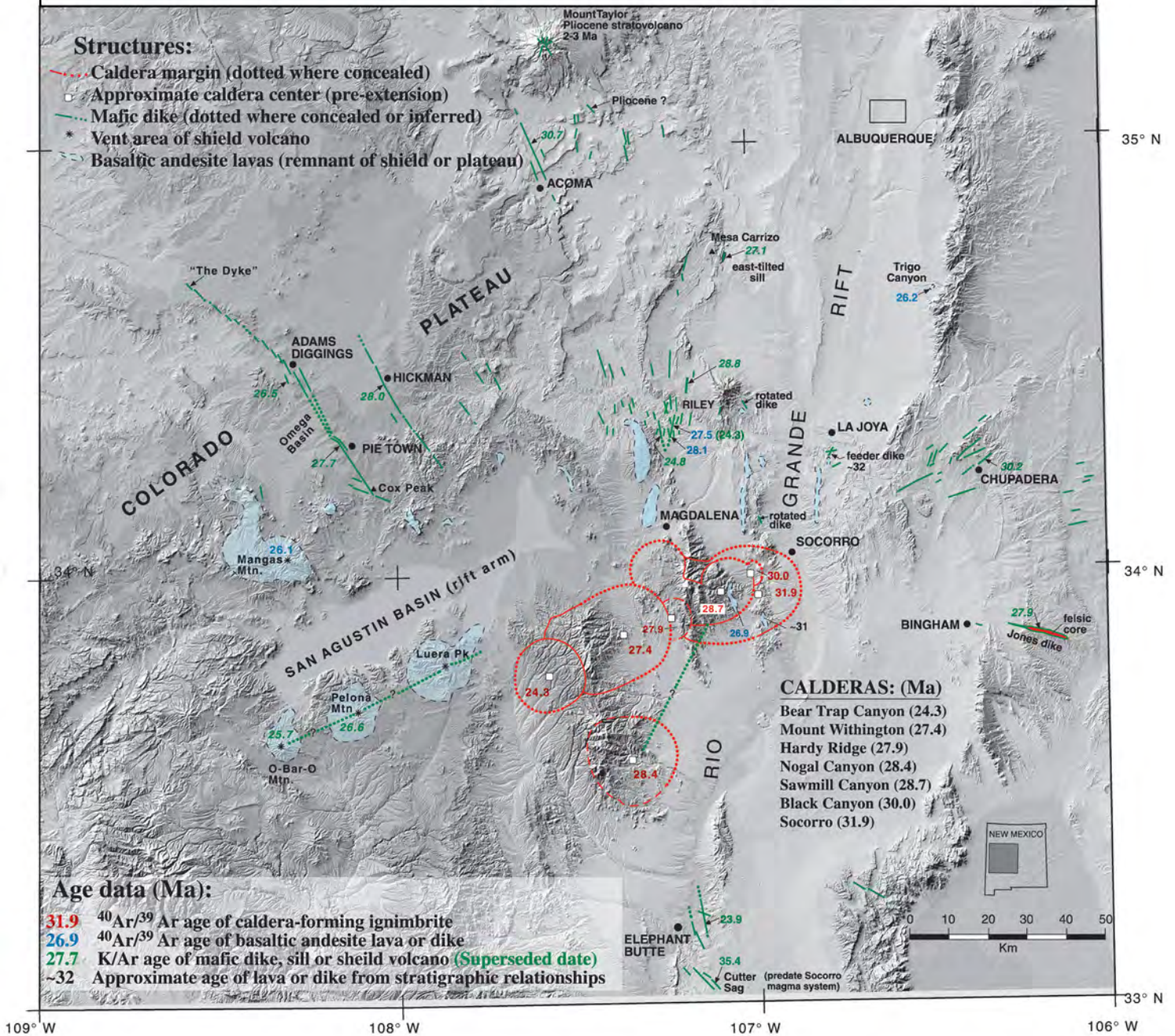


Figure 1. Oligocene calderas, mafic lavas and radiating mafic dikes of the Socorro-Magdalena magmatic system (after Chamberlin et.al., 2002). Data sources are listed at http://geoinfo.nmt.edu/staff/chamberlin/mrds/Chamberlin_2002_GSA_poster.pdf

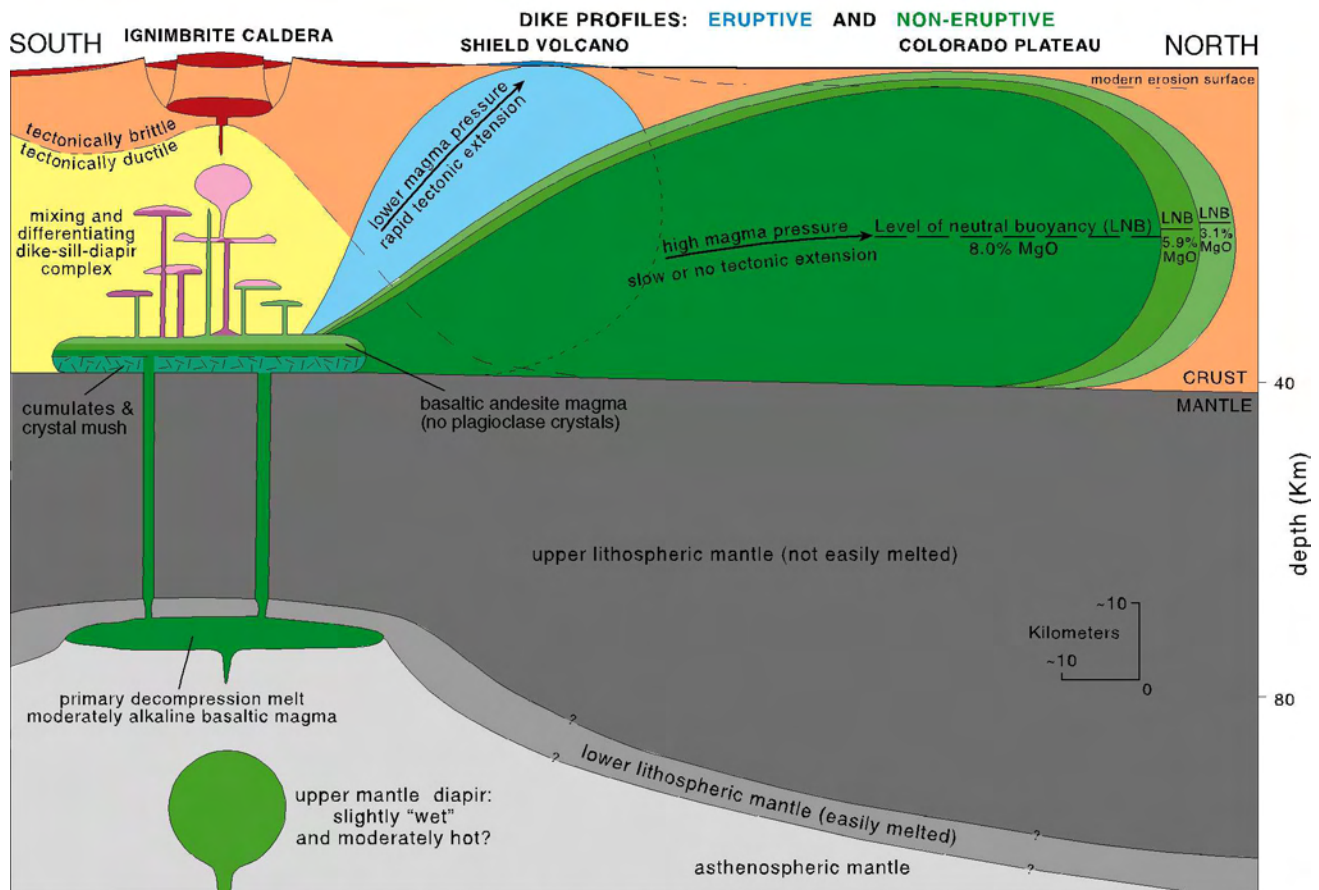
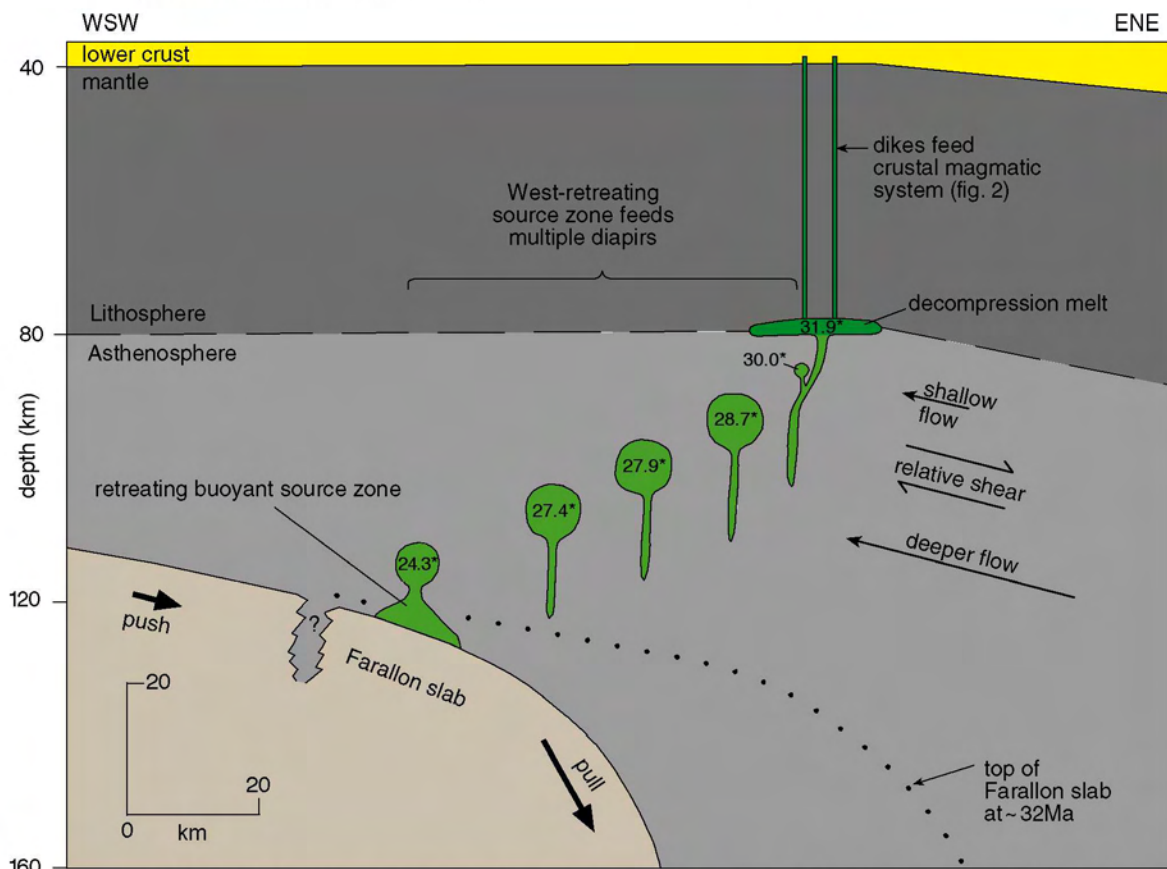


Figure 2. North-south cross-sectional model of the Socorro-Magdalena magmatic system at ~28 Ma (after Chamberlin et al., 2002).



*Numbers indicate the presumed link between an ignimbrite caldera (age in Ma) and a particular diapir

Figure 3. East-west cross sectional model of an upper mantle diapir chain associated with rollback of the farallon slab in Oligocene time (after Chamberlin et al., 2002).

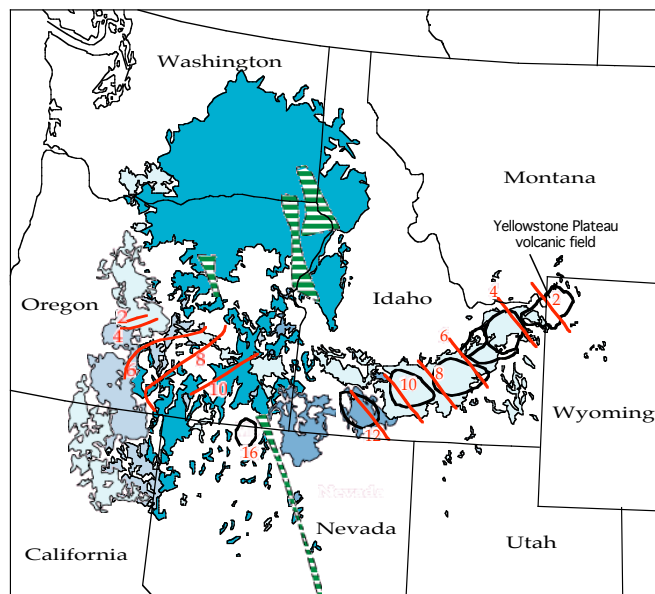
STRUCTURAL CONTROL AND PLATE-TECTONIC ORIGIN OF THE YELLOWSTONE MELTING ANOMALY

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Conventional models of lithospheric magmatism, its surficial volcanic expressions, and its tectonic contexts are commonly couched in dichotomies. The bulk of Earth's magmatism is seen as systematically related to plate boundaries, with the rest characterized somewhat vaguely as "intraplate". Certain magmatic compositions and rates of magma production at plate margins are taken as normal, in contrast to anomalous compositions and anomalously high production rates at "hotspots", typically expressed by linear chains of age-progressive volcanism. Deep-earth processes are categorized as active or passive, commonly inferring plate-margin magmatism to be a more or less passive response to plate-boundary tectonic processes and "hotspot" magmatism as representing chemical and thermal anomalies that originate deep in the mantle, perhaps even driving or modulating plate motions. The sources of diverse isotopic and trace-element compositions of primitive magmas are taken to represent partial melts of discrete mantle "reservoirs". An alternative, more unified view regards magma generation as a more or less pervasive upper-mantle process whose manifestations are controlled directly by lithospheric tectonics. In this vein, it is useful to follow Shaw and Jackson (1973), who deliberately avoided the term "hotspot" and its multiple connotations by using the term "melting anomaly" to describe a system of enhanced lithospheric magma production with compositions distinct from those of typical plate-boundary magmas.

This latter view is illustrated here in considering the origin of a linear system of volcanic features in the Northwestern United States that has commonly been viewed as defining a Yellowstone "hotspot" (Fig. 1).

Figure 1. Map of the northwestern US showing basalts younger than ~17 Ma and age-progressive rhyolitic centers. Age groups of basalts indicated by blue colors: brightest for 17-12 Ma, dark for 12-8 Ma, medium for 8-4 Ma, and light for 4-0 Ma. Ages of rhyolites younger than ~12 Ma indicated by red contours.



In the Yellowstone Plateau volcanic field, $\sim 6,500 \text{ km}^3$ of magma has erupted in the past 2.2×10^6 years, $\sim 95\%$ rhyolite and $\sim 5\%$ tholeiitic basalt. Episodic activity typically erupts 10 to 10^3 km^3 of rhyolitic magma every 10^3 to 10^5 years; the intrusive/eruptive ratio is estimated to be about 10. Basalts generated in the upper mantle cause partial melting of the lower continental crust and thermally sustain rhyolitic upper-crustal magma chambers. Magma production since 2-3 Ma has been $\sim 0.9 \text{ m}^3/\text{s}$ for Yellowstone, comparable to $\sim 3 \text{ m}^3/\text{s}$ for Hawaii. The Yellowstone Plateau volcanic field represents the youngest segment of a system that has propagated nearly 600 km across a continental plate along the eastern Snake River Plain since at least ~ 13 Ma and probably has emplaced 10^4 km^3 of rhyolitic and basaltic magma within and onto the continental crust. The age-progressive magmatic system was preceded by the 17- to 14-Ma basaltic and bimodal magmatism of the Columbia River flood basalts and other basalts and rhyolites, probably aggregating $\sim 10^5 \text{ km}^3$, in a backarc region adjacent to old, thick continental lithosphere that includes an Archean craton.

The widely espoused deep-mantle plume hypothesis for the Yellowstone melting anomaly seems deficient in several respects. Seismic studies fail to reveal any evidence for a deep-mantle plume and, in fact, point to the absence of any such feature beneath Yellowstone or the eastern Snake River Plain below ~ 200 km. The hypothesis also fails to address inherently the initiation of the melting anomaly in both spatial and temporal coincidence with a major mid-Miocene tectonic reorganization of a large region of the Western U.S., from localized extension along the Northern Nevada rift zone to accelerated uplift and regionally distributed extension of large magnitude. Similarly, the asymmetric distribution of flood basalts and related magmatism at the initiation of the Yellowstone melting anomaly, simultaneous NE propagation of the Yellowstone melting anomaly and NW propagation of the Newberry melting anomaly, and continued basaltic magmatism along both tracks all require *ad hoc* modifications of the plume hypothesis. We here interpret this system as an integrated product of regional tectonic interactions between the lithosphere and sublithospheric upper mantle, exploiting preexisting lithospheric structures

Major structural features, some cutting the entire lithosphere, were important controls on location and orientation of the track of the Yellowstone melting anomaly along the eastern Snake River Plain. The track parallels a pervasive NE-trending tectonic grain of Precambrian ancestry, including magnetic anomalies beneath the Great Plains of eastern Montana, and the Cheyenne Belt, Saint George trend, Colorado Mineral Belt, and Jemez trend to the south. More locally, numerous NE-trending tectonic features either straddle or align with the track of the Yellowstone melting anomaly, including: 1) the southwestern part of the pre-Tertiary Humboldt lineament, 2) an Early Proterozoic suture between the Archean Wyoming and Hearne Provinces, partly expressed by mylonite zones in the Beartooth Mountains, 3) the Proterozoic Great Falls tectonic zone, 4) a Late Cretaceous structural recess and consequent transverse sediment dispersal system in the Sevier orogenic belt, and 5) a conspicuous change in the orientations and expressions of Laramide foreland contractional structures.

We present a model for the Yellowstone melting anomaly that involves exploitation of preexisting tectonic features to accommodate oblique extensional strain across the Intermountain West since the mid-Miocene and strong contrasts in lithospheric architecture to initiate voluminous magmatism. In this model, lithospheric tectonic processes, including subduction, northward migration of the Mendocino triple junction, basin-range extension, and progressive outward concentration of basin-range extension toward the east and west margins in response to oblique extension and shear interact with sublithospheric upper mantle to produce the NW-propagating Newberry and NE-propagating Yellowstone melting anomalies. The volume of melting depends on source fertility, composition, focusing, prior history, the effects of horizontal temperature gradients, and thermal feedback on the dynamics of convection and melting in the upper mantle. Yellowstone and Newberry represent self-sustaining melting anomalies propagating along transform accommodation zones that bound a region of oblique large-magnitude extension. Exploitation of preexisting northeast-trending structural weaknesses along with basal shear melting of the southwest-moving North America plate and thermal feedback as magmas rise into the lithosphere and onto the surface may explain the exceptionally high magmatic productivity of the Yellowstone melting anomaly. Enhanced extension along the western margin of the Great Basin region produced crustal melting in the backarc of the Cascadian subduction system, propagating northwestward with extensional widening to produce the Newberry melting anomaly.

Thus, the lithosphere, asthenosphere, and deeper mantle interact as a dynamic feedback system that includes the generation and transport of fluid phases. Basaltic upper-mantle magmas may cause lower-crustal melting and can sustain long-lived upper-crustal magmatism. Lithospheric magmatism has long-term average rates that are coupled to tectonic displacement rates. Compositions reflect the physical state and composition of parent materials (including prior additions and subtractions by fluid phases), percentage of partial melting, rates of magma production and transport, and modification by phase separations during ascent and by thermal interactions with the lithosphere, including induced melting. All of these factors are influenced by stress distributions and ascent rates. Whereas oceanic magmas evolve in more or less transformed mantle peridotite and thin mafic crust; continental magmas evolve in mantle peridotite that may be ancient and in thick, heterogeneous but predominantly intermediate-composition crust. In such a view, plate-margin tectonic processes may simply act to focus more general upper-mantle magmatic processes also expressed in intraplate magmatism.

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The search for a primitive magma at Mount Vesuvius: possible role of a MORB-derived picrite in the genesis of Vesuvian magmas.

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Abstract

Near-primary melts, preserved as microinclusions in olivine of ultramafic ejecta (dunite and pyroxenite bearing *mosaic-equigranular* or *porphyroclastic textures* typical of mantle peridotites) of Mount Vesuvius, have a trachybasaltic, basanitic, and rarely trachybasaltic and picrobasaltic compositions (Mg# up to 68-75).

Thermobarometric estimates, obtained by using a grid of selected reactions, indicate that the *hydrous picrobasaltic melt* is in equilibrium with olivine and clinopyroxene at higher pressures: i.e., at about 17-18.5 kbar and temperatures of 1240-1300 °C, whereas a *moderately hydrous basanitic melt* is in equilibrium with the above phases at lower pressures of 11-14 kbar for temperatures ranging 1230-1350 °C. Pressure estimates are consistent with the geophysical data that indicate the crust-mantle transition, below Mount Vesuvius is at about 30-36 km.

Trace element data indicate moderate to high contents in incompatible elements for the *ultramafic ejecta* (Rb/Sr range 0.19-0.24), with enrichments up to about two order of magnitude above *primordial mantle values*. In particular spider diagrams indicate strong positive anomalies for Ta and P, whereas some sample may exhibit strong negative anomalies in Hf. Moreover, these spiders diagrams show similar patterns between the ultramafic ejecta and recent Vesuvian lavas which suggest a common source. REE patterns for *dunites* are relatively regular with moderate enrichments in LREE, and HREE contents comparable or slightly above average chondrites: (La/Yb)_N varies between 7 and 16. *Clinopyroxenitic nodules*, with (La/Yb)_N ranging 4.8-6.4, show a slightly higher degree of enrichment in terms of LREE and a relatively “flat” pattern for these elements. HREE distributions, normalized to chondrites, decrease progressively

with increasing atomic number, and show an enrichment of 1.5 to 3 times higher than average chondrite.

Isotope data obtained on ultramafic vesuvian nodules ($^{87}\text{Sr}/^{86}\text{Sr}$ range from 0.70691 to 0.70728; whereas $^{143}\text{Nd}/^{144}\text{Nd}$ range 0.51248-0.51254) confirm a close genetic link between these materials and the lavas which have rather similar isotope ratios. Isotopic parameters, such as $\epsilon_{\text{Sm}/\text{Nd}}$ (calculated at 2 Ga, according to De Paolo, 1988) show values of 0.42-0.59 both for the lavas and the ultramafic nodules. These values are consistent with 1-10% partial fusion of an eclogitic source. Moreover, $\epsilon_{\text{Sm}/\text{Nd}}$ parameters (Salters and Hart, 1989; Salters, 1996) constrain the degree of partial fusion at 5-7 %.

Mass balance calculations on major elements have been cross-checked iteratively by utilizing the computer code MELTS (Ghiorso and Sack, 1995) to identify a possible primary parental magma. Such a primitive melt is a picrite which fractionates garnet + clinopyroxene at high pressure at/or in proximity of its source region (≈ 28 kbar), followed by a moderate decompression coupled with fractionation mainly of olivine at moderately lower pressure (≈ 18 kbar) to give a picrobasaltic melt (with a fraction of melt remaining, $F \sim 0.66$). At lower pressures (approaching 12 kbar) the partly evolved parental will continue separating slightly variable amounts of olivine, clinopyroxene, spinel as well as fractions of apatite to generate basanite ($F \sim 0.12$), tephrite ($F \sim 0.06$) and phonotephrite ($F \sim 0.04$). Trace element abundance of the above primary picritic magma have been retrieved by “backstripping” the trace element distributions found within the lavas (following a two-step model to include P-T variations of selected partition coefficients for the mineral phases involved in the process). The calculated REE pattern of the “primary picrite” is a typical MORB approaching a “flat” pattern with a degree of enrichment of LREE and HREE above 10-20 times average NMORB.

Partial fusion models, constrained by phase relationships at high pressure (according to selected sections of the O'Hara tetrahedron) allowed identification of an original NMORB source. This inference suggests that Vesuvian magmas can be generated by small degree of partial fusion of a MORB derived eclogitic source (i.e., a garnet pyroxenite according to O'Hara's and Yoder's definition) which is likely represented by the remnants of the subducted oceanic crust. Therefore,

it is not excluded that mantle metasomatism (occurring with a mechanism consistent with the model proposed by Peccerillo, 1999) of subducted oceanic crust is eventually followed by slab melting below Mount Vesuvius. This petrogenetic process could play a significant role in the genesis of other magmas within the Roman Comagmatic Province and within the Southern Tyrrhenian region.

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The Origin of High-Ti Picrites from the Ethiopian Flood Basalt Province

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The Ethiopian Plateau forms the larger part of an extensive continental flood basalt province, which, today, extends into Eritrea, Djibouti, Yemen and Somalia. The province is thought to be a result of plume-head activity associated with the opening of the Red Sea and Gulf of Aden at the Afro Arabian triple junction. Tomographic images of mantle shear-wave velocities beneath Afar seem to confirm the existence of a mantle plume that extends down to the core mantle boundary (1), and elevated $^3\text{He}/^4\text{He}$ ratios for high-Ti lavas from the province support that this is sourced in an undegassed part of the mantle (2). The main phase of volcanism, which formed the traps occurred between 30 Ma and 28 Ma (3). During this period, a pile of flood basalts with an estimated thickness of up to 2000 m, covering an area of 600 000 km² (4) was erupted onto a basement of Pan-African arc-associated volcanics and intrusives, and Mesozoic sediments (5). 106 km³ of this volume were emplaced within 1 myr during the peak of the activity at about 30 Ma; this indicates an average magma flux of 1 km³/yr (3). Combined fission track data and $^{40}\text{Ar}/^{39}\text{Ar}$ age-data, together with field observations, suggest that uplift preceded volcanism and that extension was synchronous with volcanism (6). Extension in the Red Sea occurred in two distinct pulses at 30 to 35 Ma and 20 to 25 Ma respectively (7) and oceanic crust appeared around 10 to 13 Ma (8), or more recently at 5 Ma (9).

The flood basalts of the Ethiopian Traps are transitional between tholeiitic and alkaline in composition. They form three distinct magma groups, which show geographical rather than temporal variation in their major and trace element composition (5). The LT basalts, which occupy the north-west part of the plateau, show consistently low TiO₂ contents and considerable heterogeneity in their trace element geochemistry. Marked troughs for Th and Ta-Nb, and variable LREE enrichment may suggest that they have undergone extensive lithospheric contamination, and that they are derived from a LREE depleted garnet free source. The HT2 basalts, which occur in the south-east part of the plateau, on the other hand, have consistently high TiO₂ contents and more homogeneous trace element signatures. Enrichment of LREE and a relative depletion of HREE suggest the presence of residual garnet in the source, and there is little evidence for contamination. The HT1 basalts display characteristics between the two other groups. Contrasting $^3\text{He}/^4\text{He}$ ratios in the HT2 and LT basalts indicate that the two types are from different parental sources (2). The HT2 basalts are typically picritic with high $^3\text{He}/^4\text{He}$ ratios and they so define a geographically localised high-Ti sub-province representative of the primary magma generated by high temperature melting possibly within the head of a plume.

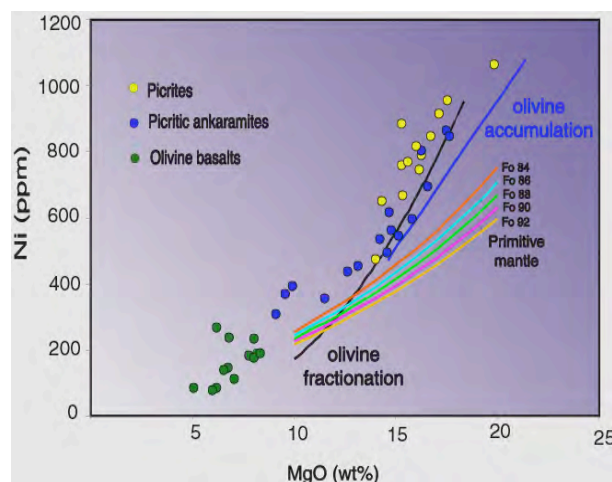


Figure 1 Plot for MgO (wt%) versus Ni (ppm) for the lavas from the high-Ti sub-province. The curves for olivine accumulation and fractionation were respectively modelled from major element data for the least and most primitive picrite. Lines for melts in equilibrium with mantle olivine of Fo84 - Fo92 were calculated using partition data from (10) and (11).

The lavas of the high Ti sub-province include picrites, ankaramites and olivine basalts, which together form a coherent suite genetically related by olivine fractionation and accumulation (Fig. 1). They have a primary melt composition of between 14 and 15 % MgO equilibrated with mantle olivine with a festerite content of 82. This relatively low Fo content is a consequence of the high iron content of the source, which is elevated compared to normal mantle, and high Fe₁₅ for the lavas compared to experimental data for KLB-1 and HK-66 seem to confirm this. High iron (12 - 16 % Fe₂O₃) and titanium (3 - 7 % TiO₂) and a primary MgO of 14 - 15 % imply that the parent magma was derived

from a high temperature small melt fraction. CaO/Al₂O₃ to Al₂O₃ ratios plotted against experimental data for KLB-1 (12) suggest that the melts were generated at between 4 and 5 GPa (Fig. 2).

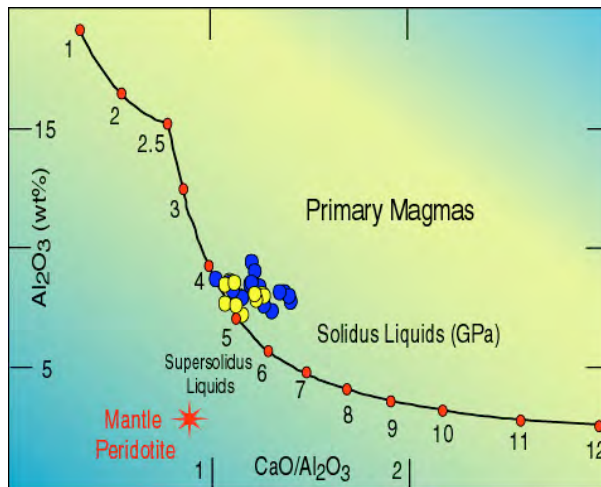


Figure 2 Plot of Al₂O₃ versus CaO/Al₂O₃ showing pressure of melting for the picrites and ankaramites from the the high-Ti sub-province. The lavas are compared to experimental data for KLB-1. Liquids formed along the solidus will have Al₂O₃ and CaO/Al₂O₃ contents indicated by the pressures in GPa (12).

These pressures are equivalent to depths of 120 - 150 km, and imply melting at temperatures between 1600 and 1700°C beneath thick lithosphere prior to extension. The random stratigraphic order of the primitive picrites and ankaramites, and the more evolved olivine basalts within the volcanic pile further implies that pulses of primary magma were released into the crust, some of which made their way directly to the surface while some were able to equilibrate in a shallow-seated reservoir before eruption.

The consistent homogeneity of the trace element patterns for the lavas of the high-Ti sub-province seems to support their common parentage, and strongly fractionated REE patterns indicate the presence of garnet in the source. Sub-parallel HREE patterns for the olivine basalts confirm that they fractionated at a shallow level compared to the picrites and ankaramites which show a distinct fanning of the REE indicative of variable degrees of partial melting within the garnet stability field. Dy/Yb and La/Yb ratios can be used to constrain the depth and degree of melting within the garnet field. Using pressure, temperature and composition dependent coefficients and phase proportions, the lavas can be modelled as 2 -3 % partial melts at pressures in excess of 3 GPa. The picrites and ankaramites define a discrete linear array for Dy/Yb versus La/Yb, and Dy/Yb versus Si₁₅; this reflects a range of melt conditions from moderate to high pressures with increasing garnet signatures.

The major and trace element data all support a model for melt generation at high temperatures and pressures. In particular the high temperature of melting, well above that expected for normal variations in the mantle, is indicative of melting within a plume.

Initial ¹⁸⁷Os/¹⁸⁸Os ratios for the picrites and ankaramites are all sub-chondritic with a restricted range (0.125 - 0.126). This reflects a source more like that expected for depleted mantle; in this respect they are quite different from any other reported plume-head lavas (Fig. 3). Total Os concentrations for the lavas are high (750 - 2025 ppt) compared to other continental flood basalts; they are more similar to komatiites. Since it is unlikely that the source contains no sulphide, this elevated Os concentration may be a feature of high pressure/temperature partitioning of Os.

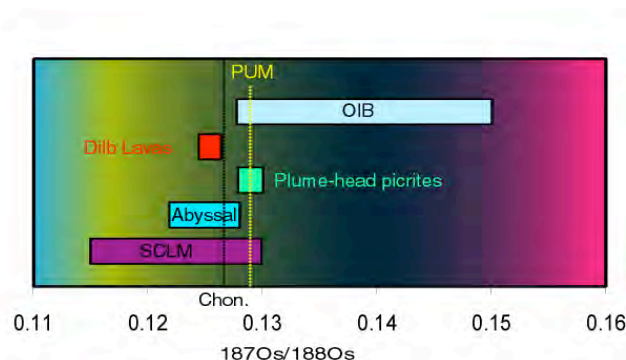


Figure 3 ¹⁸⁷Os/¹⁸⁸Os range for the Ethiopian picrites and ankaramites compared to selected mantle derivatives.

Sub-chondritic $^{187}\text{Os}/^{188}\text{Os}$ ratios suggest that there is no excess ^{186}Os in these high-Ti lavas; it is therefore unlikely that they have interacted with the core as is implicit in the tomographic images presented by Ritsema et al. (1). More recent tomographic images presented by Nolet et al. (13), however, affirm that the Afar upwelling is rooted in the upper mantle.

The Os isotope composition of the high -Ti lavas from the Ethiopian plateau is intriguing because they have values similar to what would be predicted for the depleted upper mantle. However, recent studies of both MORB and OIB lavas, and estimates of the Os isotope composition of the mantle, indicate that there is no simple delineation between sub-chondritic upper mantle and more primitive lower mantle. Some plumes have very low $^{187}\text{Os}/^{188}\text{Os}$ and some MORB have values within error of estimates for the primitive mantle. In conclusion then, although the Ethiopian lavas have Os isotope ratios similar to upper mantle their actual origin within the mantle is still elusive. What is clear, is that they are from a source significantly more homogeneous than that for MORB.

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Geochemical Structure of the Hawaiian Plume: Results from the Hawaii Scientific Drilling Project

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The major focus of the Hawaii Scientific Drilling Project is the recovery and characterization of a continuous set of lava samples from a single Hawaiian volcano. These samples provide a unique opportunity to test aspects of the mantle plume hypothesis. In 1993 a 1.06 km pilot corehole was drilled into the flank of the Mauna Kea volcano at Hilo (Stolper et al., 1996). In 1999, a 3.1 km corehole was drilled at a site 2 km south of the site of the 1993 drilling with nearly complete core recovery. In 2003 the hole is being deepened with the hope of reaching 4.5 km. The project is funded by the U.S. National Science Foundation and the International Continental Drilling Program (ICDP; <http://www.icdp-online.de/>).

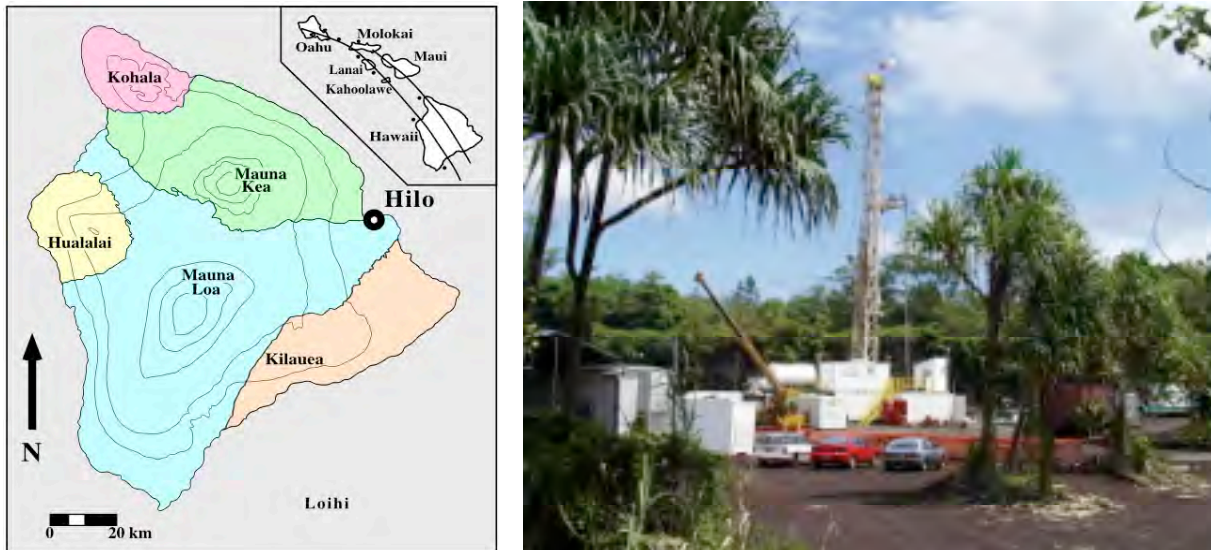


Figure 1. (left) Location map for the Hawaii Scientific Drilling Project 1.06 km pilot hole and 3.1km deep hole. (right) Drill rig in operation in Hilo in 1999.

The drillsite was chosen to be (1) far from volcanic rift zones to minimize chances of encountering intrusive rocks, alteration, and high-temperature fluids, (2) close to the coastline, to minimize the thickness of subaerial lavas that would need to be penetrated to reach the older, submarine parts of the Mauna Kea section, and (3) in an industrial area to minimize environmental and community impacts. Although the Mauna Kea lava section was the primary target, this choice of site required drilling through a 280m - thick section of ML flows, thereby providing additional information on the Mauna Loa volcanic succession. Temperatures in the upper part of the hole are indeed low; circulation of seawater through the volcanic pile causes the temperature to decrease with depth to a low of 12°C at about 800m depth, whereupon it begins to increase to 45°C at 3.1 km. Alteration is minimal. Intrusives were successfully avoided; they make up less than 2% of the core, but the percentage increases gradually in the lowermost part of the core. Digital photographs of each core box, high-resolution scans of the core, a detailed lithological column, and detailed descriptions of the entire recovered core can be accessed at <http://icdp.gfz-potsdam.de/html/hawaii/news.html>.

A major focus of our effort has been a comprehensive geochemical and petrological characterization of the samples recovered from the core. In both the pilot hole and HSDP-2, a "reference set" of over 100 samples were distributed to investigators from around the world. In addition to the reference suite, which samples the section at 50-100 m intervals, for the submarine section fresh glass samples were prepared at intervals of about 3 m. A simplified lithologic section of the core and the available age data are shown in Figure 2.

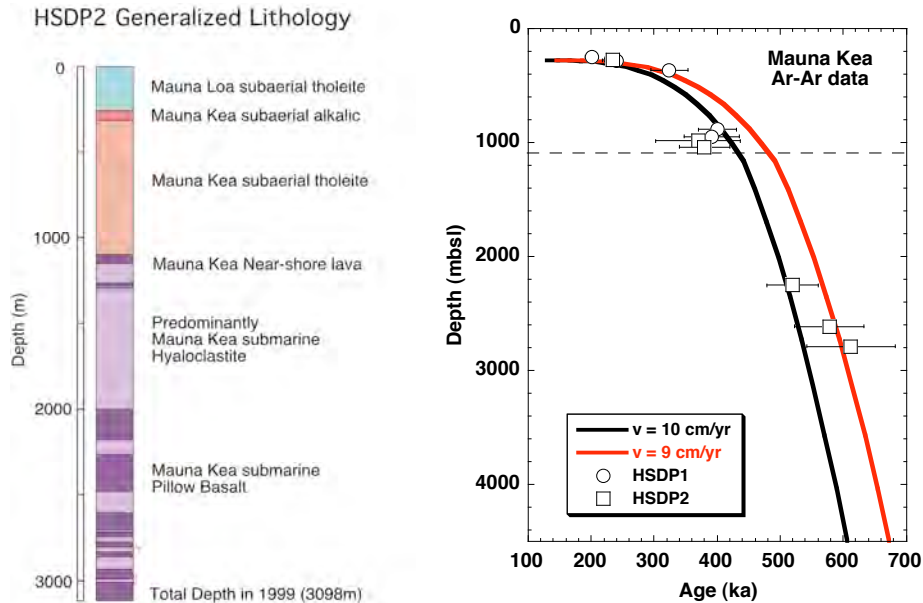


Figure 2; Generalized lithologic section of the 3.1 km HSDP core and Ar-Ar age data with 1sigma uncertainties (Sharp et al., 1996, 2003). At this scale depth in the hole (m) and depth in meters below sea level (mbsl) are approximately the same; the hole was drilled at an elevation 11m above sea level. The Argon age data are shown with two model lava accumulation rate curves based on the models of DePaolo and Stolper (1996). The parameter "v" is the plate velocity in the reference plane of the plume. DePaolo et al. (2001) speculate that the plume may be wobbling relative to a "fixed" reference frame.

The cored section includes about 250 m of Mauna Loa tholeiitic lavas varying in age from 2 to about 80 - 100 ka. The Mauna Kea section comprises subaerial flows between 250 and 1080 meters, mostly well-indurated basaltic submarine hyaloclastite from 1080 to 2000 meters, and pillow basalt with intercalated hyaloclastite from 2000 to 3098 meters. The Ar-Ar age data indicate that the age of lavas at 3098 m depth is about 600 ka. Hence the core plus subaerial exposures of Mauna Kea lavas provide a continuous record of the volcano for 600,000 years. Figure 3 shows the calculated track of the Mauna Kea volcano summit relative to a model Hawaiian plume melt production zone. The geochemical stratigraphy covers about 60% of the Mauna Kea traverse of the melting region, and encompasses the transition from tholeiitic to alkaline lavas, which is accompanied by a major decrease in eruption rate.

Examples of isotope stratigraphy, shown in Figure 4, in general show systematic shifts with age in the directions expected for a model of a radially zoned plume. Helium and ^{208}Pb show the largest shifts, but even the Nd shifts are systematic although small. Hafnium shows a deviation from expected behavior in the uppermost, alkalic and transitional lavas erupted between 300 and 200 ka. There is little change in the isotopic composition of the magma sources during the transition from tholeiitic to alkaline lavas, except possibly for Hf isotopes. The lava major element compositions are divisible into high- and low-silica groups (Rhodes et al., 2003, Stolper et al. 2003) isotopic differences between the groups are small.

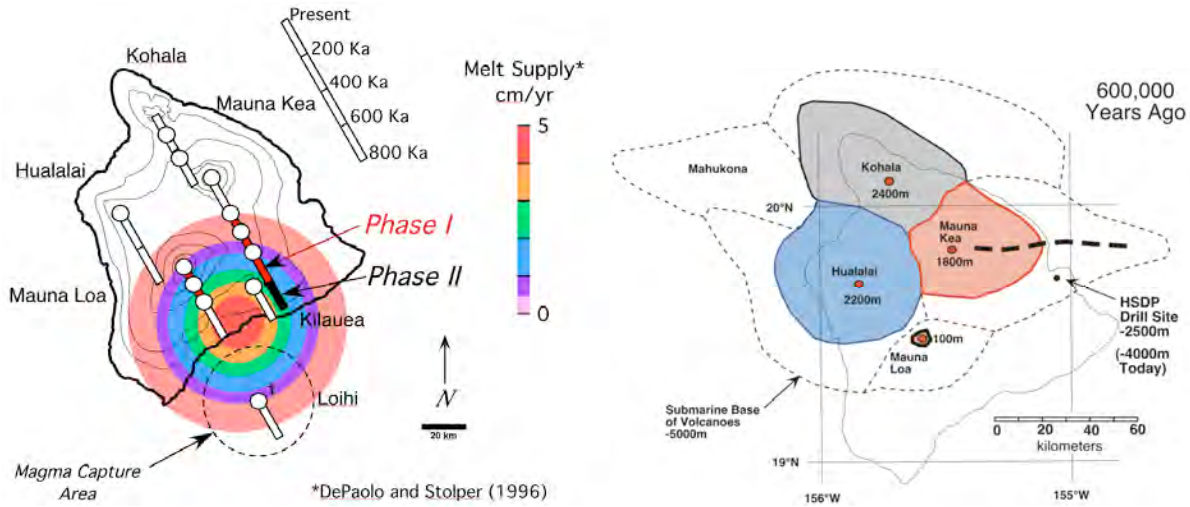


Figure 3: (left) Map showing inferred past positions of volcano summits relative to the melting anomaly that constitutes the Hawaiian mantle plume (DePaolo et al., 2001). At progressively greater times in the past, the volcanoes were farther to the southeast due to the movement of the Pacific plate. The HSDP core drilling retrieved lavas from Mauna Kea and Mauna Loa that represent the summit position tracks indicated by the filled lines colored in red. The black filled line segment shows the portion of the Mauna Kea track that will be covered by the next phase of HSDP drilling. (right) Probable configuration of the island at 600 ka, showing submarine location of the Hilo drill site.

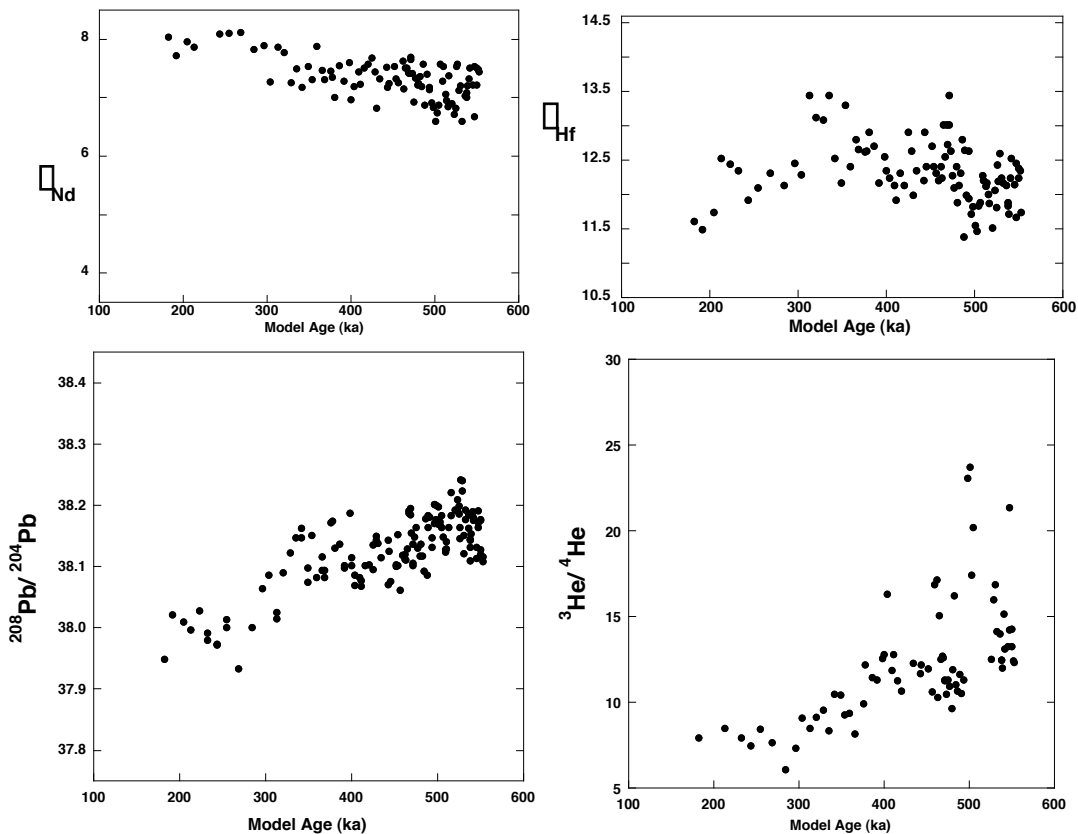


Figure 4: Plots of isotopic ratio versus age for lava samples from the 1999 HSDP2 core. The oldest lavas are at a depth of ca. 3090m below sea level. The total vertical extent of each diagram corresponds to the range of isotopic compositions reported for the volcanoes of the big island of Hawaii. (Data from Eisele et al., 2003; Blichert-Toft et al., 2003; Kurz et al., 2003; Bryce et al, 2003)

A model based on the melt generation pattern shown in Figure 3 is being used to evaluate the data. The objective is to account for the age and volume of the volcanoes of the Big Island, the lava accumulation rate of Mauna Kea (Figure 2), and the geochemical stratigraphy. The major uncertainty is in how the plume melts are sampled by individual volcanoes. We assign to each volcano the melts produced within a circular area of radius 25km centered on the summit (Fig. 3). This model is able to account for the volcano volumes as well as the geochemistry. The model lava accumulation rates are shown in Figure 2. The model requires that the radius of the geochemically anomalous material at the plume core is about 2/3 the radius of the melting region, and about 40% of the radius of the temperature anomaly. Comparison of data from Mauna Loa and Mauna Kea also shows that the plume core is longitudinally heterogeneous. There does not appear to be much longitudinal heterogeneity in the outer parts of the plume melting region, as the Kea trend volcanoes have a highly consistent isotopic composition over $> 10^6$ years. The radial structure of the plume may reflect the vertical structure of the mantle it has traversed, and implies that the lowermost mantle (plume source) is highly heterogeneous, whereas the mid-mantle is both less heterogeneous and more depleted, although not as depleted as the NMORB source.

Current models for the Hawaiian plume require upwelling velocities at the core of the plume that are upwards of 30 cm/yr, depending on the viscosity structure. To generate enough melt per unit time to construct the volcanoes, the maximum potential temperature (T) must be about 1550°C or somewhat higher, and the maximum extent of melting (F) must be about 20%. There is some tradeoff between upwelling velocity versus T and F . Both the major element compositions of the HSDP lavas (Stolper et al., 2003) and the trace elements (Feigenson et al., 2003; Huang and Frey, 2003) are consistent with the geodynamically-inferred parameters.

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How much heterogeneity in the mantle MORB source?

Henry Dick

It has long been known that there are primary compositional gradients in major and trace element composition in the residual mantle dredged from slow spreading ocean ridges. These gradients have been correlated directly to the major element composition of spatially associated basalts showing that the most depleted peridotites and refractory basalts are associated with mantle plume regions in the Atlantic and Indian Oceans. The peridotites used to show this correlation come almost entirely from fracture zones, and thus represent a sample skewed to the distal end of a mantle melting cell beneath a ridge segment. Now peridotites have been dredged in abundance from the Gakkel and SW Indian Ridge far from any mantle hot spot, and far from any transform in regions that should reflect the lowest degree of mantle melting anywhere in the oceans: ultraslow spreading ridges. Analysis of spatially associated MORB and peridotites from one of these locations, 9°-16°E on the SW Indian Ridge show that the basalts are consistently less depleted isotopically than the peridotites, implying that they represent very low degrees of melting of a depleted mantle hosted enriched vein assemblage. Many of these peridotites, however, appear to represent moderate degrees of mantle melting, despite the fact that they occur in regions where there is little or no crust, and the mantle has spread onto the sea floor for long periods. This, then raises two interesting questions: to what are the relative mass contributions of veins and host mantle to basalt generation? and How heterogeneous is the primary mantle source before melting. The new evidence suggests that both these questions remain unanswered .

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ON THE WESTWARD DRIFT OF THE LITHOSPHERE

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There have been a number of papers describing the westward drift of the lithosphere relative to the mantle (e.g., Bostrom, 1971), which is proven by independent data sets such as the hotspot reference frame (e.g., Ricard et al., 1991), space geodesy and asymmetries of subduction and rift zones (Doglioni et al., 1999; 2003). Gripp and Gordon (2002) computed an average net rotation of the lithosphere of up to 4.9 cm/yr, using classically accepted hotspot reference frame.

There still are doubts about 1) what is generating the westward drift, and 2) whether it affects the entire lithosphere or it is rather only a mean value, with most of the lithosphere moving "west" due to the dominant effect of the Pacific plate, but part of it still moving in the opposite direction relative to the mantle.

Paradoxically, the hotspot reference frame in which the westward drift was originally computed, is the strongest limitation to the westward drift itself. In fact there are plates that move eastward relative to hotspots, suggesting a mantle moving westward, such as the Nazca plate. However plates, which would move eastward relative to hotspots, they still move westward relative to Antarctica (Knopoff and Leeds, 1972).

We question the hotspot reference frame because it includes volcanic tracks permanently located on ridge zones that are detached from the underlying mantle by kinematic constraints. It is evident that oceanic ridges surrounding Africa moved away from the craton during the opening of the Atlantic and Indian oceans. This indicates that hotspots steadily located on oceanic ridges cannot be considered as fixed with respect to any global reference frame. Moreover there are a number of growing evidences suggesting that the so-called hotspots on which the 'absolute' plates motion has been computed, are rather superficial features, sourced either from the asthenosphere, or the lithosphere base itself (Bonatti, 1990; Smith and Lewis, 1999; Anderson, 2000; 2001; Harpp et al., 2002), possibly unrelated to deep mantle sources.

Therefore the asthenospheric or lower lithospheric mantle sources are unfixed relative to each other, and the hotspots appear not be a suitable reference frame for plates motion. When recomputed the westward drift disregarding those hotspots (e.g., Africa, Galapagos, Easter Island, etc.), the net rotation is affecting the whole lithosphere and it amounts to about 9 cm/yr, becoming one of the most intriguing elements of geodynamics. Moreover, geological and geophysical signatures of subduction and rift zones show a global signature, suggesting a global relative 'eastward' motion of the mantle relative to the lithosphere. Plates are moving along a sort of mainstream, which is not E-W, but it rather depicts a sinusoid (Doglioni et al., 1999). Along this flow, west-directed subduction zones are steeper than the E- or NE-directed, and the associated orogens are respectively characterized by lower structural and topographic elevation, backarc basin, and in the other side by higher structural and morphological elevation and no backarc basin. The asymmetry is striking when comparing western and eastern Pacific subduction zones, and it has usually been interpreted as related to the age of the downgoing oceanic lithosphere, i.e., older, cooler and denser in the western side. However this interpretation fails because these differences persist elsewhere regardless the age and composition of

the downgoing lithosphere, e.g., in the Mediterranean Apennines and Carpathians vs Alps and Dinarides, or in the Banda and Sandwich arcs, where even continental or zero-age oceanic lithosphere is almost vertical along west-directed subduction zones.

Rift zones are also asymmetric, being in average the “eastern” side more elevated of about 100-300 m worldwide (Doglioni et al., 2003). A test of these asymmetries is in Fig. 1 where the mean topographic and bathymetric elevations are reported for West vs East- or Northeast-directed subduction zones, and for the western and eastern flanks of rift zones.

Mantle anisotropy also provides evidence of mantle flow, e.g. beneath western North America (Silver and Holt, 2002) and beneath the southern Pacific rise (Wolfe and Solomon, 1998). Kennedy et al. (2002) have shown how mantle xenoliths recorded a shear possibly located at the lithosphere-asthenosphere interface. This supports the notion of a flow in the upper mantle and decoupling at the base of the lithosphere.

The westward drift of the lithosphere can result from the Earth’s rotation (Scoppola et al., 2003) assuming three coexisting processes (Fig. 2): i.e., 1) Earth’s rotation is decelerating mainly due to tidal torques; 2) The downwelling of the denser material toward the bottom of the mantle and in the core opposingly provides an acceleration of the Earth’s rotation, partly counterbalancing the tidal drag; 3) The decoupling at the base of the lithosphere occurs over a convective underlying mantle, with ferromagnetic properties, where both phenomena decrease relative friction. The variable relative “westward” decoupling between lithosphere and asthenosphere may explain plate tectonics.

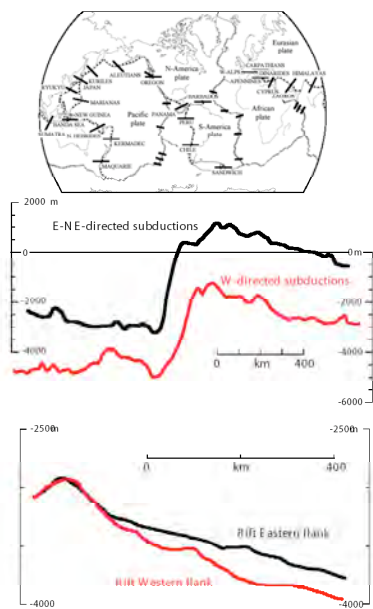


Fig. 1. Mean elevation profiles of subduction and rift zones. Both geodynamic settings show higher topography and bathymetry in the eastern or northeastern side providing evidence for a geographic polarity in global tectonics.

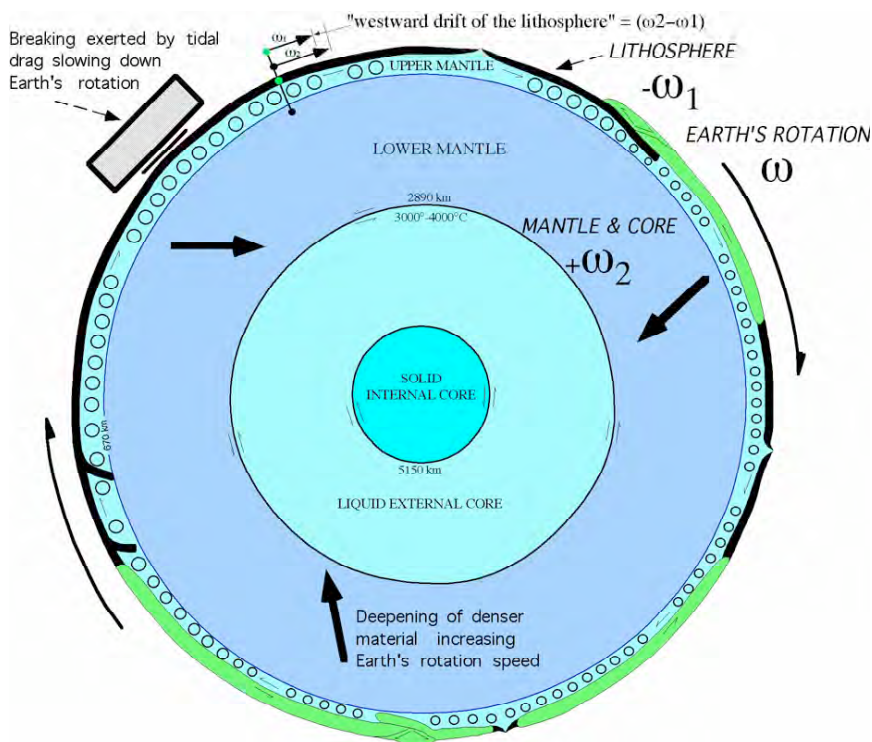


Fig. 2. Model Earth in which the westward drift is interpreted as due to different processes such as the break exerted by the tidal drag, the downwelling of heavier masses toward the interior of the Earth, and simplified rolling convection in the upper mantle related to the toroidal flow of the lithosphere. In order to decouple the lithosphere, the coexistence of these processes needs a viscosity of the asthenosphere close to the values we presently know.

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Global Seismic Tomography: What we really can say and what we make up

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At a recent meeting, someone said that it is necessary to know what are the strong constraints, weak constraints and myths. There are too many myths to discuss them in this brief presentation; it will focus on the “strong constraints”. Even though some of the classification may appear subjective, it is based on sound principles. It is also important to remember that tomography provides only a snapshot of the present velocity anomalies. Temporal extrapolations, which lead people to “see” “subducted slabs” in the lower mantle, are made at significant risk. The same is true, to a lesser extent, with respect to attributing the anomalies to thermal or compositional effects.

The observed anomalies in seismic data result from integration of perturbation in structure either along the ray path, for travel times, or volume for splitting of low order normal modes. Perturbation is a smoothing operation and, assuming constant amplitude, the effect of the component of the structure with a wavelength l_1 compared to that of l_2 is equal to the ratio l_1/l_2 . Thus long-wavelength components of the structure are much easier to recover than short-wavelength ones. It is difficult to imagine that a narrow vertical structure (plume?) of 100 km width could be mapped using teleseismic observations, since commonly observed phases such as P , S , PP or SS have rather large incidence angles in the lower mantle. Its image would be significantly widened and the amplitude greatly diminished. In addition to this smoothing effect, there is also the fact that the spectrum of heterogeneity in the upper and lowermost mantle is distinctly “red”, i.e. dominated by large wavelength components [*Su and Dziewonski, 1991; Su and Dziewonski, 1992*], which may tend to further impair recovery of short-wavelength features.

Therefore, the large-scale features are most easily recovered and most reliable, which is supported by the fact that, for example, most of the recent whole-mantle S -velocity models are quite similar [*Gu et al., 1998; Ritsema et al., 1999; Masters et al., 2000; Megnin and Romanowicz, 2000*]. The following text represents a summary of “strong constraints” for several regions of increasing depth.

Lithosphere and Asthenosphere

Credibility of seismic tomography has been to large extent established by the fact that both mantle wave dispersion maps [*Nakanishi and Anderson, 1983*] and models of upper mantle shear velocity anomalies [*Woodhouse and Dziewonski, 1984*] closely matched predictions from surface tectonics [*Nataf and Ricard, 1996*]. There are exceptions to this: the depth to which seismic velocity anomalies persist as a function of age of the oceanic lithosphere (all mantle models cited above) and the strong radial anisotropy in the Central Pacific.

Surface wave observations, particularly when expanded to include relatively short period (down to 35 s) Love and Rayleigh wave data, demand that age-dependent velocities vary to depths as large as 200 km, while the cooling plate model predicts the thickness of about 100 km. This is particularly clear in the V_{SH} model of Ekström and Dziewonski [1998] at a depth, for example, of 150 km, where the velocities monotonically increase from -4% at the East Pacific Rise, to $+3\%$ in the Western Pacific. The V_{SV} part of the model shows an added complexity in the Central Pacific with a secondary minimum (about -3%) just southwest of Hawaii. The decomposition of the model into an isotropic (Voigt average) and anisotropic part ($V_{SV} - V_{SH}$) parts clarifies the picture. The isotropic model for the Pacific plate is entirely consistent with the cooling with age (except for the segment between EPR and “Superswell”, which is characterized by low velocities and – presumably – higher temperatures at 150 km depth. The anisotropic model, however, is close to zero underneath the EPR, but shows a distinct minimum some 500 km southwest of Hawaii. The anisotropy rapidly decreases between 150 and 250 km depth. The explanation is likely to involve a significant vertical influx of material under the Darwin Rise.

The presence of radial anisotropy has also been reported under the cratons [Gung *et al.*, 2003], and it might be able to explain the long-standing controversy on the depth of the continental roots [Jordan, 1975]. Gung *et al.* [2003] suggest that the velocity anomaly associated with cratons ends at about 200-250 km, but is underlain by transversely anisotropic material with $V_{SH} > V_{SV}$.

Transition Zone

It is difficult to think of a more controversial geophysical problem than the flow of the mantle material across the boundary between the upper and lower mantle. However, there is now good evidence for strong impedance to flow across this boundary. It is based, in part, on the estimates of the patterns of large-scale velocity anomalies above and below the 650 km discontinuity. Gu *et al.* [1998] show that it is possible to satisfy a large and diverse set of travel time data with a model that has distinctly different spectral characteristics above and below the 650 km discontinuity. Even when the continuity is imposed, the radial correlation function [Jordan *et al.*, 1993] shows that the velocity anomalies 100 km below and above the 650 km discontinuity are fully de-correlated, indicating a very steep gradient in the pattern of heterogeneity.

Additional information comes from studies of the topography of the 410 and 660 km discontinuities, using $SS - SdS$ differential precursor times [Shearer and Masters, 1992; Gu *et al.*, 1998]. Simultaneous inversion for mantle shear velocity and topography of the transition zone discontinuities shows very little large wavelength topography on the 410 km discontinuity and much larger relief on the 650 km discontinuity [Gu *et al.*, 2003]. Also, the shape of the two surfaces is de-correlated [Gu *et al.*, 1998]. However, the 650 km discontinuity topography is highly correlated with the velocity anomalies in the transition zone. A simple, although not unique, interpretation is that the negatively buoyant slabs pond in the transition zone.

A shortcoming of seismic tomography is that it provides only a snapshot of the heterogeneity; attempts to extrapolate it in time are uncertain. But the ponding of slabs and - judging from volume of the anomalies - a substantial residence time, does indicate the possibility that this

material could be, at least in part, re-circulated in the upper mantle; not unlike the megalith proposed by Ringwood and Irifune [1988].

Middle Mantle

Whether or not one accepts the result of Gu *et al.* [2001], there is general agreement that, at least, *S*-wave velocity models show a distinctly different spectrum in the middle mantle (from 650 km to about 2500 km); the power spectrum is nearly flat as a function of harmonic degree and, generally, much weaker than in the upper mantle. Tackley *et al.* [1994] obtained such a change in spectral characteristic in their numerical simulation of mantle convection, in which there is discontinuity in flow across the 650 discontinuity caused by an endothermic phase transformation. There may be, of course, other causes of such a disruption of flow, but the spectral effect is eerily similar.

In many papers there have been suggestions of additional discontinuities in the lower mantle. It appears that on the global scale there is not such discontinuity with a specific radius. Even a discontinuity with a several hundred kilometers of topography and 0.5% velocity change is highly unlikely, judging from the behavior of slowness (dT/dD) of *P*- and *S*-waves as a function of distance, which change between 35° and 90° in a remarkably linear fashion [Dziewonski *et al.*, 1993]. Gu *et al.* [2001] experimented with allowing an abrupt change of heterogeneity pattern across 1000 km [Kawakatsu and Niu, 1994; Fukao *et al.*, 2001] and 1800 km [van der Hilst and Karason, 1999]; in neither case there has been a change in the pattern of the power spectra. It appears therefore, that the middle mantle has a relatively low level of heterogeneity, distributed rather evenly among the spherical harmonics up to degree 20. This does not mean that there could not be local discontinuities, although local perturbations in the gradient may cause caustics giving an appearance of a discontinuity [Tromp and Dziewonski, 1998].

The middle mantle is where it is widely believed that the coherent images of subducted slabs are visible. Of these, perhaps the most famous is the “Farralon slab” [Grand *et al.*, 1997; van der Hilst and Karason, 1999], which seems to penetrate lower mantle from the 650 km discontinuity to the core-mantle boundary in both *P*- and *S*-velocity models. However, models derived with a greater variety of data types and, consequently, more even volume sampling show large-scale high velocity features in places where there was no subduction in the past 200 Ma, do not show it in places where subduction took place and in the case of the South Sandwich Island the “slab” occupies a volume that is completely out of proportion in comparison with the duration and rate of subduction in this region [Bijwaard *et al.*, 1998]. While it is likely that there is some link between the upper and lower mantle flow, the nature of this link is not clear. Recently, Pysklywec *et al.* [2003] proposed that the dynamics in the Fiji-Tonga-Vanuatu region during the last 10 Ma might have been caused by an avalanche of the ponded slab material under the southern part of the region. In terms of seismic tomography, the region is characterized by relatively normal velocities in the transition zone, but a large volume of faster than average mantle between 650 and 1100-1200 km depth. In their model, the process is still ongoing and may present a unique opportunity of monitoring an “avalanche” in progress.

Lowermost Mantle and Core-Mantle Boundary

It has been discovered a fairly long time ago that the level of heterogeneity increases near the core-mantle boundary (CMB) [Julian and Sengupta, 1973; Dziewonski *et al.*, 1977]. This increase may be present at all wavelengths, but it is very clear at the gravest harmonics: degrees 2 and 3 [Dziewonski, 1984], which strongly dominate global maps at some 150 km above the CMB. The geometry of these ultra-long anomalies is very similar to the corresponding harmonics in the geoid field, but have the opposite sign [Dziewonski *et al.*, 1977]; the fact that led to hypothesis that then density variations are proportional to velocity variations; the sign of the proportionality coefficient positive if dynamic deformation of boundaries is taken into account [Hager *et al.*, 1985]; this would be consistent with a thermal rather than compositional origin of the anomalies. There is an overall agreement between the predicted [Hager *et al.*, 1985] and observed [Morelli and Dziewonski, 1987] topography of the CMB [Dziewonski *et al.*, 1993].

The geometry of the large-scale velocity anomalies is simple and bears a relationship to the surface tectonics. The higher than average velocities are arranged in a ring circumscribing the Pacific basin, with a slight landward shift, thus potentially reflecting accumulation of the subducted material. There are two main low velocity regions, sometimes called superplumes [Romanowicz and Gung, 2002]: one under Africa (Great African Plume) [Dziewonski *et al.*, 1993] and the other one under Pacific (Equatorial Pacific Plume Group), [Dziewonski *et al.*, 1993]. Nearly all hotspots (over 90%) can be found over lower than normal velocities in model V3.I of Morelli and Dziewonski [1986; 1991].

The velocity anomalies extend significantly from the CMB into the middle mantle and perhaps reach the 650 km discontinuity; thus, they are continuous across the so-called D'' region. Romanowicz and Gung, [2002] propose that the effects of superplumes extend into the upper mantle; they point out toward good correlation of degree-2 velocity anomalies near the CMB and Q anomalies in the transition zone, with perhaps, only the thermal coupling across the 650 km discontinuity.

Recently, the Princeton group [Montelli *et al.*, 2003b; Montelli *et al.*, 2003a] derived a P -velocity model in which some of the hotspots continue from surface well into the lower mantle and perhaps all the way to the CMB. It is difficult to assess the validity of this claim before publication, but a degree of skepticism is justified in view of discussion in the introductory part of this extended abstract and the fact that the authors seem not to recover well the dominant large-wavelength of the anomaly field in the lowermost mantle.

There are many other smaller-scale (regional) anomalies near the CMB. Among them are anisotropy; smaller scale, rather abrupt changes in the sign of the anomalies; 10-30 km thin ultra-low velocity zones at the base of the mantle [Garnero *et al.*, 1993], which seem to be preferentially located within the large-scale low velocity anomalies. Their relation to the surface tectonics is unclear.

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THE PROTEROZOIC BUSHVELD COMPLEX, SOUTH AFRICA: PLUME, ASTROBLEME OR BOTH?

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The Bushveld Complex has been cited (1, 2,) as a possible example of a large (diameter 400 km, volume 10^6 km^3) igneous province generated by rapid decompression melting at the leading edge of a mantle plume, triggered by the impact of a large ($d \geq 20 \text{ km}$, $v \geq 10 \text{ km/sec}$) iron bolide. This scenario would reconcile the widely accepted Bushveld plume model (3, 4) with the controversial proposal for an initial Bushveld impact (5, 6, 7). The evidence for an initial catastrophe comes from a group of extraordinary high-energy high-temperature debris flows at the base of the oldest Bushveld unit, the Rooiberg Group and from intense deformation bracketed between the end of pre-Bushveld shallow marine sedimentation (Transvaal Supergroup) and the coming-to-rest of the basal Rooiberg debris flows. Subsequent plume-related events are documented in the remaining 90+% of the Rooiberg Group and voluminous associated mafic and granitic units, and in a long period of structural instability that followed the initial event. A second catastrophe, documented by a zone of megabreccia blocks (diameter to 50 m) in the upper part of the Rooiberg Group, has been traced around the entire circumference of the Bushveld Complex (6, 8).

The Bushveld Complex is a unique association of (i) rocks of volcanic aspect (Rooiberg Group, diverse but predominantly siliceous, $0.1 \times 10^6 \text{ km}^3$) (9); (ii) anorthositic, mafic and ultramafic cumulates (Rustenburg Layered Suite, RLS, $\sim 1.0 \times 10^6 \text{ km}^3$) (10, 11); and (iii) A-type granites (Lebowa Granite Suite, LGS, $0.1 \times 10^6 \text{ km}^3$) (12). Granophyres, developed as facies of these rocks or at contacts between them, constitute the Rhashoop Granophyre Suite (13). Each of these has many subunits. RLS and LGS are the largest of their kind in the geological record. The principal units are sheet-like surface accumulations or sills (max. thicknesses: Rooiberg, 4.5 km; RLS, 9 km; LGS, 5 km. Max. cumulative thickness in one locality: $\sim 12 \text{ km}$) (10). Zircon U-Pb dates for RLS and the upper part of the Rooiberg Group have converged at $\sim 2,061 \text{ Ma}$. LGS activity may have continued for another 7 m.y. (14).

Discontinuous RLS outcrops outline the peripheries of three lobes or overlapping basins: eastern, western, and northern. They are truncated by erosion and their original extent is unknown. Contact metamorphism of underlying sedimentary units locally extends for tens of kilometers beyond the outcrop belt (15). The interiors of the basins are obscured by LGS granite or covered by younger sedimentary rocks, except for two 50-km inliers of pre-Bushveld rocks, respectively within the eastern and western basins (16). A large probable RLS outlier (Molopo Farms Complex; 17) is known from the subsurface of Botswana, 200 km to the west, and a small one (Losberg body; 18) crops out 100 km to the south. The total RLS volume of $\geq 1.0 \times 10^6 \text{ km}^3$ suggested by petrological calculations seems reasonable (11).

The Bushveld Complex is roofless (19), not intrusive in the usual sense. The later members intrude earlier ones but the entire complex cooled on the surface. The Rooiberg Group erupted in a previously stable tectonic setting but rests on a regional unconformity (20), probably enhanced by catastrophic scouring. This lower contact became the principal conduit for RLS and LGS sills. Consequently, the Rooiberg Group now forms the roof of the Bushveld Complex, of which it is itself a part. Upper units of the Rooiberg Group may have been synchronous with the later pulses of RLS and early LGS (3, 21). If so, siliceous flows piled up on top of the Rooiberg stack while sills invaded its bottom. With few exceptions, invasions of RLS and LGS have destroyed the critical basal debris flows by contact metamorphism and rheomorphic melting (22). Exceptions include the intrabasin inliers and a slice of the lowest unit of the Rooiberg Group (Dullstroom Formation) locally caught beneath the RLS in the southeastern Bushveld. The recognizable Dullstroom Formation has a limited distribution in the southeastern part of the Bushveld Complex (8). The basal debris flows are best preserved in scoured (?) paleochannels.

What kept the 12-km surface pile from collapsing? Rooiberg flows and RLS sills must have accumulated in a pre-existing basin many kilometers deep, even allowing for Rooiberg overflow. There could have been no significant subsidence until after the invasion of the third major RLS pulse, the Main Zone. Paleomagnetic data (23) showed the RLS as horizontal until that time. Gravity data by Cousins (23) indicated an annular RLS basin fed by ring intrusions, not the central vent for the gigantic lopolith (25) that still haunts our textbooks. Hamilton (5) and Rhodes (6) invoked quasi-simultaneous multiple impacts to create a three-lobed Bushveld crater and the smaller Vredefort dome, 250 km to the SSW, both surrounded by multiple rings. In Rhodes' view, Cousin's annular basin was the inner ring syncline. I interpret Cousin's ring dike as an impact-related deep fracture system (as seen at Chicxulub; 26), originally inward-dipping but now rotated to near-vertical by basin subsidence. It was partly filled with RLS mafic melts, released from a rising mantle plume by decompression and contaminated with crustal material (27).

The present interpretation evolved from Rhodes', who regarded the two intrabasin inliers as central uplifts. From available maps (16) and a seismic profile (28), I reinterpret them as wall segments of a transient impact cavity. Each inlier consists of two fragments, in fault contact: deformed and undeformed. Deformed fragments are interpreted as either parts of the transient cavity wall or of a collapsed central uplift, ramped against a cavity wall (29). Rocks of the Transvaal Supergroup are tightly folded and metamorphosed, up to pyroxene hornfels facies (16). On a microscopic scale, shearing is intense. Compressional deformation on this scale is unknown in Transvaal rocks elsewhere. Neither is it known from any volcanic terrain; extension is the rule for large calderas (30). The undeformed fragments are interpreted as gigantic Transvaal blocks that slid into the unstable transient cavity, meeting the emerging debris flows. The present study has concentrated on the eastern (Stavoren) fragment. There, unmetamorphosed Transvaal rocks (Makeckaan Subgroup) have low ($<10^\circ$) dips and brecciation is confined to fault zones. Only a few tens of meters of overlying debris flows are preserved. Elsewhere, large gravity slides have been related to the western Bushveld basin in Botswana (31).

Rhodes regarded the entire Rooiberg Group as impactite crater fill. It is here interpreted as outflow; there are as yet no documented outcrops of crater fill or signs of an eruptive source. In the present interpretation only the basal debris flows are impact-related; subsequent Rooiberg,

RLS and LGS rocks are interpreted as plume-related. Three facies of the basal Rooiberg debris flows were studied: Proximal, in the undeformed Stavoren fragment, intermediate in the annular basin, and distal in the Dullstroom slice beneath the RLS. Superficially, the proximal facies resembles spherulitic rhyolite lava studded with relict quartzite blocks, some exceeding 10 m, in every stage of shearing, comminution, recrystallization, melting, and obliteration. In a paleochannel, the base rests on a polished and grooved surface, resembling a glacial pavement. Along the fault that juxtaposes the undeformed Stavoren fragment against the neighboring deformed (Marble Hall) fragment, the Stavoren fragment broke up. Overturned slabs of Makeekaan quartzite, tens to hundreds of meters long, were engulfed in a debris flow of hot quartz sand, now recrystallized (formerly identified as felsite).

Proximal-facies debris flows, derived from quartz arenite, have sedimentary mineralogy and chemistry (up to 90% SiO₂) (6, 9). They contain 191-304 ppm Zr (9) but no zircon (32). Apparently the temperature was above the stability range of zircon. It was high enough to partially melt quartzites with varying amounts of interstitial sericite. Quartz inverted to high-temperature polymorphs of SiO₂, probably ordered and disordered forms of high tridymite (33), which became the dominant phases. This was recognized 70 years ago (34) but the entire sequence of stages has not been previously described: (i) Solid-state inversion of quartz into a network of needles with optical continuity, confined within single quartz grains. (ii) Solid-state inversion of relict quartz grains into stout lath-shaped crystals ("stubbies"). (iii) Partial melting, leaving pseudospherulites (rounded mm-sized remnant aggregates of stubbies and relict quartz) in a matrix of melt. (iv) Quenching of melt, with rapid growth of mm-sized needles, some swallow-tailed, in a glass matrix. (v) Inversion of all previous forms into paramorphous quartz. Stage (i) also transformed >10 m of underlying quartzite and subgraywacke. Such sanidinite-facies contact metamorphism is unknown below volcanic rocks but occurs in cm-sized quartzite xenoliths in gabbro (35). The closest analog is tridymite formed in silica-brick linings of high-temperature industrial furnaces between 1,200 and 1,370°C (36). At Sudbury, in a 60-m quartzite breccia at the base of the Onaping impactite, tridymite of stage (iv) formed in interstitial melt between quartz grains (37).

The intermediate facies is barely preserved in the type section of the Rooiberg Group, at Loskop Dam (38). Most of it was obliterated at the RLS contact. The rock is black quartzite with a matrix of glass and devitrified glass. Coexistence of delicate uncompressed glass shards and lithic clasts suggests inflated ignimbrite-like transport. This exceedingly complex rock requires more study. In the distal facies, 300 m of melt rock and several flows of black quartzite are preserved in three paleochannels at the base of the type-Dullstroom section (9). From a distance, the black quartzite could be taken for ignimbrite with lithic clasts, black fiamme, and basal sand-wave and planar surge deposits. However, it contains no pumice, shards, glass, devitrification features, or phenocrysts. Rounded quartz sand grains are supported by a dark matrix of fine amphibole, chlorite, plagioclase and quartz. Chemically, the rock is a mixture of quartz and a mafic component. The lithic clasts are hornfels and the "fiamme" are cm-size zoned lenses with cores of pyrite, surrounded by zones of carbonate-epidote and actinolite and a reaction rim of fine quartz. These clasts were evidently metamorphosed to amphibolite in situ. The rock could also be taken for a quartzose suevite variant, in which amphibolite lenses take the place of glass lenses. In one locality (Kwaggaskop), m-size brecciated quartzite blocks are plastered against the wall of the paleochannel. Ignimbrite-like inflated transport of a hot debris flow is indicated. In

another locality (Messchunfontein) accumulations of lithic clasts up to 50 cm resemble co-ignimbrite lag deposits. Dark matrix material, concentrated in a finely laminated interbed, resembles a co-ignimbrite ash-cloud deposit. Chemically similar material (low-Ti basaltic andesite) forms flows with quench textures. Above the debris flows, a rhyolite-like lava flow with m-size relict boulders and contorted flow bands has the same amphibolite lenses as the debris flows and similar chemical composition. It is the same material, hot enough to have partially melted. Some samples have microscopic quartz-tridymite paramorphs (9).

Aside from the basal debris flows, the Rooiberg Group is here interpreted as outflow from one or more melt pools created by impact but rapidly replenished by mafic melts released from a decompressing mantle plume. This can explain the abundance of diverse mafic lavas in the ~1,200 m Dullstroom Formation. In time, the pools coalesced and increasingly homogeneous siliceous melts developed by differentiation and crustal assimilation (3, 4). They increasingly resembled conventional volcanic rocks. Quench textures and scarcity of phenocrysts testify to continuing superliquidus temperatures. Repeated influxes of water triggered explosive eruptions of ignimbrite-like flows, interspersed with high-energy influxes of cold sediment, with exotic clasts from distant sources, and hot outflows (39). Concurrently, mafic plume-related RLS sills invaded the base of the Rooiberg Group, followed by LGS crustal melts. Return to stability allowed segregation of RLS cumulate zones traceable over hundreds of km, even as crustal diapirs locally rose into the RLS (40). During the second catastrophe (collapse of the present basins during LGS invasion?), an inflated flow carried both 10+-m quartzite blocks and delicate mm-size glass spherules. Tilting during basin subsidence turned horizontal RLS sills in Cousin's annular ring (24) into the inward-dipping sheets of a more recent gravity model (41). As the lithosphere readjusted, faulting and minor siliceous magmatism continued into post-Bushveld time (15, 42).

Objections to the proposed initial catastrophe have been raised by authors (4, 43) who dismissed the basal Rooiberg debris flows as conventional rhyolite and sandstone. The present combined impact-plume model reduces the volume of impactite, previously considered excessive. Other authors have cited the absence of shock criteria (planar deformation features, high-density polymorphs and glass, shatter cones, pseudotachylite). The rocks discussed here were either recrystallized or melted and were far from the proposed impact point. Inversion of quartz to tridymite occurs only under low-P, high-T conditions (36), outside the high-P, high-T stability field of coesite and stishovite. Indications of moderate shock (46) in contact metamorphosed rocks and in clasts of debris flows include mosaicism, deformation lamellae and twins, and cataclasis.

The Vredefort dome, which has all of the diagnostic impact criteria, is bracketed by the same stratigraphic units as the Bushveld Complex. To date, no unequivocal field evidence has proved a difference in ages. If the difference in zircon U-Pb dates between Bushveld (~2,060 Ma) and Vredefort (~2,020 Ma; 47, 48) could be resolved, a multiple Bushveld-Vredefort impact event would be plausible. Closure dates from quenched rocks in the upper part of the Rooiberg Group are close to times of emplacement. The thermal history of exhumed Vredefort rocks is more complex and controversial (49, 50). Gibson and Stevens (49) interpreted two stages of metamorphism by impact into a site preheated by a Bushveld-age plume, so "that parts of the granulite-facies terrane were still at temperatures above the granite solidus at the time of impact."

Could this sequence be reinterpreted as impact-plume-exhumation? Roberts and Finger (51) concluded that, in the presence of melt, zircon U-Pb dates of granulites "rather than representing the age of high-pressure metamorphism, ...most likely date a stage where the rocks had already been exhumed to medium-pressure levels." For the Paleozoic granulites cited by (51), the time lag was 30 m.y. A Vredefort 40-m.y. impact-plume-exhumation lag would also have affected $^{39}\text{Ar}/^{40}\text{Ar}$ dates (52). It may be significant that Archean granulite from a locality near the north side of the Bushveld Complex yielded a "Vredefort" zircon U-Pb date of $2,027 \pm 6$ Ma (53).

The Bushveld Complex meets most of the criteria for an impact-triggered plume, as set by the hydrodynamic model of Jones et al. (1): Melt volume $\sim 10^6$ km³, crater auto-obliterated by melts, high rate of eruption (11), no initial doming, "plume-like geochemical signature," and no "deep geophysical fingerprint." It differs from the prediction that "melt extrusion would start with...low-viscosity peridotitic melts" but the mafic components of the Dullstroom Formation may qualify.

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Tertiary Eifel volcanism – intraplate mantle plume or extension-related activity?

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The Eifel volcanic fields are located in the Hercynian Rhenish Massif, Western Germany, and belong to the Cenozoic Central European Volcanic Province (CCEVP). About 300 Tertiary volcanic relics occur in the so-called Hocheifel (e.g., Huckenholz, 1983) a region of about 1000 km² in between the neighboring Quaternary West and East Eifel volcanic fields.

Seismic tomography has shown that there is a low velocity anomaly in the upper mantle underneath the Eifel (Ritter et al., 2001). This anomaly extends at least to a depth of 400 km and can be explained by a mantle plume with an excess temperature of 150-200±100 K. Whereas the activity of the Quaternary East and West Eifel volcanism appears to be related to this mantle plume, the geodynamic setting of the Tertiary Hocheifel volcanism is not clear. As alternatives, precursor activity of the intraplate Quaternary Eifel plume or magma generation related to extension are plausible models.

⁴⁰Ar/³⁹Ar step-heating measurements by laser fusion on amphibole, feldspar and groundmass separates from 19 prominent Hocheifel volcanic occurrences yield an age range from 44 to 34 Ma (Middle and Upper Eocene). Previously published conventional K-Ar Oligocene and Miocene ages cannot be confirmed. For the Tertiary Hocheifel activity, two time intervals can be distinguished each related to a discrete geographical setting. An older phase of volcanism occurs along a central ca. 50 km long north-striking lineament with a south to north age progression from 44 to 40 Ma. A younger phase comprising an age range from 38 to 34 Ma developed ca. 10 km to the west and to the east of the central north-striking lineament.

The two age phases of the Tertiary Hocheifel volcanism form two discrete data fields in a diagram ⁸⁷Sr/⁸⁶Sr_i vs. ¹⁴³Nd/¹⁴⁴Nd_i, where i stands for initial. In addition, these phases are also distinct from the data groups defined by the Quaternary Eifel, as well as by further Tertiary and Quaternary volcanic fields of the CCEVP. Initial Sr and Nd isotope compositions as well as trace element patterns show that the Tertiary Eifel mantle is inhomogeneous at least in the order of magnitude of 10 km and that the mantle sources tapped by Quaternary and Tertiary Eifel volcanism are different.

Based on the local geographical distribution of the Tertiary Hocheifel volcanic occurrences and their regional geographical setting as north-west prolongation of the Upper Rhine Graben, it appears that the Tertiary volcanic activity is rather related to Eocene Rhine Graben rifting than to precursor activity of the Quaternary Eifel plume. In order to test whether there is a relationship between the Eocene Eifel and the Upper Rhine Graben volcanism we are pursuing a detailed geochemical study to compare the tectonic regimes as well as the volcanic rocks from both regions in terms of ages, mantle sources and magma generation processes. However, at the moment we cannot exclude the hypothesis that the Tertiary Eifel volcanism represents precursor activity of the Quaternary Eifel plume.

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Definition of a Cenozoic alkaline magmatic province in the southwest Pacific without rift or plume origin

Finn, Carol A., Müller, R.Dietmar, and Panter, Kurt

A Cenozoic (<50 Ma) bimodal, but largely basaltic, mostly alkaline igneous province covers a broad area of continental and oceanic lithosphere in the southwest Pacific (Fig. 1) has been conjecturally linked to rifting, mantle plumes, or hundreds of hot spots, but all of these associations have flaws. For example, plate reconstructions demonstrate that the last episode of major regional rifting in west Antarctica, eastern Australia and New Zealand occurred during the Mesozoic break-up of Gondwana. GPS and stress-field measurements show no extension in Australia, New Zealand and much of west Antarctica, suggesting that the widespread magmatism cannot be explained by rifting alone. Estimates of volumes of magmas erupted in west Antarctica and Australia, as well as magma production rates are low compared to areas associated with plumes. Uplift and doming typically associated with mantle plumes are also largely absent. Also, to explain the areal distribution of the volcanism, an unusually large plume would have to underlie the entire southwest Pacific, or there would have to be hundreds of hot spots, which are not observed. Clearly, new models for volcanism are required.

Comparison of the location of volcanoes and seismic shear wave perturbation models (Fig. 1), show that this alkaline volcanic province occurs in thin (<80 km) lithosphere (e.g. Fig. 2). Offshore areas of old (> 100 Ma), thick lithosphere like the Weddell Sea (Fig. 2) do not contain Cenozoic alkali basalts. The province correlates with distinct low seismic velocity anomalies (Fig. 1) restricted to a zone in the mantle between ~60 and 200 km depth (e.g. Fig. 2).

Geochemical studies show that for most of the region, the magmatism is a result of small degrees of melting ($F = 1-3\%$) of a source enriched in incompatible elements relative to primitive upper mantle. The enrichment may have involved the introduction of volatile-rich fluids or melts into pre-existing upper mantle. This suggests that melting of metasomatized upper mantle can occur without excessive temperatures (Fig. 3) and that the low seismic velocities are primarily related to slightly elevated temperatures, water and, in places, melt.

The key to generating long-lived, low volume alkaline magmatism is the combination of thin (<80 km) lithosphere underlain by metasomatized, mostly Pacific mantle at only slightly elevated temperatures. The age of the metasomatism is not known but may be related to a combination of Paleozoic-Mesozoic subduction along the Pacific margin of Gondwana and possible plume-related activity in the Jurassic. During Cretaceous break-up of Gondwana, rifting in east Australia and west Antarctica did not result in voluminous magmatism despite thinning and regional extension of continental lithosphere containing metasomatized mantle. This suggests that a regional heating and/or mantle upwelling event is required to allow alkaline magmatism.

Plate motion and seismic tomography studies propose that high density and velocity subducted slabs lying in the lower mantle detached from the mantle transition zone in the Eocene (e.g. Fukao et al., 2001). Geodynamic model of the effects of an “avalanche” of detached slabs suggest that it is possible to generate vertical and lateral flow and high temperatures in the upper mantle. Although we do not call for a dramatic increase in temperature or mantle flow near the detached slabs, such a mechanism may induce slight heating, catalyzing melting of metasomatized mantle and eruption during extension.

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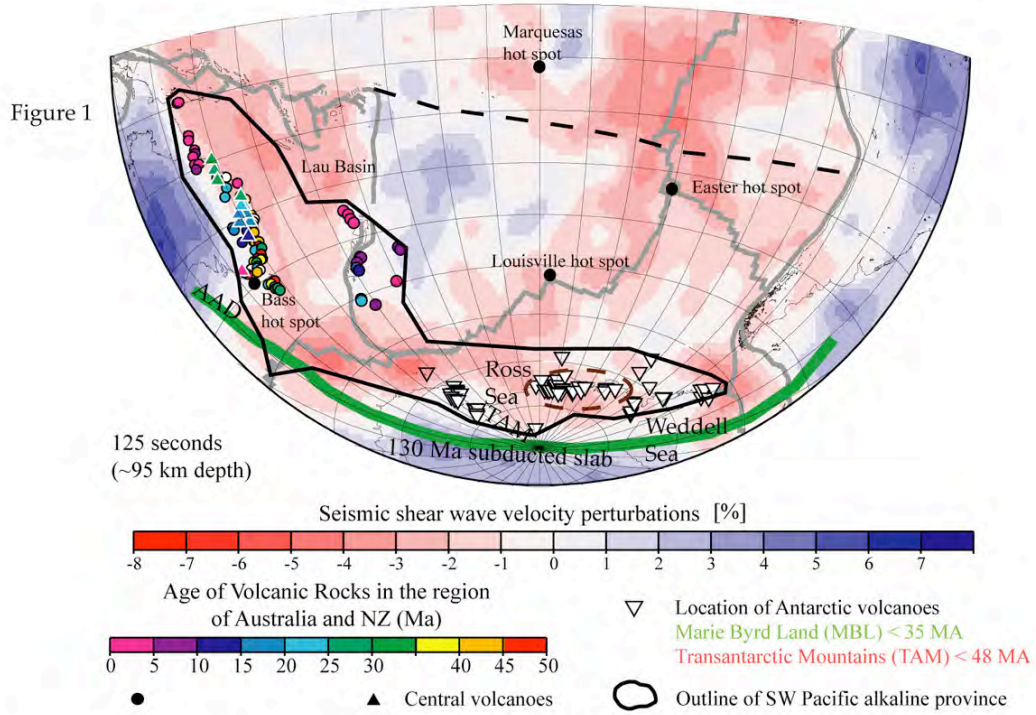


Figure 1. Rayleigh wave 125s group velocity map (~95 km depth) of the SW Pacific (Larson and Ekström, 2001). Only robust, hot spots with long traces (Clouard and Bonneville, 2001) are shown.

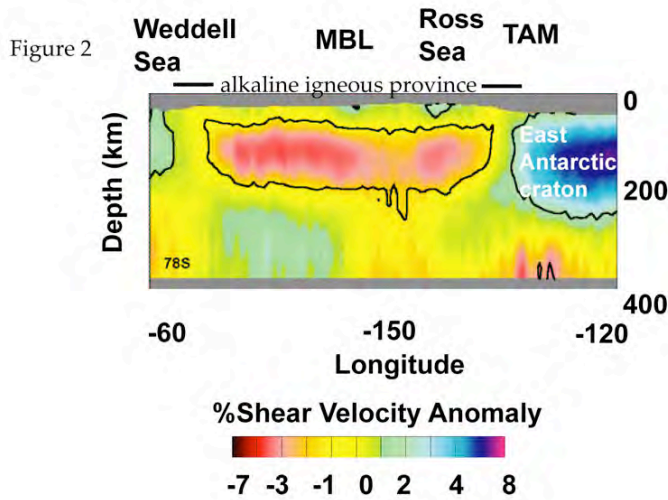


Figure 2. Shear wave velocity perturbation model derived from inversion of Rayleigh and Love wave data and relative to AK135, Antarctica (from Ritzwoller et al., 2001).

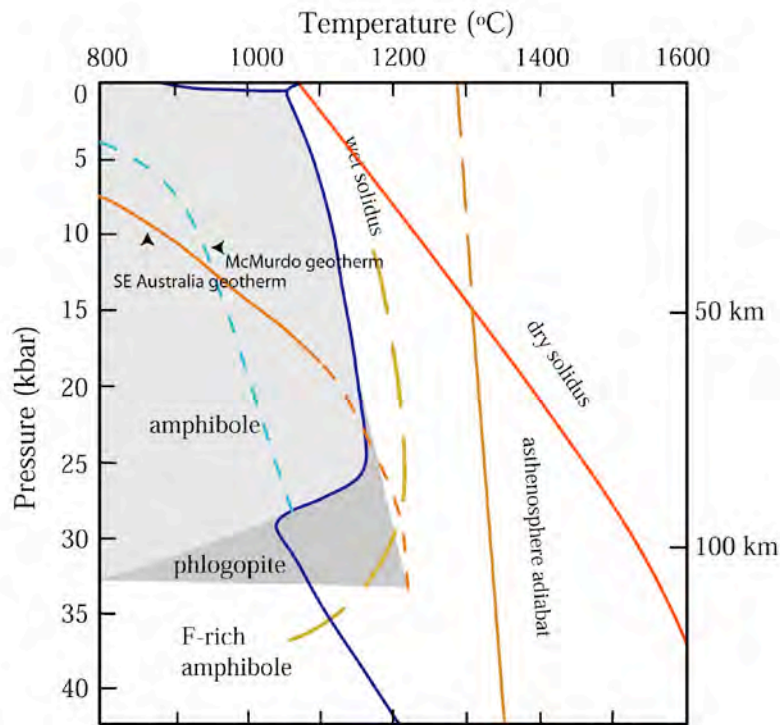


Figure 3. McMurdo/SE Australia geotherm (from Berg et al., 1989; O'Reilly and Griffin, 1985) with 2 possible extrapolations (solid and dashed lines), compared to stability fields of amphibole and phlogopite. Also shown are water-saturated and water-undersaturated solidi and an adiabatic path for asthenospheric mantle.

A plume origin for the Ontong Java Plateau?

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The submarine Ontong Java Plateau (OJP) is the most voluminous of the Earth's large igneous provinces (Coffin & Eldholm 1994). The plateau, defined mostly by the 4000-m bathymetric contour (Fig. 1), covers an area of 2×10^6 km³ (comparable in size with Western Europe), but OJP-related volcanism extends over a considerably larger area into the adjacent Nauru and East Mariana basins. Collision with the Solomon arc has resulted in folding and uplift of the southern margin of the OJP in the last 6 m.y., exposing thick (up to ~3.5 km) sections of basaltic rocks on land in the Solomon Islands, notably in Malaita, Santa Isabel, and San Cristobal (e.g. Petterson 1999). OJP basaltic basement rocks have also been sampled at ten Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) drill sites on and around the plateau (Fig. 1).

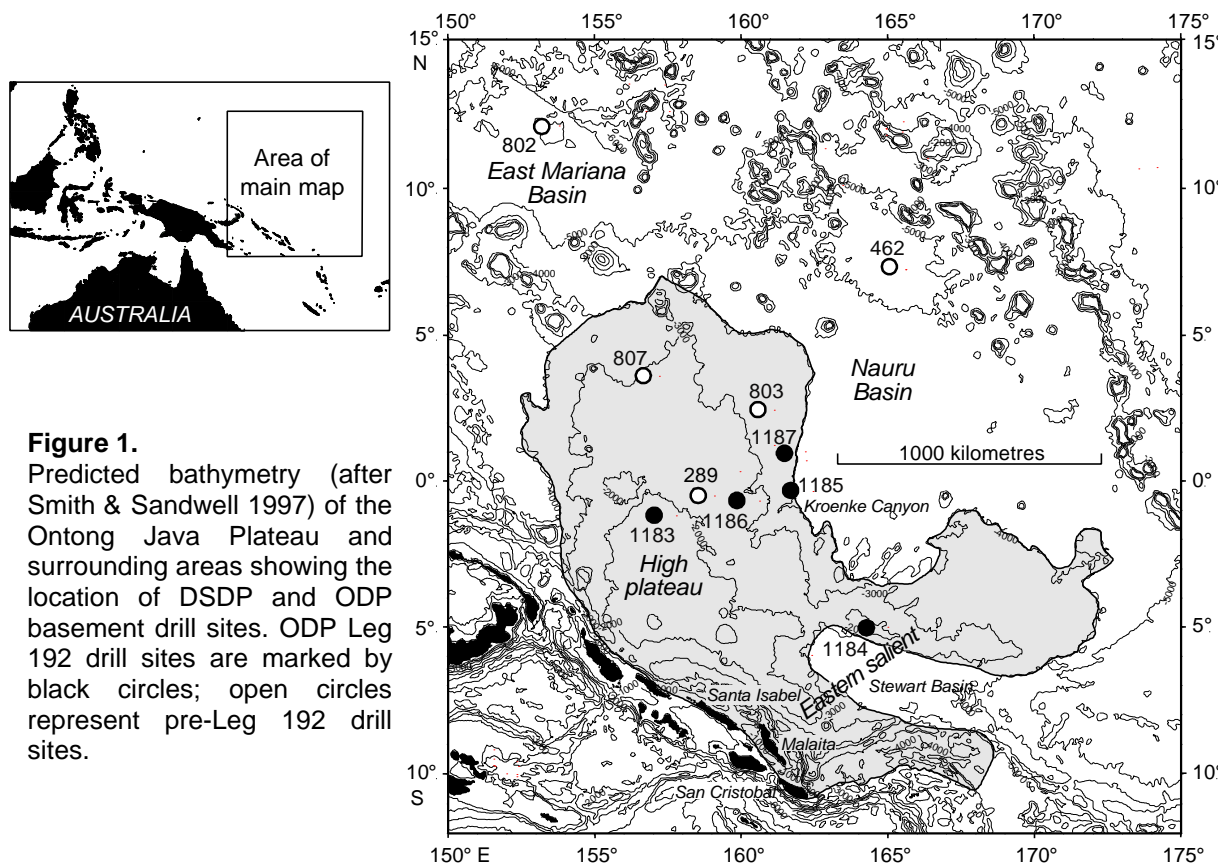


Figure 1. Predicted bathymetry (after Smith & Sandwell 1997) of the Ontong Java Plateau and surrounding areas showing the location of DSDP and ODP basement drill sites. ODP Leg 192 drill sites are marked by black circles; open circles represent pre-Leg 192 drill sites.

With an average thickness of crust beneath the plateau of 30-35 km (Gladchenko et al. 1997; Richardson et al. 2000), the volume of igneous rock forming the plateau may be as high as 6×10^7 km³. The OJP seems to have been emplaced rapidly at around 120 Ma (e.g. Mahoney et al., 1993; Tejada et al., 1996, 2002; Chambers et

al. 2002; Parkinson et al. 2002) and the peak magma production rate may have exceeded that on the entire global mid-ocean ridge system at the time (e.g. Coffin & Eldholm, 1994).

Despite its size, the OJP's basaltic crust appears to be remarkably homogeneous in composition (Fitton & Godard, in press; Tejada et al., in press). The most abundant rock type is a uniform low-K tholeiite, represented by the Kwaimbaita Formation on Malaita and found at all but one of the DSDP and ODP drill sites on the plateau and in the adjacent basins. Kwaimbaita-type basalt is capped by a thin and impersistent veneer of a slightly more incompatible-element-rich tholeiite (the Singgalo Formation on Malaita, some flows in Santa Isabel, and the upper unit of flows at ODP Site 807). Singgalo-type basalt has lower $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$, and higher $^{87}\text{Sr}/^{86}\text{Sr}$ than does Kwaimbaita-type basalt.

A third magma type is represented by high-Mg (Kroenke-type) basalt found in thick (>100 m) successions of lava flows at two drill sites (ODP Sites 1185 and 1187) 146 km apart on the eastern flank of the plateau (Fig. 1). The high-Mg basalt is isotopically indistinguishable from Kwaimbaita-type basalt and may therefore represent the parental magma for the bulk of the OJP. Low-pressure fractional crystallisation of olivine followed by olivine+augite+plagioclase can explain the compositional range from high-Mg Kroenke-type to Kwaimbaita-type basalt. The Singgalo-type basalt probably represents slightly smaller-degree, late-stage melting of an isotopically distinct component in the mantle source.

Identification of a parental magma type allows the primary-magma composition to be estimated by incremental addition of equilibrium olivine until the residual mantle olivine composition is reached. Taking residual olivine compositions ranging from Fo_{90} to Fo_{92} gives primary magma with MgO ranging from 15.6 to 20.4 wt.%, respectively. Incompatible-element contents in the calculated primary magma, coupled with radiogenic isotope ratios, are consistent with a mantle source consisting of primitive mantle depleted through the extraction of 1% by mass of average continental crust (Fitton & Godard, in press; Tejada et al., in press). The degree of melting required to produce the primary magma from this source ranges from 27% (in equilibrium with Fo_{90}) to 31% (Fo_{92}). Independent estimates of residual olivine composition and degree of melting can be obtained from the major-element composition of primitive basalt through the combined forward- and inverse-modelling method of Herzberg & O'Hara (2002). Applying this method to Kroenke-type basalt and a fertile mantle source gives 25% melt with residual $\text{Fo}_{90.4}$ for perfect fractional melting, and 30% melting with $\text{Fo}_{91.6}$ for equilibrium melting. The remarkable agreement in the results of trace- and major-element modelling provides compelling evidence that the OJP had a fertile peridotite mantle source.

Peridotite mantle with a potential temperature >1500°C will melt to a *maximum* of around 30% if decompressed to shallow levels (i.e. at or close to a spreading centre). To achieve an *average* of 30% melting requires that the mantle is actively and rapidly fed into the melt zone, and a start-up mantle plume provides the most obvious mechanism. This should have caused uplift well above sea level, but the abundance of essentially non-vesicular pillow lava and the absence of any basalt showing signs of subaerial weathering show that the OJP was emplaced below sea level (e.g. Neal et al., 1997; Mahoney et al., 2001). Volatile concentrations in quenched pillow-rim glasses suggest eruption depths ranging from 1100 m at Site 1183 to 2570 m at Site 1187 (Roberge et al., in press).

We have not yet been able to resolve the paradox of apparent high mantle potential temperature coupled with submarine emplacement. An eclogitic source

does not provide a solution because the high-Mg parental magma would require almost total melting, and consequently a very high potential temperature would be needed to provide the latent heat of fusion. We can also rule out a hydrous mantle source because the magmas have very low H₂O contents (Michael, 1999; Roberge et al., 2002, in press). Widespread melting of the mantle following the impact of an asteroid provides an attractive means of avoiding uplift, but the magma would be generated entirely within the upper mantle and would have the chemical and isotopic characteristics of NMORB. OJP basalt is isotopically (Tejada et al., in press) and chemically (Fitton & Godard, in press) distinct from NMORB and seems to have a lower mantle source.

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Collision-induced mantle flow during Tethyan closure: a link between magmatism, lithosphere ‘escape’, and arc-trench rollback?

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The transition from pre- to post-collision settings is marked by sequences of calcalkaline, potassic, and basaltic magmatism each representing thermal and compositional probes of the convecting upper mantle. This pattern characterizes recent collisions between migrating fragments of Gondwana (Africa, Arabia, India, and Australia) and accreting Eurasia, and match lithospheric collision responses that include crustal shortening and thickening, post-orogenic collapse, and continental escape tectonics. Syn- and post-collision potassic magmas include shoshonites, lamproites, and (rarely) kamafugites which typically coincide with the inception of continent-scale shearing (as recorded by Ar-Ar thermochronology). Tomographic and S-wave splitting studies also suggest potassic magmatism corresponds with post-collision slab detachment and incipient continental crust subduction. Post-collision basaltic phases typically comprise tholeiites, alkali basalts, basanites, and nephelinites and are confined largely to pull-apart basins and extinct offshore spreading axes. Usually post-dating major shearing events they correspond with continuing shifts in transtensional stress fields representing late stage adjustments to ‘escape’-related block rotations.

Complex isotopic and trace element enrichment-depletion histories characterize magmatic sources of both types, indicating contributions, via both subduction and lithospheric mantle delamination, of sialic components and hydrated, refractory peridotite. At a regional scale, Tethyan magmatism taps compositionally distinct mantle domains whose boundaries with each other and with contiguous Atlantic and Pacific N-MORB domains have migrated with time in accord with global plate motions. For example, as northward-moving Africa approached Eurasia in the early Mesozoic, HIMU-rich mantle infiltrated beneath Neo-Tethyan marginal basins, accreting western Eurasia, and eastern parts of the Atlantic Ocean. At about the same time, Indian Ocean spreading centers began tapping EM1-contaminated ‘DUPAL’ mantle, the latter infiltrating beneath remnant Neo-Tethyan basins - preserved in central and eastern Eurasian ophiolites - and eastward-propagating western and southwestern Pacific back-arc basins.

The HIMU-rich western domain has been ascribed to north-northeast-directed channeling of a high- $^{238}\text{U}/^{206}\text{Pb}$ Central Atlantic plume while eastern Tethyan DUPAL mantle was also attributed to deep plumes – the latter invoked as the principal cause of Gondwana breakup. Mantle plume models are problematic for several reasons, however. Lower mantle low velocity anomalies inferred beneath Indian and Atlantic Ocean ‘hot spot’ loci are probably suspect, given the relatively poor resolution of deep mantle tomography. Secondly, if, as is commonly assumed, plumes are stable and persistent features, the spatial-temporal distributions of Tethyan volcanic centers do not match those expected from recorded lithospheric plate motions. Finally, mantle flow fields inferred on the basis of geochemical tracers and S-wave splitting

studies are at variance with the domain boundary migrations implied by Atlantic and Indian Ocean plume models. While low- $^{206}\text{Pb}/^{208}\text{Pb}$ DUPAL contaminants have been interpreted to represent lithospheric detritus accumulated at the core-mantle interface, they may be more plausibly explained by selective delamination of dispersed Gondwana cratonic roots in a context of shallow-level mantle convection. This type of model is supported by new thermobarometric and isotopic data for continental mantle xenoliths, suggesting that substantial thinning of Gondwana cratons has occurred since the early Paleozoic.

Most current or recent Tethyan volcanism corresponds with shallow mantle (< c. 300 km) P- and S-wave velocity minima which indicate ‘swell-’ rather than plume-like upper mantle thermal anomalies. Potential temperatures inferred from melt thermobarometric data are mostly $\leq 200^\circ\text{C}$ above those expected beneath undeformed continental lithosphere. Accordingly, the observed scale of mantle melting requires both rapid asthenosphere decompression and lithosphere stretching. Significantly, the spatial extent of shallow tomographic anomalies may define laterally-continuous flow channels linking regions of collision-displaced asthenosphere (e.g. beneath the Himalayas and Tibet) with those underlying: 1) extensional volcanic centers (e.g. eastern China and Indochina), 2) retreating intra-oceanic forearcs (e.g. the Izu-Bonin-Mariana terrain), and 3) their conjugate marginal basins (e.g. the Parece Vela Basin and Mariana Trough).

While Tethyan continental basalts tap variably (if slightly) contaminated fertile asthenosphere, the potassic magmas tap strongly contaminated sources in mantle ‘wedge’ regions. Geochemical studies indicate two major types of contamination accompany the transition from plate convergence to collision: 1) refractory serpentized peridotite, delaminated from the overriding plate, following its rheologic ‘conversion’ and incorporation by asthenospheric corner-flow, and 2) continental crust, either delaminated from thickened orogenic crust or introduced by subduction as slivers or metasomatic melts. Hitherto subduction-driven flow will be substantially modified if post-collision slab detachments allow fresh asthenospheric influx from below.

In summary, mantle wedge flow paths represent a critical aspect of collision-related mantle dynamics the evidence for widespread ‘non-plume’ magmatism, Neo-Tethyan marginal basin propagation, and concomitant infiltration of thermally and compositionally anomalous asthenosphere offers a compelling case for continent-scale, collision-induced mantle extrusion. This model is in accord with upper mantle thermal structures and uniquely explains the observed kinematic and temporal relationships between plate collisions, intra-plate magmatism, and distal arc-trench rollback processes. Mantle extrusion also provides a plausible driving mechanism for lithosphere escape tectonics. Petrologic interpretations of mantle thermal and compositional will be developed as a template for understanding collision-related mantle flow, amenable for testing by geophysical methods, tectonic facies analysis, and numerical simulation.

An alternative model for Iceland & the North Atlantic Igneous Province

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Summary

It is almost universally assumed that Iceland is underlain by a hot plume rising from deep within the mantle. At Iceland, probably the best-studied hotspot on Earth, this hypothesis is inconsistent with many first-order observations, such as the lack of evidence for high mantle temperatures, a time-progressive volcanic track or a seismic anomaly in the lower mantle. Iceland is essentially a melt anomaly with normal, ridge-like mantle temperatures.

Iceland lies where the mid-Atlantic ridge crosses the Caledonian suture, which marks the site of a ~ 400 Myr-old subduction zone. The great melt production at Iceland is explained well by enhanced fertility there resulting from ancient subducted slabs that still remain in the shallow mantle. This model is consistent with the historical locus of melt production, and the lack of geophysical indicators of a plume [Foulger *et al.*, 2003a]. It is also consistent with the geochemistry of Icelandic basalts, which differs only subtly from mid-ocean ridge basalt (MORB) [Foulger *et al.*, 2003b]. In this way, Iceland is explained as a natural consequence of relatively shallow processes related to plate tectonics. Unlike the hot, thermal plume model, this model can account for all the major geophysical, geological, petrological and geochemical observations at Iceland without special pleading, paradoxes, or invoking coincidences.

A plume model is inconsistent with many observations

Virtually none of the results of Earth science research in Iceland agree with the predictions of the plume hypothesis [Foulger, 2002]. Evidence for the 200-600°C mantle temperature anomaly required for a plume is absent. Marine heat flow measurements around Iceland provide no evidence for high temperatures [Stein & Stein, 2003]. The petrology of Icelandic basalts is similar to MORB, and geothermometers involving Mg# and MgO indicate temperatures no more than approximately 70°C higher than the average beneath mid-ocean ridges (e.g., http://www.mantleplumes.org/HawaiiFocusGroup/Sisson_abs.html). Even in central Iceland, primitive lavas have eruptive temperatures of only ~ 1240°C [Breddam, 2002], close to those of similarly magnesian N-MORB [Ford *et al.*, 1983]. Picrite glass, an indicator of high temperatures, is not found.

A plume is postulated to have migrated southeast from beneath the Greenland craton at ~ 60 Ma to underlie southeast Iceland at present [Lawver & Muller, 1994] (Figure 1). However, no time-progressive volcanic track is observed, such as occurs at Hawaii. Volcanism has been focused at the mid-Atlantic ridge (MAR) since the opening of the Atlantic at ~ 54 Ma.

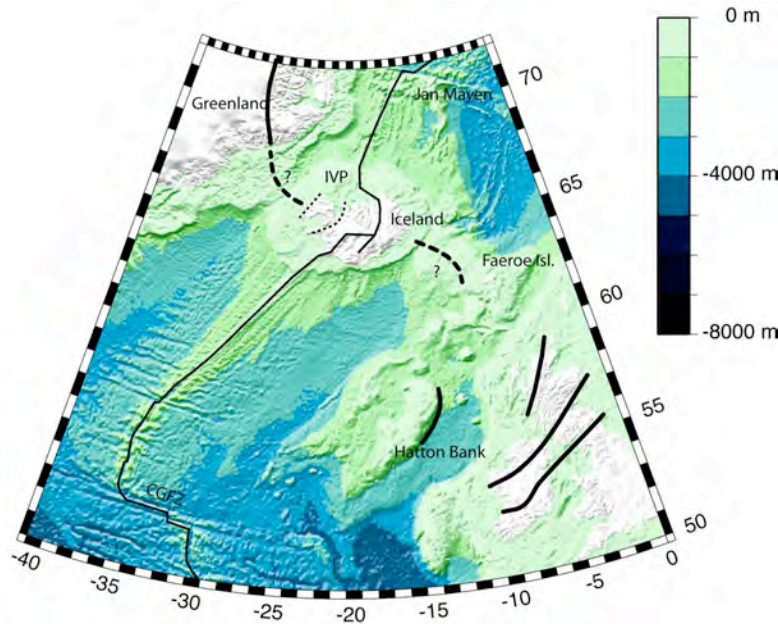


Figure 1: Present-day bathymetry of the north Atlantic, showing the Greenland-Iceland-Faeroe bathymetric ridge which is underlain by crust ~ 30 km thick. Other shallow areas are blocks of stretched continental crust. Thin black line: MAR; thin dashed black lines: extinct ridges; thick lines: faults of the Caledonian suture [Soper et al., 1992]; thick dashed line: inferred trend of suture crossing the Atlantic Ocean [Bott, 1987], IVP: Iceland volcanic plateau, CGFZ: Charlie Gibbs fracture zone.

Seismic tomography yields no evidence for a plume-like structure in the lower mantle [Ritsema et al., 1999], which is predicted by the plume hypothesis. Instead, both the strength and the non-cylindrical morphology of the upper-mantle seismic anomaly show that it extends only down to the mantle transition zone (Figure 2) [Foulger et al., 2000; Foulger et al., 2001]. Tomographic cross sections illustrating a continuous, low-wave-speed body extending from the surface to the core-mantle boundary beneath Iceland have been produced by over-saturating the colour scale, which imparts the visual impression of significance to weak bodies in the lower mantle that are at the noise level and have not been confirmed by other studies [Bijwaard & Spakman, 1999]. Cross sections are also truncated to remove similar bodies beneath the Canadian shield and Scandinavia, where plumes are not expected.

Crustal structure in Iceland provides no support for the plume hypothesis. The crust is thinner beneath western Iceland than eastern Iceland, the opposite of what is expected for an eastward-

migrating plume (Figure 3). The local gravity field can probably be explained by crustal structure alone, without need for large density anomalies in the mantle. Historically, crustal seismic data from Iceland have been interpreted both as indicating that the crust is thin and the mantle beneath hot, and that the crust is thick and the mantle beneath cool. Both of these models have been considered to be consistent with the plume hypothesis, illustrating well that the model of a plume beneath Iceland is an *a-priori* assumption, and not an hypothesis [Foulger *et al.*, 2003c]. Current interpretations of crustal seismic data suggest that the crust beneath Iceland is cooler than at similar depths beneath the East-Pacific Rise [Menke & Levin, 1994].

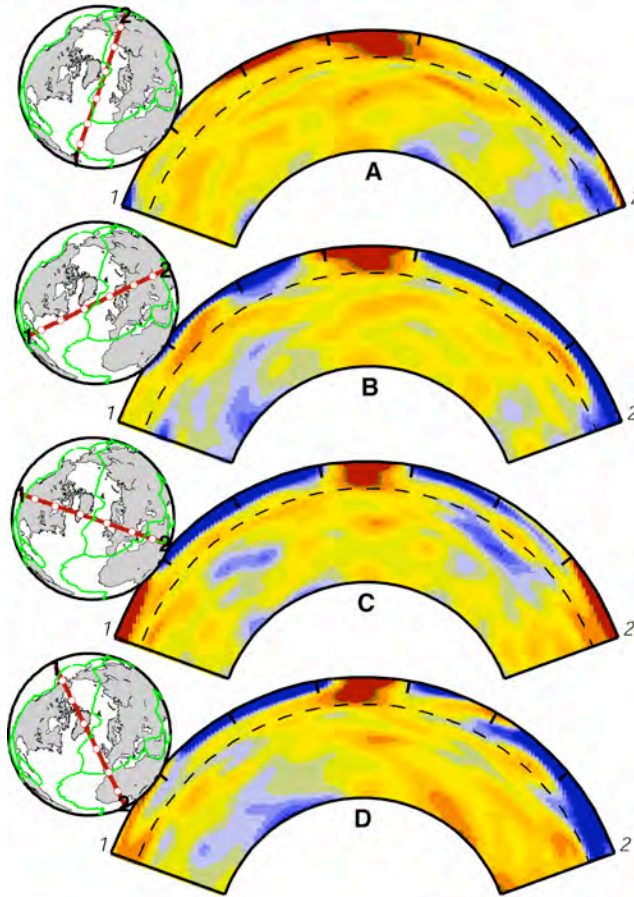


Figure 2: Tomographic cross sections through Iceland [Ritsema *et al.*, 1999]. Courtesy of J. Ritsema.

Although the geochemistry of Iceland is somewhat different from that of the adjoining Reykjanes ridge, interpretation in terms of mixture of a plume component and MORB is problematic. The geochemistry cannot be explained by mixing two components, but “depleted” and “enriched” plume components must be invoked, and perhaps an additional component to explain the helium isotope data. $^{87}\text{Sr}/^{86}\text{Sr}$, postulated to be a plume tracer, increases away from the presumed centre of the plume to the Icelandic shelf edge [Schilling *et al.*, 1983] – the opposite of what is expected. Na_2O and TiO_2 are much higher in Iceland than on the adjoining ridges, again the

opposite of what is expected for the greater extent of melting and depth range of melting required to explain the thicker crust [Klein & Langmuir, 1987].

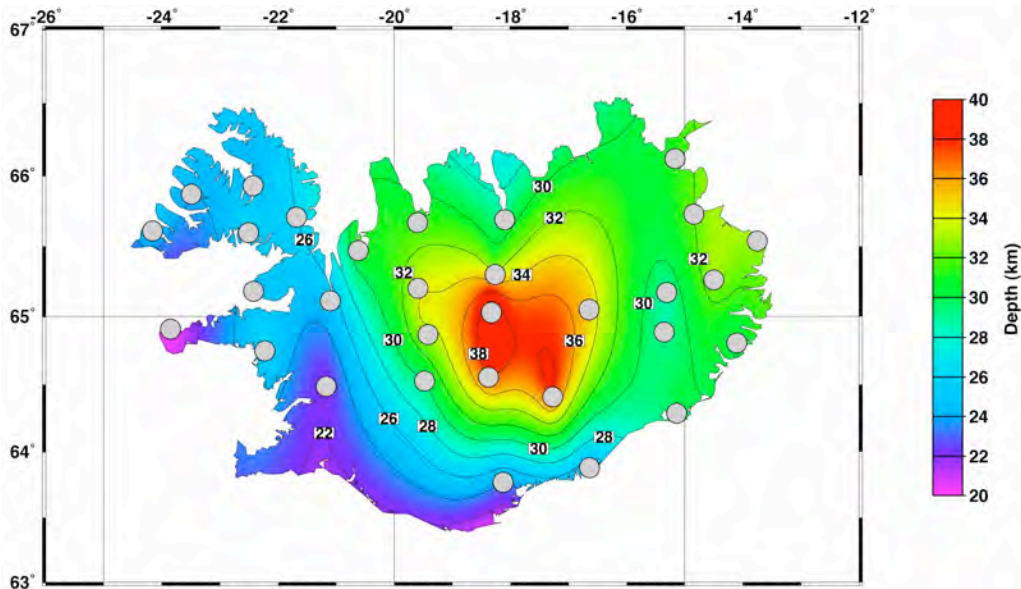


Figure 3: Contour map of the depth to the base of the lower crust (defined as the depth to the $V_s = 4.2$ km/s horizon) (from Foulger et al., 2003c).

What must be explained?

In order to explain Iceland, a model is required that can account for the production of 2-3 times more melt at the MAR between $\sim 63^{\circ}30'$ and $\sim 66^{\circ}30'$ than elsewhere on the ridge, at similar temperatures. A mantle composition that is more fusible than normal peridotite is probably the only option. Iceland and the North Atlantic Volcanic Province formed in the Caledonian suture, which was created at ~ 400 Ma when what are now Greenland and Scandinavia collided as the Iapetus ocean closed (Figure 4). This suture is thus the site of earlier subduction, and abundant eclogite is expected. Eclogite is the high-pressure form of basalt, and is created when oceanic crust is transported into the Earth at subduction zones. The latest-subducting crust would probably have been fairly young and hot. It would thus not have been dense enough to sink deep into the mantle, but would have remained, abandoned, in the upper mantle.

A shallow model involving plate tectonic processes

Eclogite, and eclogite-peridotite mixtures, have lower liquidus and solidus and a narrower melting interval than peridotite [Yaxley, 2000]. At typical mantle temperatures, where peridotite melts to the extent of just a few percent, eclogite is almost completely molten. In the case of eclogite-peridotite mixtures, up to several times the amount of melt is expected than from pure peridotite. Thus the volume of melt at Iceland can be explained by processes the same as elsewhere along

the MAR, but occurring where the mantle is fertilised by eclogite in the ancient subduction zone within which it lies. This model suggests that Iceland can be explained by passive upwelling only, and it predicts that isentropic decompression melting of an eclogite-rich source can produce 2 – 3 times as much melt volume than the same process involving peridotite only.

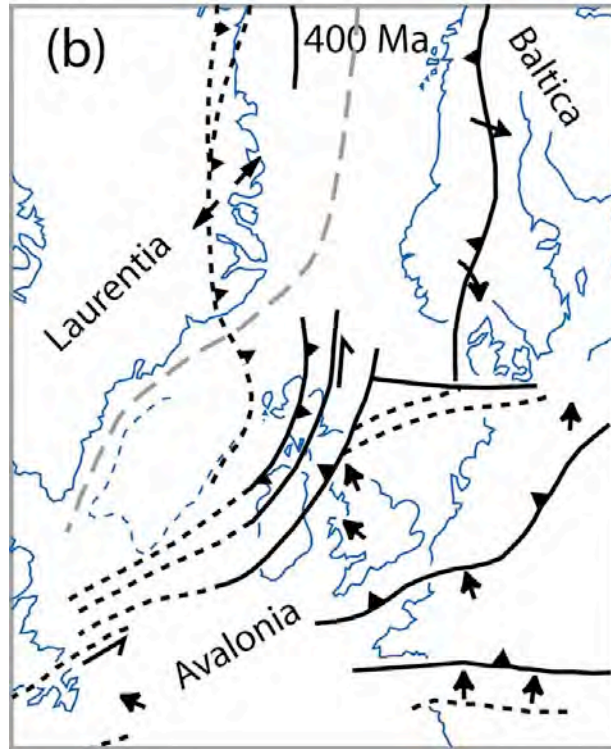


Figure 4: Closure of the Iapetus ocean at 400 Ma. Arrows: convergence directions; thick lines: faults and orogenic fronts. Black triangles indicate sense of thrust faults. Slabs were subducted beneath Greenland, Baltica and Britain [after Soper et al., 1992]. Dashed grey line indicates position of MAR that formed at ~ 54 Ma.

There is evidence for remelted crust of Caledonian age in the basalts of east Greenland, Iceland and Britain from calculated compositions of parental melts, trace- and rare-earth elements (REE) and radiogenic isotope ratios [Breddam, 2002; Chauvel & Hemond, 2000; Korenaga & Kelemen, 2000; Lesher et al., 2002]. A source in extensively melted subducted Iapetus crust can explain the subtle differences in geochemistry between Icelandic basalts and MORB, including REE, trace elements such as Zr/La, Nb-Y-Zr systematics, isotopic and noble-gas data (Figure 5, Foulger et al., 2003c).

Oceanic crust comprises a variety of lithologies, including troctolite, olivine gabbro, gabbro-norite, oxide gabbro, and minor residual granitic material. Remelting these produces basalts that reflect subtle variations in geochemistry inherited from the fractionation history of the corresponding mineralogy. What has been termed the “depleted plume component” [Kempton et al., 2000] in Icelandic basalts may be derived solely from abyssal gabbro, and the

“enriched plume” component may be derived from remelting axial or seamount E-MORB, AOB, and associated intrusive rocks.

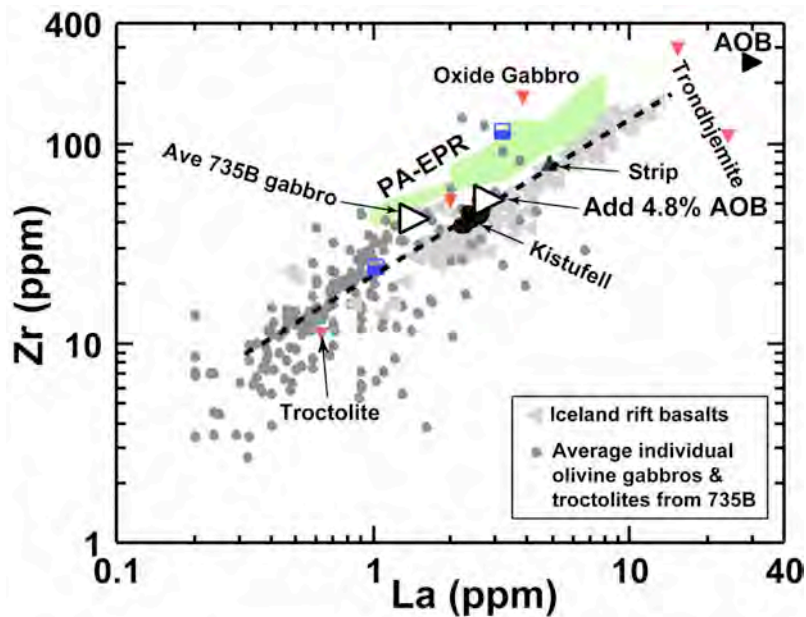


Figure 5: Icelandic basalts can be well modelled as average gabbro from DSDP hole 735B plus 4.8% alkali olivine basalt.

The high $^3\text{He}/^4\text{He}$ observed in Iceland is probably of Caledonian age, and preserved in olivine crystals in the subducted slabs. Olivine traps helium in gas bubbles, and since U+Th is essentially absent from olivine crystals, old, high- $^3\text{He}/^4\text{He}$ ratios are preserved until such time as the olivine is remelted [Anderson, 1998; Natland, 2003].

Summary

This model for the Iceland melt extraction anomaly, which involves shallow plate tectonic processes only, can explain the melt distribution, temperature, seismology, petrology and geochemistry of the region more plausibly and with less special pleading and fewer appeals to coincidence than models involving a hot plume.

The Iceland region serendipitously offers many clues to its genesis, and has been unusually well studied. The lessons learned there may give clues to the origin of other “hotspot” melt anomalies. The essential ingredients are mantle fertility arising from crust subducted when it was still relatively hot and young, or from some other source such as the continental lithosphere, and lithospheric extension. In this context it would seem no coincidence that many “hotspots” occur on or near ridges, triple junctions, faults and sutures. This kind of model attributes melt anomalies to the by products of plate tectonics, and may provide a generic alternative to the plume model.

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On the apparent eastward migration of the spreading ridge in Iceland

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The part of the north Atlantic where Iceland is currently forming has functioned as a tectonic divide since the opening of the ocean, and has persistently featured paired spreading ridges and intervening microplates.

At the time of ocean opening a ~ 100-km long, right-stepping transform fault, the Faeroe transform fault (FTF), formed where the new spreading ridge crossed the Caledonian suture [Bott, 1985] (Figure 1a). Tectonic complexities subsequently rafted several continental blocks into the ocean, including Jan Mayen, the Jan Mayen microcontinent (JMM) and the Faeroe block.

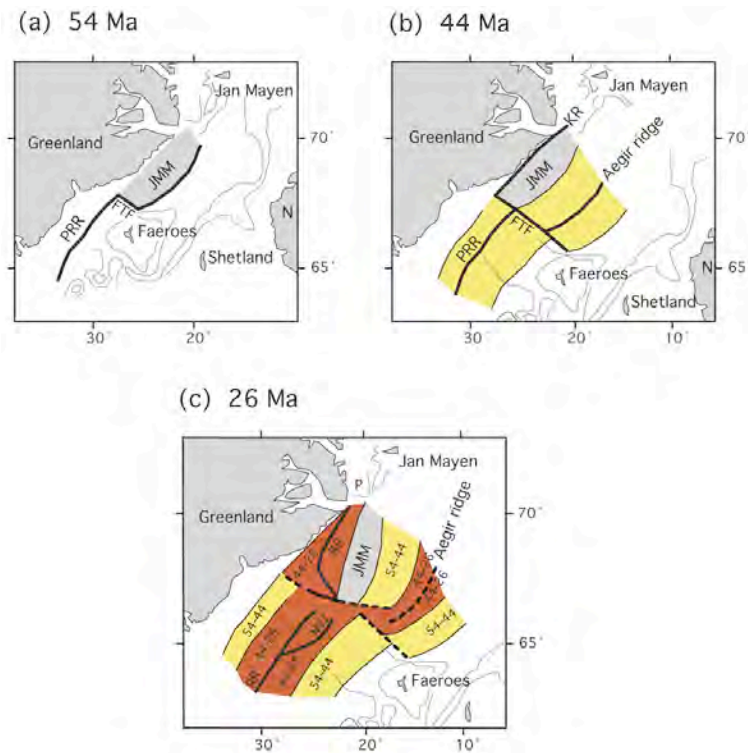


Figure 1: Tectonic evolution of the Iceland region at 54, 44 and 26 Ma [from Bott, 1985]. Gray: continental crust, heavy solid lines: active plate boundaries, heavy dashed lines: extinct plate boundaries, thin lines: bathymetric contours, FTF: Faeroe transform fault, KR, PRR, RR: Kolbeinsey, proto-Reykjanes, Reykjanes ridges, NVZ: Northern volcanic zone, JMM: Jan Mayen microcontinent, N: Norway, P: pole of rotation of JMM 44 – 26 Ma, numbers: sea floor age in Myr.

Spreading proceeded relatively simply for the first ~ 10 Myr following opening, but at ~ 44 Ma, a major reorganization occurred north of the FTF. A second spreading center, the Kolbeinsey ridge, developed within the Greenland craton (Figure 1b). Complimentary fan-shaped spreading then occurred along both the original Aegir ridge and the Kolbeinsey ridge during the period 44 - 26 Ma, causing ~ 32° of counterclockwise rotation of the intervening, continental JMM (Figure 1c). This resulted in up to ~ 60 km of transtensional extension across the FTF, which corresponds to opening at up to ~ 15% of the local full spreading rate of 1.9 cm/a. The onset of this phase of transtensional extension coincides with a time when the magmatic production rate increased greatly, and formation of the ~ 250-km-wide Iceland-Faeroe ridge of thickened crust gave way to formation of the ~ 600-km-wide Iceland volcanic plateau. This increased magmatism may have been permitted by the transtensional opening across the FTF.

In the neighborhood of Iceland spreading continued about a pair of parallel centers that progressively migrated south through the repeated extinction of old rifts and the opening of new. At ~ 26 Ma the Aegir ridge became extinct, spreading became confined to the Kolbeinsey ridge, and a second parallel spreading center formed to the immediate south (Figure 1c). The eastern center developed into the currently active Northern Volcanic Zone (NVZ) (c.f., Figures 1c and 2a). Continuity of the lava succession in eastern Iceland, and the absence of an eastward extension of the Tjornes fracture zone beyond the NVZ shows that the eastern center has been long-lived and has remained approximately fixed with respect to the Kolbeinsey ridge [Jancin *et al.*, 1985; Saemundsson, 1979].

As a consequence of the fixed spatial relationship of the eastern spreading center with the Kolbeinsey ridge, the western center was progressively transported west relative to the Kolbeinsey ridge. This center experienced at least two rift extinctions, accompanied by the opening of new rifts further east with better colinearity with the Kolbeinsey ridge. At ~ 15 Ma the western rift became extinct and a new rift opened ~ 80 km further east (Figure 2a), and at ~ 7 Ma this new rift in turn became largely extinct and the presently active Western Volcanic Zone (WVZ) formed (Figure 2c). Extension across a pair of parallel spreading centers has occurred in south Iceland since ~ 2 Ma when the Eastern Volcanic Zone (EVZ) formed [Saemundsson, 1979] (Figure 2d).

The continually evolving, parallel-pair spreading center configuration has resulted in a jigsaw of ephemeral microplates. These include the JMM and two microplates in Iceland. One lay between the northern pair of parallel spreading centers (the Trollaskagi microplate) and one between the southern pair (the Hreppar microplate). The palinspastic reconstruction (Figure 2) shows that oceanic crust up to ~ 30 Myr old has been trapped beneath central and southeast Iceland as a consequence of the complex history of spreading. Recent ancient ages determined for zircons from basalts in southeast Iceland suggest the presence of continental crust beneath that area, suggesting that the JMM may have continued further south than shown in Figure 1c [Amundsen *et al.*, 2002]. In this case, a sliver of continental crust might have been captured between the parallel pair of spreading centers that formed at ~ 26 Ma, and currently underlie central and southeastern Iceland. The presence of such a sliver may have been influential in the formation of the parallel spreading-center pair.

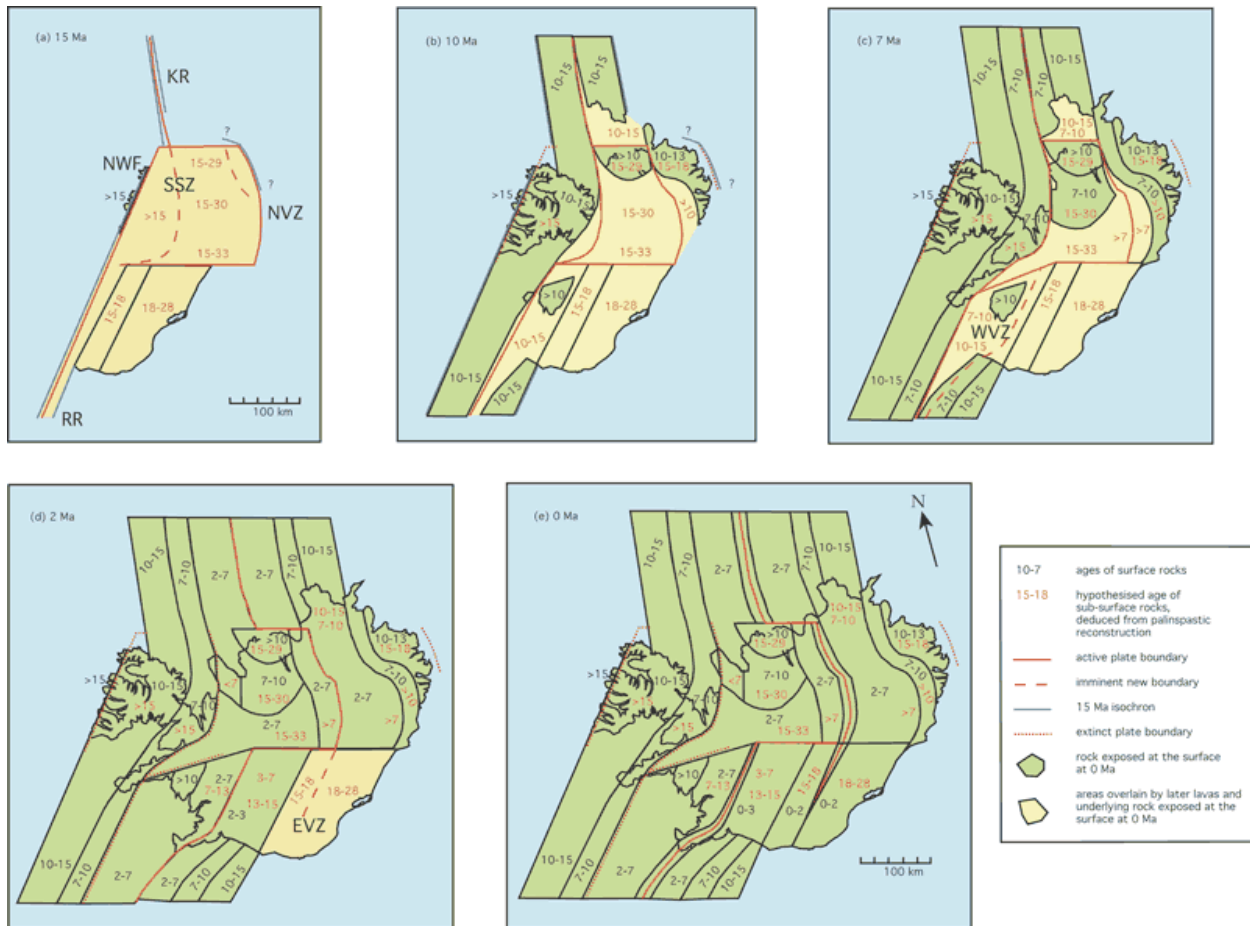


Figure 2: Tectonic evolution of Iceland at 15, 10, 7, 2 and 0 Ma. Black lines indicate the boundaries of modeled blocks. The oldest rocks exposed in Iceland are 17 Myr old. Blue: unmodeled oceanic areas, yellow: areas that are part of present-day Iceland but are now covered with later lavas, green: rock currently exposed at the surface, solid, dashed and dotted red lines: active, imminent and extinct plate boundaries, blue lines: inferred position of the 15 Ma isochron, red and black numbers: ages of rock in Myr, red: ages of rocks currently covered with younger lavas, black: ages of surface rocks. Where ages are shown in both black and red for blocks in later panels, older rock underlies younger surface rock. The extent of rock in a given age range offshore is inferred assuming a spreading rate of 1.9 cm/a about the KR and the RR. On land, the possible age range of rocks in a given block is deduced assuming spreading is equally distributed between parallel spreading center pairs, where these exist. The ages of rocks onland are taken from the published literature. The positions of the extinct NWF (Northwest Fjords) and SSZ (Snaefellsnes-Skagi zone) rifts are taken from Saemundsson [1979]. Other extinct rifts are required by space considerations, given the ages of surface rocks.

Eastward migration of spreading in Iceland is often cited as evidence in support of an eastward-migrating plume. The palinspastic reconstruction (Figure 2) shows, however that the locus of spreading has not migrated east. The regular progression of basalts extending from the NVZ to

the east coast shows that spreading has proceeded about the eastern center since at least ~ 15 Ma [Bott, 1985; Jancin *et al.*, 1985; Saemundsson, 1979] and marine observations suggest that this situation has probably been relatively stable since ~ 26 Ma [Bott, 1985]. Spreading about a parallel pair of ridges is unstable and, given the fixed location of the eastern rift relative to the Kolbeinsey ridge, the western rift has been required to re-equilibrate by repeated relocations to the east or else be rafted away with the north American plate. These relocations do not represent an eastward migration of the general locus of melt extraction in the region, which has remained fixed relative to the Kolbeinsey ridge. Indeed, southerly ridge migration is a more prominent characteristic of the Iceland region than easterly migration.

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The Emperor and Hawaiian Volcanic Chains

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For about 20 years there has been no serious challenge to the deep thermal mantle plume hypothesis for the Hawaiian and Emperor chains, despite the fact that many features do not conform to this hypothesis. These include:

1. The great “bend”, near the Mendocino fracture zone, where the Emperor seamount chain ends and the Hawaiian chain begins does not result from a change in direction of motion of the Pacific plate [Norton, 1995; Richards and Lithgow-Bertelloni, 1996; Raymond *et al.*, 2000] (see Figure). The collision of India and Asia has been ongoing for much longer than the time taken for the bend to form, and this has little effect on the motion of the Pacific plate [Richards and Lithgow-Bertelloni, 1996]. The driving forces of plate tectonics are mainly thermal (ridge push and slab-pull) and the integrated effect, which dominates the direction of plate motion, cannot change rapidly, although local stresses and resisting forces can.
2. The locus of active volcanism has not remained fixed in any reference frame except itself [Clague and Dalrymple, 1987; Tarduno and Cottrell, 1997]. It moved south by ~ 800 km relative to a geographic reference frame while the Emperor Seamount chain formed [Butt, 1980; McKenzie *et al.*, 1980]. At the bend, the hotspot migration rate increased from ~ 7 cm/year to ~ 9 cm/year relative to the Pacific sea floor. The Hawaiian and Emperor chains may be separate and independent phenomena.
3. The Emperor chain began at or near a ridge, as shown by MORB-like ⁸⁶Sr/⁸⁷Sr values at the oldest end [Keller *et al.*, 2000], Pacific plate palinspastic reconstructions [Engebretson *et al.*, 1985; Smith, 2003], and the elastic thickness of the lithosphere beneath the northern Emperors [Watts, 1978]. This is a coincidence in the plume hypothesis. Many other hotspots are also on ridges or their tracks started on ridges.
4. There is no evidence for a Hawaiian “plume head”. Oceanic plateaus are not subductable [Abbott *et al.*, 1997], and if a “plume head” had existed it would have been scraped off, accreted or obducted onto the Aleutian/Kurile/Kamchatka arc. There is no evidence for such material. Alternative hypotheses for the chain e.g., propagating cracks, stress-induced

volcanism, do not require a large igneous province at the beginning of the chain, but this association is fundamental to the plume hypothesis [*Campbell and Griffiths, 1990; Campbell and Griffiths, 1993*].

5. The volume flux along much of the chain has typically been $\sim 0.01 \text{ km}^3/\text{yr}$. It dropped essentially to zero for $\sim 10 \text{ Myr}$ following the bend, but over the last 5 Myr has been an order of magnitude greater than the average rate [*Bargar and Jackson, 1974, B. Eakins, USGS, unpublished results*] which is similar to current Pu'u 'O'o eruption rates of $0.113 \text{ km}^3/\text{yr}$. The eruption rate correlates with the propagation rate of the melt locus, which has doubled over the last 2 Myr [*Shaw et al., 1980; Clague and Dalrymple, 1987*]. Magmatic rate would be expected to be anticorrelated with migration rate for a plume source with a steady supply rate and, in the plume head-tail model, to decline with time. This is the opposite of what is observed. Thermal models do not explain the high flux rate beneath thick plates, where the top of the productive part of the melting column is missing [e.g., *Cordery et al., 1997*].
6. Heat flow across the Hawaiian bathymetric swell shows no significant anomaly [*von Herzen et al., 1989; Stein and Stein, 1992; 1993*], and the swell surrounding the southernmost part of the Hawaiian chain cannot therefore be explained as a thermal effect. In the plume model it must be attributed to compositional effects with no discernible thermal effect at the surface [*Liu and Chase, 1989; Sleep, 1994*]. Other models do not have this problem.
7. There is no discernable thermal rejuvenation or thinning of the lithosphere as it passes over the site of active volcanism, as predicted in the plume hypothesis.
8. Petrological estimates of the temperature anomalies beneath Hawaii compared with ridges vary from zero to a maximum of 200°C [*Green et al., 2001; Gudfinnsson and Presnall, 2002*]. This is no greater than the range observed for ridges away from hotspots relative to the mean ridge temperature. Plume models typically require temperature anomalies of $200\text{--}600^\circ\text{C}$ [e.g., *Cordery et al., 1997*]. A small change in temperature, volatile content or fertility of the upper mantle can lead to a large change in the extent of partial melting and melt volumes [*Yaxley, 2000; Green et al., 2001; Asimow and Langmuir, 2003*].
9. The petrology of Hawaiian lavas suggests that the melt comes from $\sim 80\text{--}120 \text{ km}$ depth or shallower – near or above the base of the lithosphere. No petrological data require a deeper source.
10. In addition to not requiring a hot source, the geochemistry of Hawaiian basalts does not require a deep mantle source. The ultimate origin of OIB material has been suggested to be the deepest mantle, but may also be old mantle wedge material, the asthenosphere, and a shallower layer which accumulates subduction-zone products [*Anderson, 1989; 1994; 1995; 1996*]. In the deep-plume model, subduction-zone material is carried from the surface down to the core-mantle boundary and back up in the core of the plume [*Hofmann and White, 1982*]. This is inconsistent with the widespread distribution of OIB at rifts and seamounts throughout the Pacific. The geochemistry varies geographically and temporally. *Mukhopadhyay et al.* [2003] report variations in $^3\text{He}/^4\text{He}$ ratio of up to 8 Ra during a single century in Kauai volcano, and spatially, such that different volcanoes do not appear to be fed

by the same magma source. “End-member” and principal-component interpretations require at least four different source components to explain the geochemistry of Hawaiian lavas, which suggests a spatially distributed, compositionally inhomogeneous, and temporally variable source that is sensitive to shallow lithospheric features. High maximum $^3\text{He}/^4\text{He}$ ratios have been attributed to a lower-mantle component, but this interpretation is flawed [Anderson, 1998b; Anderson, 1998a; Foulger and Pearson, 2001; Meibom et al., 2003]. Hawaiian basalts exhibit a wide variation in helium contents and ratios and generally have low ^3He contents.

11. Seismic tomography reveals no plume-like low-wave-speed anomalies in the upper ~ 150 km under the big island [Ellsworth, 1977; Wolfe et al., 2002]. Shear waves from a large earthquake there in 1973, that reflected off the core and registered on a seismometer on Oahu (*ScS* waves), indicate that the average *S*-wave speed of the mantle beneath the Hawaii region is higher than the average beneath the southwestern Pacific ocean [Best et al., 1975], and that the propagation efficiency is high, contrary to expectations for regions of high temperature or partial melting (see abstract by B. Julian, this conference). Whole-mantle tomography [e.g., Ritsema et al., 1999] reveals that the mantle beneath whole southern half of the Pacific ocean has low wave speeds, and on the scale of resolution of these studies (a few hundred km) Hawaii is not anomalous compared with the region as a whole. Thick pancakes of high-wave-speed material, that might be expected to characterize plume heads beneath the lithosphere or the 650-km discontinuity, have not been detected.

Beneath the Hawaiian region, transition zone thickness [an index of mantle temperature Anderson, 1967] is ~ 229 km [Gu and Dziewonski, 2001]. This is ~ 13 km thinner than the global average of 242 km, but not significantly thinner than the transition zone throughout much of the central Pacific and other oceans.

Contrary to popular belief and countless textbooks, courses and web pages, Hawaii is poorly explained by the plume model. A fully quantified alternative hypothesis is long overdue.

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The Kerguelen Plume: What We Have Learned From ~120 Myr of Volcanism

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The Kerguelen Plume has had a major role in creating major volcanic features in the Eastern Indian Ocean over the last ~120 myr. In order to understand this role, igneous basement has been drilled and cored at 9 sites on the Kerguelen Plateau, 2 sites on Broken Ridge and 7 sites on the Ninetyeast Ridge^(1,2,3,4). In addition, stratigraphic volcanic sections on the two relatively young islands (Kerguelen and Heard) constructed on the Kerguelen Plateau have been studied^(5,6), as well as dredged samples from seamounts defining a linear trend between these islands⁽⁷⁾. Major results are:

(a) The Kerguelen Plateau began forming at ~120 Ma, after Gondwana breakup. Eruption ages decrease from ~120 Ma in the southern plateau to ~95 Ma in the central plateau. This age range is not consistent with a pulse of volcanism associated with melting of a single, large plume head.

(b) The sampled volcanic portion of the plateau is dominantly tholeiitic basalt that formed islands, but the waning stage of volcanism included alkalic basalt and highly evolved, explosively erupted trachytes and rhyolites.

(c) At several geographically dispersed locations on the plateau, the Cretaceous tholeiitic basalt has been contaminated by a component derived from continental crust. Geophysical data are consistent with continental crust in the oceanic lithosphere and clasts of ancient garnet-biotite gneiss occur in a conglomerate intercalated with basalt on Elan Bank.

(d) The Ninetyeast Ridge is a 5000 km volcanic feature composed of tholeiitic basalt whose eruption age increases from south to north (~38 to 82 Ma), as expected for a hotspot track.

Although these lavas have diverse geochemical characteristics there is no evidence for a continental component.

(e) The oldest lavas (~30 Ma) on the Kerguelen Archipelago are tholeiitic to transitional basalt but younger archipelago lavas (<25 Ma) and Heard Island lavas are alkalic. This change to alkaline volcanism reflects a decreasing magma flux from the plume accentuated by the thick Cretaceous plateau lithosphere underlying the islands.

(f) Seamounts between the Kerguelen Archipelago and Heard Island are formed of 18-21 Ma alkalic basalt that may reflect the Tertiary track of the Kerguelen plume.

(g) Isotopic ratios (Sr, Nd, Hf and Pb) range widely in lavas from the Kerguelen Plateau, Ninetyeast Ridge, Kerguelen Archipelago and Heard Island. A significant role for continental crust is obvious at several locations on the Kerguelen Plateau and in a single trachyte from Heard Island. In contrast, there is no evidence for continental crust contamination in lavas from the Ninetyeast Ridge and Kerguelen Archipelago. The isotopic diversity of these lavas reflects plume heterogeneity and mixing of plume and MORB-related asthenosphere and lithosphere.

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Perturbations to the Galápagos Hotspot due to Interaction with the Galápagos Spreading Center

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The Galápagos Islands have a number of attributes that are not readily explicable by simple mantle plume theory. The unique aspect of the Galápagos is that they lie close to, but not on, a nearby plate boundary, the Galápagos Spreading Center, which is between 30 and 300 km north of the active volcanoes. The plate boundary results in high heat flow, anomalous stresses in the Nazca Plate, and lithosphere that has large changes in thickness over short distances. Thus, the exceptional features of the Galápagos hotspot may be explained either in terms of alternatives to deeply-rooted mantle plumes, or by perturbations to the plume by the nearby plate boundary.

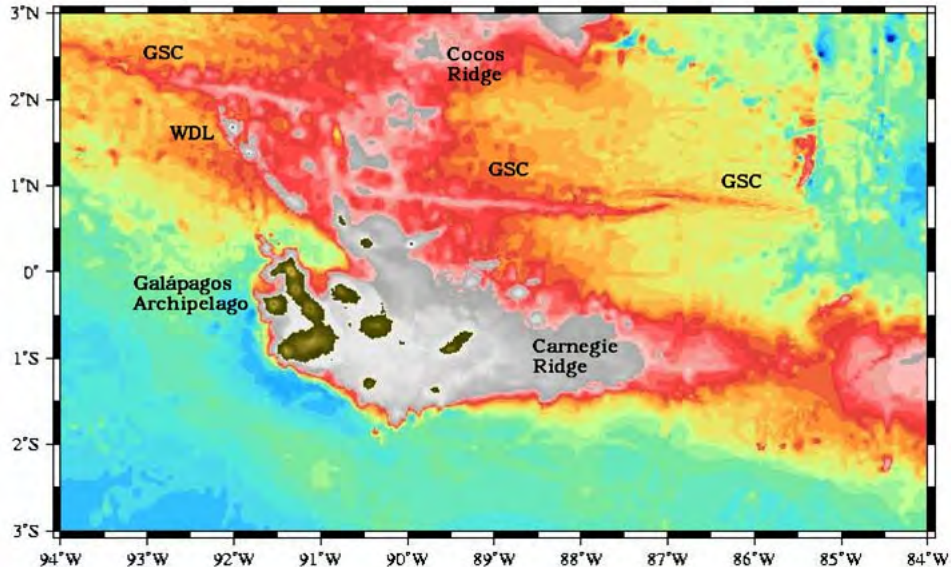


Figure 1: Bathymetric map of the Galápagos region, produced by Dr. William Chadwick. GSC = Galápagos Spreading Center, which separates the Cocos plate to the north from the Nazca Plate to the south; WDL = Wolf-Darwin Lineament (Harpp and Geist, 2002). Carnegie and Cocos Ridges are thought to be traces of the hotspot.

Some of the features that may be related to hotspot-ridge interaction are:

1. The volcanoes form lineaments (e.g. the WDL) whose orientations have no straightforward relationship to the direction of plate motion or the nearby Galápagos Spreading Center. A steadily-deepening aseismic ridge, the Carnegie Ridge, extends eastward from the islands, however. Thus, depending on the scale at which one looks, there is either no relation between the distribution of islands and plate motion, or a near-perfect relation.
2. The age progression of the volcanoes is irregular. Although the eastward plate motion of the Nazca plate would suggest that only the westernmost volcanoes

- should be active, Holocene volcanism has occurred throughout the archipelago, and there is no geological or geochemical evidence that the young volcanism is comparable to posterosional volcanism at Hawaii. The vast majority of historical volcanism has been in the western part of the archipelago, however, and the oldest volcanoes are in the east. Also, sparse data indicate that seamounts on the Carnegie Ridge are progressively older to the east.
3. Some lavas from the central and northern part of the archipelago (particularly Wolf volcano on Isabela, Santa Cruz, and Genovesa volcanoes) are among the most isotopically depleted ocean island volcanoes on the planet. These lavas are indistinguishable from EPR MORB in their Sr, Nd, Hf, He, and Pb isotopic ratios. In contrast, the volcanoes from the periphery of the archipelago erupt lavas with more enriched isotopic ratios. This has been attributed both to plume-asthenosphere mixing and simply to intrinsic features of the Galápagos hotspot.
 4. There is not a single enriched source. Instead, the southern enriched source is isotopically distinct from the northern one, which is in turn distinct from the western one. Whether or not the source of the magmas is a plume, the melting region is isotopically zoned.
 5. Although the Galápagos is a high $^3\text{He}/^4\text{He}$ hotspot, the helium isotopes are decoupled from other indicators of an enriched source. In other words, the volcano with the highest $^3\text{He}/^4\text{He}$ is Fernandina, but those lavas possess intermediate Sr, Nd, Pb, and Hf isotopic ratios.
 6. There is no Loihi-type edifice at the leading edge of the hotspot; instead, the subaerial volcanoes abruptly rise from the surrounding seafloor. The building and coalescence of large terraces (10 x 20 km in spatial dimension and up to 500 m high) appear to be the major means by which the Galápagos Platform is constructed.
 7. Cocos Island, which lies on the Cocos Ridge, is millions of years younger than could have been produced at the hotspot. The Cocos ridge is essentially flat; it does not follow the subsidence path predicted for cooling lithosphere.

As an alternative to a plume origin, many of the Galápagos Islands' features might be explained by a zoned, long-lived melting anomaly that resides in the upper mantle. A critical observation, however, is that the Galápagos hotspot has produced different amounts of crust on the Nazca and Cocos plates over time, as shown by closely-spaced bathymetric profiles across the Carnegie and Cocos ridges. The most straightforward explanation for the bathymetry is that there has been differential motion between the plate boundary and the melting anomaly. Results from a recent seismic experiment indicate that the transition zone is 50 km thinner to the SW of Fernandina (Hooft et al., 2003), which appears to require a thermal anomaly of about 320° both at the transition zone's upper and lower boundaries. This anomaly in the transition zone immediately underlies a vertically-oriented, seismically-slow region in the upper mantle (Toomey et al., 2001 and in prep). At shallow depths in the mantle, the low velocity anomaly is bent to the east, as if it were being sheared by the eastward motion of the Nazca plate, which was predicted by geochemical models.

In search for an Iceland plume: long period magnetotellurics

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Introduction

The method of magnetotellurics is used to determine the electrical resistivity distribution within the earth from measurements of natural, time-varying, magnetic and electrical fields at the earth's surface. The fields originate from electrical currents in the ionosphere and magnetosphere which induce secondary currents in the conducting earth. Since electromagnetic fields of longer periods are less attenuated than fields of shorter periods (skin effect), the observation of longer periods reveals information about greater depths. The estimation of frequency dependent complex transfer functions between different field components can be related to models of the resistivity distribution. To gain information about the lower crust and upper mantle periods ranging from minutes to several hours are required, which leads to the specialized method of long period magnetotellurics (LMT).

The resistivity of crust and mantle materials is influenced by several factors, e.g. the presence of partial melt. Given a certain geodynamical model, theoretical resistivities can be calculated using empirical relationships between resistivity, material composition, melt content and temperature, known from laboratory experiments. The resulting resistivities can be compared to results from LMT measurements. This way LMT measurements can contribute a constraint for geodynamical models. Modeling studies by Kreutzmann (2002) within the „Iceland plume dynamics project“ (IPDP) show that the presence of a plume head beneath Iceland would produce a significant magnetotelluric signal if melt content is higher than 1%.

Earlier work by other authors give detailed insight into the resistivity distribution within the Icelandic crust. One remarkable result is the presence of a sheet-like conductor in approximately 20 km depth beneath most parts of Iceland (e.g. Thayer 1981, Beblo 1983). However, these datasets don't cover periods long enough to study features in greater depths. Therefore three field sites for long period

monitoring were set up by Beblo and Björnsson in 1999 and in 2001 a joint project was initiated by the universities of Akureyri (Björnsson), Munich (Beblo) and Frankfurt (Junge), titled „Continuous Monitoring of the Icelandic Crust and Mantle Resistivity“ (CMICMR). Its objective is to collect, process and analyse continuous LMT data from several stations on Iceland to obtain a reliable resistivity model of the upper mantle beneath Iceland, which hopefully will help to reveal information about the presence or absence of a plumehead.

Data collection and processing within the CMICMR project

Since 1999 three LMT sites were in continuous operation (see Figure 1). They were located at Akureyri (AKU), Húsafell (HUS) and Skrokkalda (SKR). In 2002 SKR was replaced by a new site at Grímstaðir (GRI). The other sites are still in operation. Each station records the two horizontal electric field components and all three magnetic field components with a sampling period of about 1 s, theoretically enabling the coverage of the period range from about 8 s down to DC. So far the data processing was concentrated on periods between 8 s and 9 h.

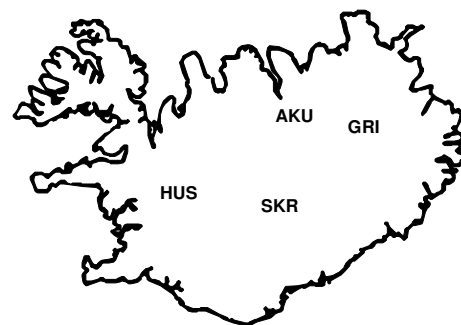


Figure 1: Location of LMT sites on Iceland.

The observed timeseries are split up into segments which are Fourier transformed to calculate transfer functions between different field components and stations. Thereby a quasi uniform source field is assumed. Deviations from this assumption can lead to systematic errors of the transfer function estimates,

known as so called source-effects. Due to Iceland's location at the edge of the auroral oval it is close to the polar electrojet and the possible presence of strong source effects had to be taken into consideration. Great care was taken to avoid source effects by automatically selecting time segments with low source field inhomogeneity. However, in the end the transfer functions turned out to be more stable in regard to source effects than it was assumed. This improves our confidence in the correctness of the transfer function estimates.

Preliminary results

Early analyses of the CMICMR dataset were performed by M. König (1999) and C. Salat (2002) and are continued by S. Golden. Figure 2 shows apparent resistivity and phase curves for the station Húsafell. The apparent resistivity first descends to a minimum around 200 s, which is related to the good crustal conductor. Beyond that there is a slight increase of the apparent resistivity followed by a second decrease which at most sites/polarizations forms a second minimum around 5000 s. This minimum can be explained by a second good conductor in a depth of about 200 km. 3D modelings by A. Kreutzmann showed that such a minimum can be caused by a mantle plume head or a ridge effect. To discriminate between those two possible causes further modelings are needed, which additionally take transfer functions between horizontal and vertical magnetic field components (tipper) into account.

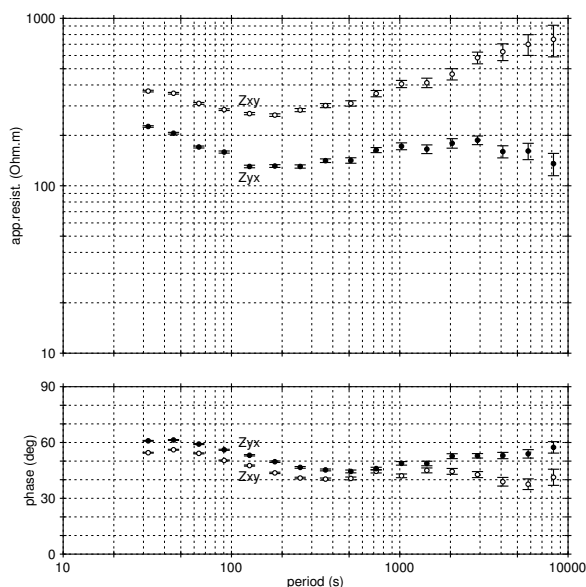


Figure 2: Example of apparent resistivity and phase curves at the site Húsafell (HUS).

Conclusion

Until now magnetotellurics was not able to answer the question on the existence of an Iceland plume. Forward modeling by Kreutzmann has shown that a plume head could produce a significant signal in the magnetotelluric data if it contains more than 1% melt. The detection of such a signal would provide strong evidence for the plume hypothesis. However, the absence of such a signal can't be used as proof against it. In either way, reliable resistivity estimates for the upper mantle will result in additional constraints for geodynamical models. Ongoing work in the CMICMR project is directed to increase the resolution and reliability of resistivity estimates to increase their usefulness for such interpretations.

Further information:

<http://www.geophysik.uni-frankfurt.de/em/icelmt/>

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PRIMARY MAGMAS AT MID-OCEAN RIDGES, 'HOT-SPOTS' AND OTHER INTRAPLATE SETTINGS: CONSTRAINTS ON MANTLE POTENTIAL TEMPERATURES

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The melting regime for mantle-derived basaltic magmas from intraplate settings (kimberlites, olivine melilitites, olivine nephelinites, basanites, alkali-olivine basalts) requires the presence of carbon and hydrogen (dissolved ($\text{CO}_3^{=}$) and (OH) in the melt phase. Magma genesis in intraplate settings can be understood in terms of the lherzolite+(C-H-O) system. Magmas are derived from an "incipient melting" regime which lies at temperatures below the volatile-free lherzolite solidus – marking entry to the "major melting" regime. Experimental petrology and observations of natural magmas concur in defining magma segregation at deep levels and magma transport within channels with only limited wall-rock reaction. In the intraplate settings, the role of an "incipient melting" field in lherzolite + (C-H-O) at depths of >90 km to ? <200 km creates conditions in which both depletion (from N-MORB to D-MORB sources) and enrichment (from N-MORB to E-MORB sources); occur by migration of a 1-2 % melt fraction.

These relationships are summarized in a specific model for a petrological lithosphere [below the silicate solidus for lherzolite +(C-H-O)] and asthenosphere (above the silicate solidus) The proposed model is used to explain production of specific primary magmas in the P,T field between the regionally applicable conductive geotherm and adiabatic upwelling of mantle with potential temperature (T_p) \square 1430°C (FIG. 2).

In many earth models, Mid-Ocean Ridge magmatism is attributed to decompression melting of upwelling upper mantle/asthenosphere at normal mantle temperature and melting has been considered to occur in the absence of significant volatiles (C-H-O). Nevertheless, primitive MORB have water and carbon contents which are not negligible but rather are very low because of relatively high degrees of melting of a source with low carbon and hydrogen contents. Most importantly, primitive mid-ocean ridge basalts have remarkably restricted compositions compatible with saturation with olivine, orthopyroxene, minor clinopyroxene and Cr-Al spinel at pressures around 2 GPa. Processes of melt migration by porous flow may have occurred at deeper levels but at 2 GPa equilibrium between melt and lherzolite mineralogy is indicated. Primary magmas move from this depth through dykes or channels without significant modification by wall-rock reaction, to sub-ridge magma chambers or sea-floor eruption.

In the 'hot-spot' setting, combinations of geophysical arguments (the buoyancy implication beneath the Hawaiian Swell) and geochemical arguments (thicker crust, "garnet signature" on REE implying deeper melting if the anhydrous solidus is used) have been used to infer higher temperature in the mantle, in comparison with the Mid-

Ocean Ridge setting. The comparison of magma and phenocryst compositions between MOR and Hawaiian primitive basalts is a way of testing for evidence for such temperature contrasts, argued in some models to be of the order of 200°C.

The composition of olivine phenocrysts in Hawaiian picrites and in Mid-Ocean Ridge picrites vary up to $Mg^{\#}_{91.3}$ and $Mg^{\#}_{92.1}$ respectively. The compositions and liquidus temperatures of the magmas crystallizing the most magnesian phenocrysts can be estimated and anhydrous liquids temperatures (at 1 bar pressure) of Hawaiian tholeiitic picrites average 1365°C, and for MOR picrites average 1335°C. Water contents of the magmas decrease in the order Hawaiian picrites, E-MOR picrites to N-MOR picrites, and consideration of liquidus depression by these water contents leads to the conclusion that all primitive magmas had liquidus temperatures of approximately 1325°C at \square 1 bar. The data from primitive magmas suggests that the temperature contrast between “Hot-Spot” and MOR primary magmas is $\leq 20^{\circ}C$. Application of information from partial melting studies of model (pyrolite) source compositions and of the liquidus phases of the Hot-Spot and MOR picrites leads to the conclusion that both “Hot-Spot” and MOR primary basalts are derived from mantle with potential temperature $T_p \square 1430^{\circ}C$. Insofar as primitive magmas may be used to infer the potential temperature of their sources, there is no evidence for a temperature contrast of $\Delta T_p = 100-250^{\circ}C$ between “Hot-Spot” or “Deep Mantle Plume” sources and ambient (MOR source) asthenospheric mantle.

Although magma temperatures are similar, the residual mantle compositions for Hawaiian picrites are refractory harzburgites, more refractory (including Cr/Cr+Al ratio) than the lherzolite to harzburgite residue from MOR picrite extraction. It is argued that the buoyancy plume and geophysically anomalous mantle beneath the Hawaiian Swell is due to compositional and not temperature contrasts in the upper mantle. The four-component mixing identified in the Hawaiian source is attributed to interaction between old subducted lithospheric slabs, buoyant or suspended in the upper mantle, and surrounding ambient mantle at $T_p = 1430^{\circ}C$. A cartoon representing the model is presented in Figure 1.

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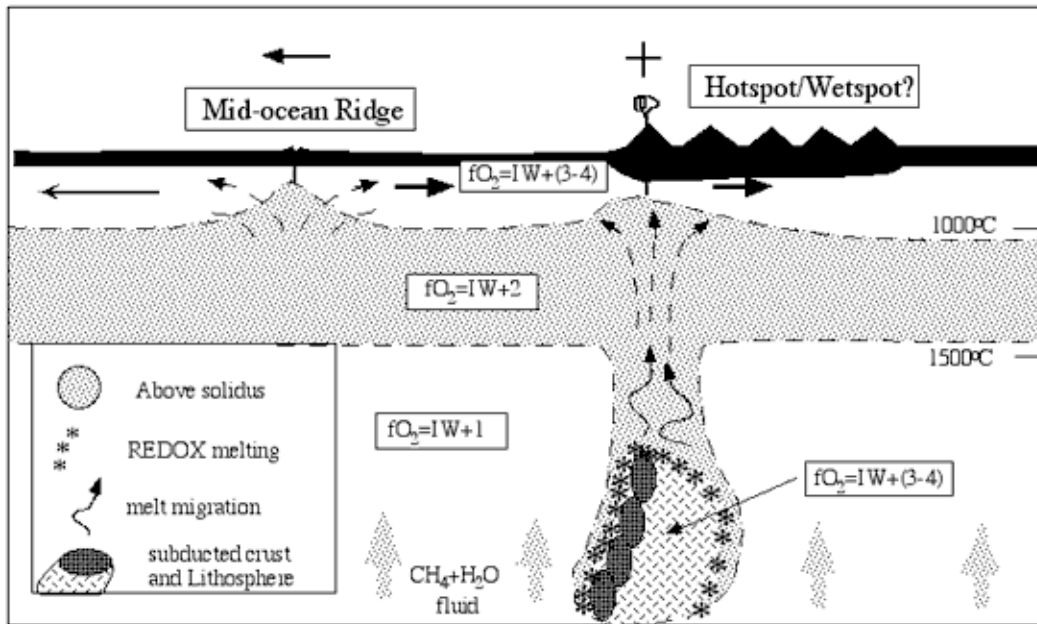


Figure 1

Contrasting Origins of the Most Magnesian Glasses from Iceland and Hawaii

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In a melting column produced by upwelling of the mantle, the most magnesian melts are generated in the deepest part of the column at the highest temperatures. With increasing degree of melting, at lower temperatures and pressures, the primary melts become less magnesian but at the same time have higher Mg# and coexist with more forsteritic olivine. Hence, melts with high Mg# need not be generated at extremely high temperatures. The presence of volatiles also lessens the requirement for high temperatures as the solidus of mantle lherzolite is lowered 25-60°C for each wt% H₂O in the melt and about 10°C for each wt% CO₂ in the melt. The most magnesian volcanic glasses found in Iceland and Hawaii provide evidence for magmas generated in dissimilar parts of melting columns. The most magnesian basalt glasses from Iceland, so far published, contain up to 10.5 wt% MgO, and are found in several picrite formations on the Reykjanes Peninsula and in the northern volcanic zone. Generally, the Icelandic picrite formations appear to date back to the glacioisostatic period at the end of the last ice age. In terms of major and trace element compositions, these picrites are the most depleted volcanics found in Iceland. Furthermore, they show LREE depletion, depleted ¹⁴³Nd/¹⁴⁴Nd, and low concentrations of radiogenic Pb. The associated high-MgO glasses have unusually low concentrations of K₂O and P₂O₅ (generally ≤0.05 wt%) and the TiO₂ content is generally about 1 wt% or less. When all of these characteristics are taken into consideration, it is clear that this kind of high-MgO melt cannot be parental to most of the less magnesian basaltic melts found in Iceland. Instead, we propose that the high-MgO melts are generated at the top of the melting regime under Iceland at a pressure close to 1 GPa, approximately at the low-pressure extreme of the transition from spinel to plagioclase lherzolite, after segregation of the more fusible part of the upwelling mantle column. By assuming that these melts are close to being primary products of the melting of lherzolite (or harzburgite with some clinopyroxene and spinel and/or plagioclase remaining), use of the CMASNF geothermometer yields a relatively modest temperature of up to 1280°C. The most magnesian glasses from Hawaii, with up to 15 wt% MgO, are found in turbidite sands at the base of Puna Ridge, the submarine extension of Kilauea volcano. In contrast to the primitive Icelandic glasses, the Hawaiian picrite glasses are derived from a relatively undepleted source. For example, the most magnesian glasses contain about 0.3 wt% K₂O, 0.2 wt% P₂O₅, and 1.9 wt% TiO₂. Unlike the Icelandic high-MgO basalt glasses that lie close to the olivine-plagioclase saturation boundary, the Hawaiian picrite glasses clearly align along olivine fractionation lines. On the basis of the CMASNF geothermometer, Gudfinnsson and Presnall (2002) argued that the glasses with about 15 wt% MgO could be generated from mantle lherzolite with potential temperature of about 1420°C at a pressure of about 2.5 GPa. Thus, we propose that the Hawaiian picrite melts are generated in the deeper levels of the melting column where only a relatively small amount of melting and melt segregation has occurred.

Global departure from equilibrium in a self-gravitating system and global tectonics.

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Many dynamical processes take place in our planet Earth because of some departure from the equilibrium state to which it spontaneously tends as a system subjected to its own gravity field. A sound recognition of this equilibrium state is an essential requirement in order to interpret correctly the already available evidence of processes that occurred in the past and/or now in progress, and to focus on the most appropriate investigations to undertake in the future.

Taking into account only the gravitational field, one should expect that, in the absence of 'external' actions, the matter making up the self-gravitating system Earth would tend spontaneously to be arranged in a disordered way along every equipotential surface, and in an orderly way in directions normal to these surfaces. The disordered arrangement along the potential surfaces would give rise to lateral statistical homogeneity and isotropy; the radial order would consist in the stacking of less dense over denser matter. It is worth pointing out that the departures from *lateral homogeneity*, from *local isotropy* and from *equilibrium density gradient* are strictly interlaced. In fact, the arising of each one of these three departures necessarily implies that of the other two.

Unfortunately, the current state of the art forces us to deal with the above departures from equilibrium as they were unrelated.

A departure from lateral homogeneity can be, in a way, considered as a departure from thermodynamic equilibrium. After all, the tendency to lateral homogeneity - extendable to the third dimension when dealing with not too wide systems in which the slight difference between "up" and "down" can be disregarded - led to the conception of thermodynamics. Anyhow, since in this theory the gravitational field is considered as external, the spontaneous tendency to lateral homogeneity of composition is limited to mixable phases.

A departure from the equilibrium pressure and density gradients in the gravity field, can be explained, without thermodynamic argumentations, as a departure from hydrostatic equilibrium.

Finally, it goes often unnoticed that a departure from local isotropy along equipotential surfaces - that is, that departure from isotropy a volume element included between two close equipotential surfaces shows being in a non-hydrostatic state of stress and/or having acquired strain anisotropy - is a departure from equilibrium. Notwithstanding, this anisotropy gives rise to spontaneous processes tending to restore isotropy of both state of stress and structure. Thermodynamics and hydrostatics cannot explain such a phenomenon.

The disjointedness of the current way of perceiving and explaining different facets of a single global physical phenomenon, certainly does not help in investigating causes and development of a process such as global geodynamics. To help in reducing the unavoidable groping, comprehensive *thermo-gravitational* equilibrium conditions are here defined, without raising any doubt about the validity of the conventional *thermodynamic* and *hydrostatic* equilibrium conditions, being the tendency to the latter ones, together with the tendency to *isotropy of structure*, expression of the spontaneous tendency to *lateral* disorder and *radial* order, jointly imposed by the existence of the gravity field. To define such a global equilibrium condition, one has just to extend to the whole self-gravitating system Earth what may be easily observed in its atmosphere: namely, *the spontaneous tendency to homogenisation of entropy per unit mass*.

All the thermodynamics principles should be considered valid, as they have been stated, *only* in systems included between any two close equipotential surfaces of the gravity field. Anyhow, in a

whole system subjected to its own gravity field, or in any isolated large fraction of it, an equilibrium thermal gradient must be associated to the equilibrium pressure and density gradients.

More than once, the existence of an equilibrium thermal gradient has been hypothesized by someone to justify the thermal gradient observed in the atmosphere and then discarded by some other one as incompatible with the validity of the second law of thermodynamics, or has been hypothesized by someone else to invalidate the second law and then discarded by the opponent on duty who gave as granted the validity of this law. However, relativistic thermodynamics seems to enforce in equilibrium an inhomogeneous temperature distribution in the atmosphere.

Unanimity may be reached just by considering that at least part of any increase (or decrease) of gravitational potential energy should correspond to a decrease (or increase) of heat. Since, considering the gravitational potential energy as part of the internal energy, one has to recognize that a self-gravitating system spontaneously tends to lateral homogeneity and radial inhomogeneity of temperature, the second law should be restated for a whole self-gravitating system in the following way: heat does not pass from one body to another if such a passage would destroy a state of equilibrium.

The assumption that heat supply at depth should be responsible for convection in the lower mantle is basically supported by the observed existence of the geothermal gradient that, as a whole, is currently considered expression of a departure from equilibrium. Besides looking for further observational and experimental verification of the tendency to *thermo-gravitational* equilibrium, it would be worth trying to check if all the already available evidence can be considered in agreement with the above suggested working hypothesis. Such a trial could lead to a new way of conceiving and evaluating what is currently considered evidence of heat flow in the mantle and the crust, perhaps allowing a distinction between the departure from this gradient responsible for the overall mass transfer we call global tectonics, and the local ones, responsible for heat flows stemming out from dissipative processes and giving rise to shallow secondary mass transfer. Some of the local departures from thermal equilibrium, resulting from the ignored or undervaluated heat release during the spontaneous isotropization of previously strained rock masses, should be taken into account besides the ones resulting from heat released during deformation.

P. S. - A book on this subject has been published in Italian (www.geocities.com/pino_guzzetta). For a free copy of the whole book in English please write to guzzetta@unina.it.

Convection in a self-gravitating system

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In order to inquire on the role of convection in global tectonics, one has to take into account that any action exerted from the exterior on a self-gravitating system, or on any fraction of it, produces, all together, departures from lateral homogeneity, local isotropy and radial gradient of density⁽¹⁾. As a consequence, mass and/or energy transfer spontaneously takes place tending to reduce the above departures.

The beginning of mass transfer known as ‘convection’ has been observed in systems consisting of one or more fluids subjected to gravity, when introduction of heat makes laterally not homogeneous their temperature, leading to departure from lateral homogeneity of density⁽²⁾ and from local lateral isotropy (i. e., local equilibrium hydrostatic pressure). Convection tends to reduce these departures.

Within the bulk of low viscosity fluids, the departure from equilibrium hydrostatic pressure is so slight that the lateral non uniformity of density and the departure from the equilibrium radial density gradient can be considered as the only driving forces responsible for mass transfer. In a horizontal liquid layer, the warmer, less dense fluid moves to overlap the colder, more dense fluid. If the lateral gradient persists in time, the mass transfer shows a cyclic course (formation of Bénard convection cells), with the warmer fluid going upward to the top of the layer, and the colder fluid going downward. As a consequence, temperature at the top free surface loses its uniformity making the surface tension not homogeneous.

If the lack of homogeneity of the surface tension is obtained heating not uniformly the liquid layer from the top, the ensuing departure from local isotropy becomes relevant, so that mass transfer tending to reduce such a departure may prevail: especially in micro-gravity conditions. If the lateral temperature gradient is maintained in time, it becomes cause of convective mass transfer (formation of Marangoni convection cells), with the colder fluid going upward and the warmer fluid going downward.

The outer shell of the ‘solid’ Earth is more complex than the top surface of a liquid layer, but it seems to behave, as a whole, like a very high viscous fluid floating over less viscous matter. Lateral unevenness of temperature may explain the not homogeneous state of stress of the crust and the consequent spontaneous mass transfer tending to reduce the local departures from equilibrium hydrostatic pressure. Most of the mass transfer consists in the wedging of underlying less viscous matter, tending to reduce local deficiency of lateral pressure, and in the subduction of portion of the overlaying more viscous shell, tending to reduce local excess of lateral pressure. The mass transfer within the outer shell, as shown by its deformation, is subordinate.

The persistency of the mass transfer suggests that the latter is following a cyclic course and, consequently, that the action ‘from the exterior’ on the Earth crust persists in time.

To know if the state of stress is cause of, or is caused by convention - that is, if the system is open at its top, or at its bottom - we have first to locate the energy source needed to keep in existence the departure from equilibrium.

(1) - See the abstract “Global departure from equilibrium in a self-gravitating system and global tectonics”.

(2) - In a vertical liquid layer heated from one or both sides, convection starts as soon as temperature becomes laterally not homogeneous. In a liquid layer heated from below, convection starts when - because of the uneven heating, or because of the unavoidable local fluctuations of temperature and density if the heating is homogeneous in the average - one or more hotter spots whose density is lower than that of the adjoining regions come in existence at the layer bottom.

AN ALTERNATIVE VENUS – PLUME-FREE PLANET PRESERVES PRE-3.9 Ga ACCRETIONARY SURFACE

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The model for Venus accepted by most specialists assumes a radioactive composition similar to that assumed (wrongly?) in the standard model for the Earth. As Venus lacks bimodal topography and plate tectonics, common corollary speculation is that Venus sheds its assumed excess heat primarily by rise from deep mantle of large and small plumes and upwellings that spawn myriad plumelets and diapirs. In most variants of this general scheme, the rising material represents a mantle overturn that resurfaced the planet structurally and magmatically and that predated only bolide impacts younger than 1 or 0.5 Ga. The rising masses are manifested at the surface mostly by circular rimmed and multiring structures up to 2000 km in inner diameter, and by vast basalt-surfaced lowlands. Among hundreds of reports embellishing variants of such assumptions are Aittola and Kostama (2002), Basilevsky and Head (1998), Brown and Grimm (1999), and Smrekar and Stofan (1999).

The assumptions on which these rationales are based may be invalid. Venusian plume scenarios are imaginative extrapolations from conjectures, likely false, regarding existence of such structures on Earth (see last paragraph). Speculation that the lower mantles of Earth and Venus are fertile and unfractionated—prerequisites for whole-mantle convection and for plumes from deep mantle—descends from 1950s conjecture (e.g., Urey, 1951), reasonable then but not now, that these planets accreted cold and slowly from enriched material like that of meteorites from beyond Mars, then heated gradually and are still largely unfractionated. Cosmological, orbital, and other considerations require that the inner planets instead accreted fast, violently, and hot, from material much less volatile than those meteorites, and fractionated early and irreversibly (references in Anderson, 2002, and Hamilton, 2002, 2003). Observed heliocentric compositional zoning of the asteroids (coming sunward: ices + organics; carbonaceous chondrites and other volatile-rich compositions; ordinary chondrites etc.) continues through progressively-less-volatile Mars and Earth and, arguably, on through Venus. Direct indicators of a less volatile composition of Venus than Earth include Venus' probably lower uncompressed bulk density (more magnesian), lack of silicic crust as sampled by 7 landers, and its very low atmospheric ^{40}Ar (low planetary K_2O ?). Earth itself is much less radioactive than commonly assumed, and has only about 2/3 the heat flow (Hofmeister and Criss, this symposium).

Venus commonly is postulated to have been magmatically resurfaced, and modified by extension and shortening, before ~1000 impact craters with little-modified ejecta blankets were blasted into it. These obvious impact craters have diameters <270 km, and mostly >50 km although only ~10 are >100 km, and are distributed randomly, or nearly so, on other units. The dense atmosphere retards all bolides; nearly all small bolides, and a great many mid-sized ones, are destroyed in transit. Conventional calculations of the maximum age of pristine craters are based on estimation of size distribution of the few bolides that produced the >100-km craters, with the critical

assumptions that these craters represent all of the large impactors that reached the atmosphere and that these large bolides neither fragmented nor lost velocity in transit. A maximum age of ~800 Ma is inferred on this basis (McKinnon et al., 1997). This age limit likely overestimates atmospheric survival of large bolides, and it does not incorporate either atmospheric slowing or scaling of crater mechanics for the dense atmosphere, both of which markedly decrease crater dimensions for a given bolide. Integration of these neglected factors permits much of the pre-existing surface of Venus to be 3 or 4 Ga old (Schultz, 1993). The conventional young-age calculation assumes further that all young impact structures are visible and thus overlooks the possibility that the largest impacts (transient craters, say, >400 km diameter and >100 km deep) might have generated enough decompression melt, in addition to impact melt, to have buried the craters and produced large volcanic constructs (cf. Jones et al., 2002). In conventional explanations, Venusian uplands are regarded as saturated with products of pre-impact plumes, plumelets, and diapirs, and lowlands as vast fields of pre-impacts basalt.

A proposed alternative Venus is now internally much cooler than Earth (though graded to hot greenhouse surface temperature): its initial heat was all or mostly lost long ago, and its radiogenic replenishment is much less. Much of the surface of Venus has been little modified since main accretion ended -3.9 Ga ago. Uplands are saturated with small to huge ancient impact craters (not with young magmatic edifices), from which rubble was eroded and deposited in a transient ocean (unrecognized in consensus interpretations) in nonvolcanic lowlands whose floors also are saturated with impact structures.

The pre-pristine-craters upland surface of Venus is dominated by rimmed circular structures, variably superimposed, many of them multi-ringed, of widely varying degrees of preservation and interference. I see these as erosion-modified impact craters that had varying amounts and patterns of impact-related melt. (My first statement of this view was in Hamilton, 1993.) Well-defined inner rims range from 50 to 2000 km in diameter, and outer margins, beyond troughs and outer swells, reach 3000 km. The majority view of specialists is that these circles (most “coronae” and “arachnoids”, and many “novae” and “paterae”, in Venusian jargon), throughout their huge range of size, formed atop plumes and plumelets that spread, sublithospherically in some schemes and extrusively in others, into circular shapes no matter where, or against what, they formed. This widely-accepted conjecture lacks discussion of bases or alternatives, or of such critical problems as lack of the lobate shapes required by the speculation.

To me, large rimmed circles on a solid-surface planet require gigantic explosions, and only impacts are plausible sources. (Earthbound geologists went through this in the 1960s and 1970s, when circular “crypto-volcanic structures” were proved to be ancient impact craters.) Volcanic explosions are trivial by comparison; indeed, the obvious volcanic constructs on Venus have, at most, small calderas. Many of the older large Venusian circles are moderately deformed, and stresses, and their relaxation, related to impacts may account for much of the modest early deformation of the planet’s surface. Correlation of gravity with topography indicates great strength—thick lithosphere—in outer Venus. (Contrary inference that topography is supported dynamically by rising plumes or cells, that lithosphere is thin, and that high upper mantle is near solidus temperature, is circular rationalization.) The young nearly-pristine impact structures retain obvious ejecta blankets that distinguish them from older rimmed circles of similar to larger

size, assigned to plumes. Inner rims of large Venusian circles, as exposed in uplands, are eroded to bedrock, so particulate ejecta have been removed from rims, although perhaps only smoothed on outer slopes. In the broad lowland plains, hundreds of small to large buried impact craters appear as rimmed depressions (“ghost coronae”) where plains surficial material is compressed into them, and huge circles (mostly-buried impact maria?) may be defined by partly-exposed rims, to ~4000 km in diameter. This printing-through of subjacent structures shows plains material to be weak and thin.

Lunar analogy indicates giant impact craters to be older than 3.9 Ga. Venusian uplands, like the lunar highlands, are saturated with craters that include such great terminal impacts of main accretion. The surface of Venus is an eroded accretionary landscape mostly older than 3.9 Ga. The weak Venusian heat engine has been incapable of much modifying the surface throughout most of geologic time.

The resurfacing prior to formation of little-modified impact craters was not planet-wide magmatism, but instead was a brief era of erosion that removed and smoothed ejecta blankets in the uplands, and of mostly-marine sedimentation that buried impact structures in the lowlands (Jones and Pickering, 2003). Timing is constrained primarily to be younger than 3.9 Ga, but the aqueous era was brief—subaerial valley systems are poorly integrated (Baker et al., 1997), and marine deposits are thin. (Subsequent eolian erosion and deposition have probably been minor: Greeley et al., 1997.) The atmospheric deuterium/hydrogen ratio, ~150_ that of Earth, may indicate that Venus once had an ocean equivalent to ~270 bars of water (Hunten, 2002). Erosional river valleys enter the lowland plains, into which channels (“canali”) extend, one possibly 6800 km long but all others shorter than 500 km. The channels were made by debris-carrying fluid and some have cutoff meanders, braids, point bars, and deltas (Williams-Jones et al., 1998). Jones and Pickering (2003) recognized that the uniform-cross-section lowland channels dimensionally and morphologically resemble terrestrial submarine density-current channels, and hence that the lowlands likely are formed mostly of turbidites. (See Habgood et al., 2003, for further description of terrestrial analogs.) A transient early ocean apparently was present, perhaps more than once and perhaps hot, and disappeared as the atmosphere heated because of increasing solar radiation and increasing greenhouse gas. Mass, temperature, and composition of the Venusian greenhouse atmosphere—now 93 bars, surface T ~475EC, ~96.5% CO₂, 3.5% N₂, traces of other gases—must have changed greatly over time, importantly by loss of water by oxidation of CO to CO₂ and escape of H₂.

Radar properties of the gentle and mostly radar-dark Venusian plains accord with sedimentary character (Cochrane and Ghail, 2002). Soviet landers showed plains materials to be soft, porous, low in density, and horizontally layered—all consistent with sedimentary deposits—and vaguely basaltic in semiquantitative partial chemical composition (e.g., Basilevsky et al., 1985). Nearly all Venusian geologists, however, assume that liquid water cannot have existed during the planet’s geologically recorded history, so they devise ad hoc explanations, devoid of actualistic analogs, of the plains and channels in terms of volcanism. For example, the superabundant small, low shields, mostly 0.1-5 km in diameter, that speckle many lowland regions are termed, without evaluation, basalt volcanoes. Their distribution makes no sense in this context, and I presume them to be mud volcanoes.

The plains sediments and channels predate broad, open deformation (Stewart and Head, 2000). Smaller-scale surficial deformation of the plains—wrinkle ridges, and gridded and polygonal patterns—is likely a product of secular changes in post-oceanic atmospheric temperature (e.g., Anderson and Smrekar, 1999), and not of global tectonics or volcanism (cf. Banerdt et al., 1997). Relative fluid pressurization responsible for mud volcanoes presumably also relates to atmospheric change.

The surface of Venus may have been modified since late in the era of main planetary accretion primarily by early aqueous erosion and deposition. Internally-driven tectonism and volcanism have been minor. The change, likely before 3 Ga, from aqueous modification to pristine preservation of impact craters, represents thresholds in evolution of atmosphere and hydrosphere, and not an internal turnover of the planet.

Widespread acceptance of feebly-based conjecture that plumes from deep mantle profoundly influence evolution of Earth's crust and upper mantle has retarded consideration of alternatives (Anderson, 2000; Hamilton, 2002, 2003; various authors in this symposium). Terrestrial plumology is based on bad assumptions of planetary composition and evolution. Geophysical rationales and purported evidence do not withstand scrutiny and improved data, but as fast as their plume predictions are falsified, proponents make their conjectures more convoluted, untestable, and unique to each example. After early speculation that earthly plumes are fixed in the mantle was disproved by plate-kinematic and paleomagnetic data, and after it was learned that ridge and island basalts largely overlap in composition, evidence-evading salvage notions of gyrating and squirting plumes were made ever wilder. Pro-plume geochemical rationales display unawareness of thermodynamics, phase petrology, and much more. Export of dubious terrestrial plume conjectures, with unconstrained modifications, retards comprehension of very-different Venus.

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Alternative Mechanisms for Volcanic Activity in Hotspot-Ridge Systems: The Northern Galapagos Province

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Many geologic features of the Galápagos Archipelago conform to the Hawaiian-style hotspot model (Morgan, 1971). Most importantly, the islands near the central axis of the chain get older to the east as predicted from Nazca plate motion, and recent volcanic activity is focused in the western archipelago. Upon closer inspection, however, the Galápagos deviate from the simple hotspot model: (1) the islands are broadly distributed instead of forming a linear chain in the direction of plate motion; (2) most of the volcanoes are active (e.g., White et al., 1993); (3) geochemical variations define complex spatial patterns, reflecting contributions from multiple mantle reservoirs (Geist et al., 1988; White et al., 1993; Kurz and Geist, 1999; Harpp and White, 2001); and (4) the hotspot may be interacting extensively with the adjacent Galápagos spreading center (e.g., Schilling et al., 1982).

Many of the deviations from the conventional hotspot model are most prominently expressed by the enigmatic volcanic features that grace the northern perimeter of the archipelago, including Genovesa, Marchena, and Pinta Islands, the northeast platform seamounts, and a bathymetric lineament including Wolf and Darwin islands and several elongate seamounts (Figure 1). In general, the northern volcanoes have been challenging to incorporate into regional models for the plume (e.g., Harpp and White, 2001). The present-day hotspot is thought to be centered near Fernandina Island (Kurz and Geist, 1999), and absolute motion of the Nazca plate is essentially eastward (Gripp and Gordon, 1990). Therefore, the northern volcanoes do not lie “downstream” from the plume and cannot be explained by the traditional mantle plume hypothesis.

Lavas erupted on the northern islands exhibit a wide range of chemical compositions, from enriched (Pinta) to depleted (northeast platform seamounts), spanning nearly the entire range observed in the archipelago (White et al., 1993). The northern islands are some of the smallest volcanic centers in the Galápagos, exhibiting little resemblance to the large western shields. Many are morphologically unusual, and possess elongate, composite structures rather than symmetrical, conical shapes (Christie et al., 1992). Perhaps most importantly, the northern volcanoes are located between the main archipelago and the Galápagos spreading center, one of the few places in the world where two fundamental mantle processes operate in such proximity.

We present results from a series of detailed field studies of Northern Galápagos volcanic centers, including Genovesa Island, a submarine ridge that extends from Genovesa, and the Wolf-Darwin Lineament.

A. Genovesa Island and Genovesa Ridge

Genovesa Island, which lies between the central Galápagos Platform and the Galápagos Spreading Center, is unusual in several respects. Exposed Genovesa lavas are the most isotopically depleted intraplate ocean island basalts on the planet (Georoc database). Despite its circular coastline and calderas, it is crosscut by both eruptive and non-eruptive fissures trending NE-SW. The 075° bearing of the fissures parallels that of Genovesa Ridge, a 55 km-long volcanic rift zone that is the most prominent submarine rift in the Galápagos and constitutes the majority of the volume of the Genovesa magmatic complex (Figure 1). Genovesa Ridge was the focus of detailed multibeam and side-scan sonar surveys during the Revelle/Drift04 cruise in 2001. The ridge consists of three left-stepping en echelon segments; the abundances of lava flows, volcanic terraces, and eruptive cones are all consistent with constructive volcanic processes. The non-linear arrangement of eruptive vents and the ridge's en echelon structure indicate that it did not form over a single dike.

Major and trace element compositions of Genovesa Ridge glasses are modeled by fractional crystallization along the same liquid line of descent as the island lavas, but some of the glasses are more primitive than any lavas sampled from the island. Most lavas on both the ridge and island have accumulated plagioclase. Incompatible trace element abundances of dredged Genovesa Ridge rocks are lower than the island's lavas, but ratios of the elements are similar in the two settings. The geochemical similarities indicate that the island and ridge lavas have nearly identical mantle sources. Glass inclusions in plagioclase phenocrysts from the ridge are compositionally diverse, with both higher and lower MgO than the matrix glass, suggesting homogenization at shallow levels.

The structural and geochemical observations are best reconciled if Genovesa Ridge did not form in response to injection of magma laterally from a hotspot-supplied central volcano, which is how Kilauea's Puna Ridge is thought to operate. Instead, Genovesa Ridge and its western extension appear to be the result of passive upwelling directed by far-field tectonic stresses that are generated by tension across the 91°W transform. The proximity of the hotspot system causes magmatism in the extensional zones where it would not ordinarily occur.

B. The Wolf-Darwin Lineament

The Wolf-Darwin Lineament (WDL) lies between the focus of the Galápagos hotspot and the Galápagos Spreading Center, extending to the SE from its intersection with the ridge (Figure 1). Consequently, most researchers have attributed its origin to interaction between the plume and the adjacent ridge. We propose that the WDL is caused only partially by plume-ridge interaction, and instead that it is primarily the result of tensional stresses emanating from the inside corner of the transform fault at 91°W . An additional factor that amplifies the tension in this region is the oblique orientation of the major transform fault with respect to the Nazca plate's spreading direction. This setting creates a transtensional zone whereby strain is partitioned into strike-slip motion along the transform and extension throughout the inside corner of the ridge-transform system (Figure 2). The area under tension is magmatic owing to the overlapping effects of the ridge and the Galápagos plume.

The extensional model predicts no age progressive volcanism, which is supported by observed age relationships. The WDL volcanoes define two distinct chemical groups; lavas erupted south of Wolf Island have compositions similar to those produced along the GSC west of 93°W, while those from the northern WDL resemble GSC lavas from the segment directly north of the lineament. This geographic distribution implies that the WDL is supplied by the same type of hotspot-affected mantle as the segment of the GSC that produced the lithosphere underlying the volcanoes. The observed WDL geochemical gradients are consistent with the extension model; the region under tension simply taps hybrid products of mixing at the margins of the sub-ridge convection system and the periphery of the hotspot.

The results of our Genovesa Ridge and WDL studies, in essence, are pointing toward a single, coherent explanation for the distribution, orientation, eruptive behavior, and compositional variations of all the Northern Galápagos volcanoes. These volcanic centers may be explained not as the direct result of hotspot activity, but instead as the serendipitous product of the proximity of a plume hotspot to the major transform that offsets the GSC at 91°W. The adjacent plume and mid-ocean ridge magmatic systems serve to generate widespread melting throughout the northern Galápagos region by causing upwelling in the mantle where it might not otherwise occur. The melts then migrate to the surface along tectonically controlled zones of tension imposed by an extensional transform (Figure 2). In effect, then, the adjacent plume illuminates the regional stress field with magma; at mid-ocean ridge systems without a proximate plume, such zones are not rendered observable. The range of lava compositions erupted throughout the Northern Galápagos, then, reflects the heterogeneous nature of the underlying mantle and provides a rare opportunity to assess the extent and distribution of the Galápagos plume as well as its interaction with the nearby ridge.

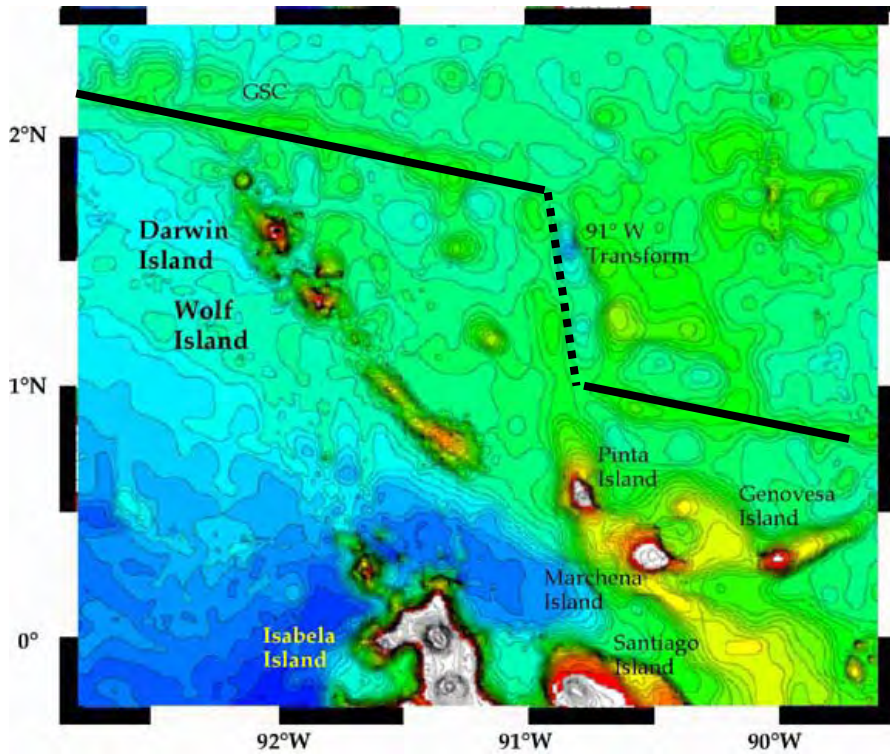


Figure 1. Bathymetric map of the northern Galápagos region, as compiled by Dr. William Chadwick of NOAA and Oregon State University and available at <http://newport.pmel.noaa.gov:80/~chadwick/Galápagos.html>. Solid and dashed black line is the Galapagos Spreading Center (GSC).

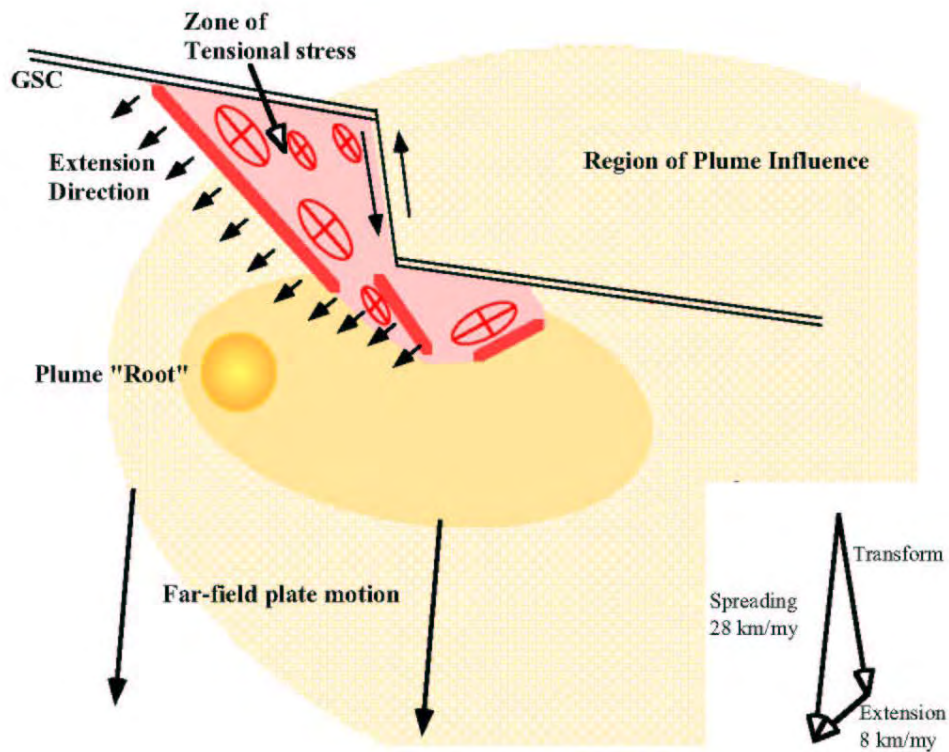


Figure 2. Conceptual model of the tectonic origin of the WDL. Orange tones indicate zone of influence of the Galápagos plume. Long arrows at the bottom of the figure oriented 186° show spreading direction of the Nazca plate. Short arrows represent extension necessary to accommodate plate spreading after strike-slip component of transform fault is removed. Bold red lines are observed volcanic lineaments (interpreted from bathymetric map, Figure 1). Ellipses are from the model of Gudmundsson [1995] for stresses around a transform fault-ridge intersection. Inset shows vector representation in velocity space of a model for strain partitioning through a combination of transform and extensional motions.

Evidence for layered mantle convection: implications for lower mantle plumes

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Experimental and theoretical studies of convecting systems generally yield plumes of some sort. For the Earth, of particular interest is whether deep plumes reach the surface. A critical limit on the egress of lower mantle plumes is whether mantle convection is whole or layered. Numerical models are under-constrained: the system is multi-phase, with extreme pressure and temperature differences, an uncertain chemistry, and the behaviors of the relevant physical properties are incompletely known. Given this complexity, controversy exists concerning whole or layered mantle convection. The points numbered below set forth key results from existing literature, mainly from the perspective of mineral physics, and provide new information from relatively unexplored avenues, all of which indicate that mantle convection is layered.

(1) Available rock samples have invariably originated at shallow depths. Inclusions in diamonds have the deepest source, near 670 km¹. A plausible interpretation is that the circulation of the lower and upper mantles are separate, whereby significant amounts of material from the lower mantle should not pass through the transition zone (TZ).

(2) The steep velocity profiles found in the TZ cannot be explained by adiabatic gradients for the phases possibly stable in this region, but can be modeled as a gradual transformation of pyroxenes to majoritic garnets as depth increases². Their comparison to laboratory measurements of shear and bulk moduli of majoritic garnets suggests enrichment of Ca in the lower mantle, and thus chemical layering. Alternatively, the observed velocity profiles are compatible with a sub-adiabatic gradient, which could arise from heat flow from the lower mantle being impeded by the low thermal conductivity of majoritic garnet³. For this to happen, the proportion of garnet in the TZ must be larger than that provided by a pyrolitic upper mantle composition, again pointing to chemical layering. A sub-adiabatic gradient in the transition zone could lead to catastrophic overturns of this layer with the upper mantle. Mantle avalanche has recently been inferred from length of day perturbation in the Cretaceous⁴.

(3) Grain size (d) differences seem to exist between the upper and lower mantles. Rock samples from the upper mantle have variable d with megacrysts as large as 1 cm. Mantle viscosity has been used to estimate d as 0.3 cm near 670 km⁵, whereas grain growth modeling⁶ suggests d is roughly 0.01 to 0.1 cm for the lower mantle. Whole-scale mixing would not produce such differences. More importantly, two of the physical properties which exert key controls on mantle convection, viscosity⁷ and diffusive radiative thermal conductivity⁸, strongly depend on grain-size. The latter dependence results from the emission spectrum of a mineral grain and physical scattering depending on d , and the results support small grain size in the lower mantle. Importantly, heat transfer in the lower mantle is largely radiative, whereas in the upper mantle, phonon transfer dominates. Because the sign of dk/dT determines stability,⁹ the differences in mode of heat transport suggest layered convection.

(4) Dynamical models limit the secular cooling delay to 1-2 Ga, given favorable conditions for heat retention that include a low conductivity layer, low radiative transfer, and high initial heating.¹⁰ A 1 Ga delay requires that radioactive decay contributes at least 80% of the total power. The heat generation of K, U, and Th now and at 1 Ga constrains the bulk silicate earth (BSE) composition, irrespective of uncertainties in surface heat flux and on the secular delay¹¹. Our BSE composition differs from known compositions of the upper mantle suite¹², again indicating chemical differences between the upper and lower mantles.

(5) To understand the effect of layering heat producing elements, we model the Earth as an onion with thick shells. The heat flow through any given shell equals that of its internally generated heat, plus the heat produced in all deeper shells. Therefore, the most efficient means for the Earth to expel heat is if the radioactive elements are located in the shallow shells. The continental crust is the known, extreme example. Both chemical differentiation and the dynamics of heat flow lead to stratification of U, Th, and K within the mantle. The amount of radiogenic heating inside the mantle is also overstated, as this hinges on the current estimate of total power as 44 TW. However, this value was deduced assuming that the thermal conductivity be unrealistically constant.¹¹ Several independent approaches to analyzing surface flux provide a global power of 31 TW.¹¹

(6) A two-layer model of the convection pattern inside the Earth can be inferred from the surface expressions, i.e., the long-lived hot spots¹³ and the subducting plates. Most of the surface complexity comes from upper mantle and TZ phenomena. In a three-dimensional visualization, convection in the lower mantle consists of two hemispherical cells, with zones of up-welling centered on and underlying Hawaii and South Africa (see figure). Thus, these antipodal hot spots do not originate from plumes, but instead are edge effects for the simplest possible configuration for 3-D convective flow in the spherical cell representing the lower mantle.

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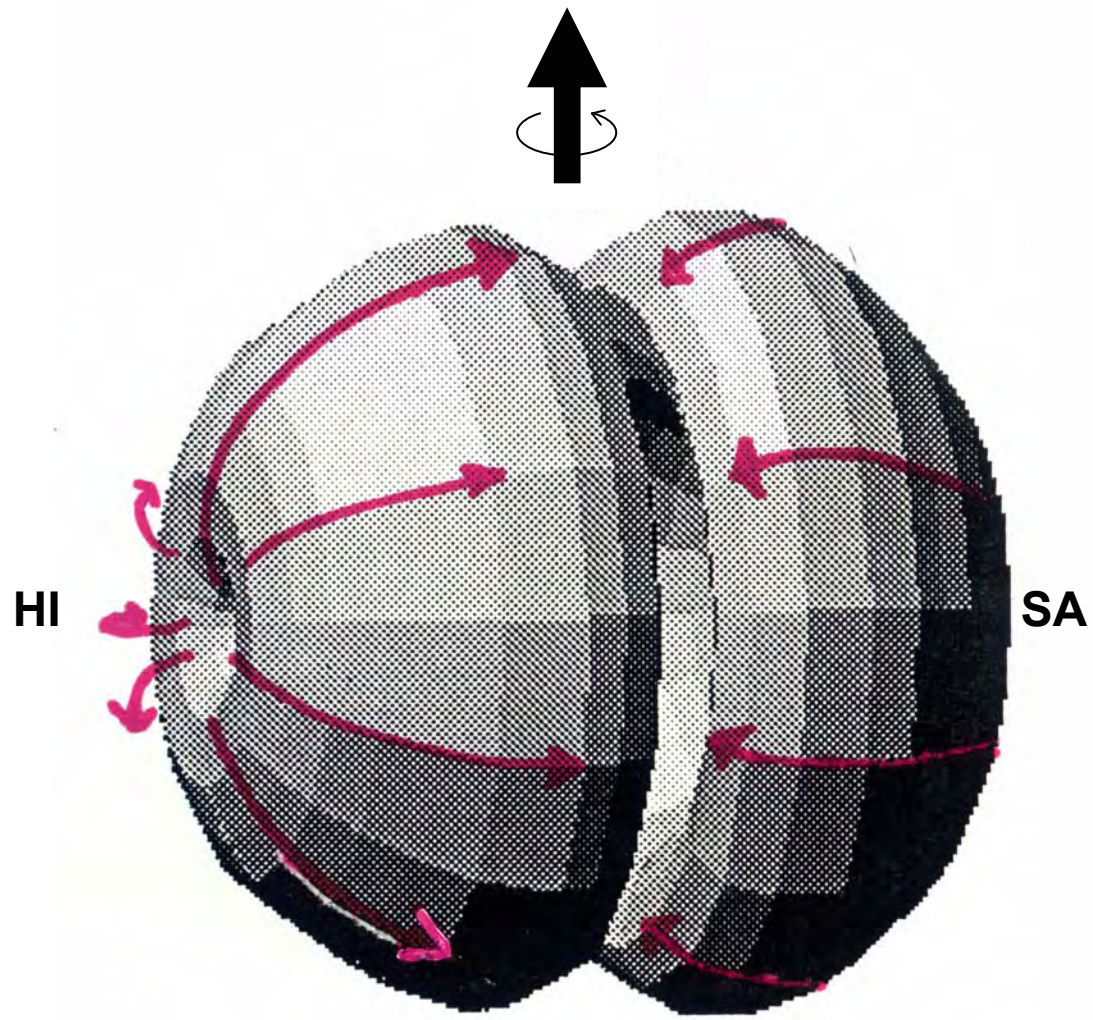
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A schematic comparison between Albanian Ophiolites with Idealized Ophiolite Sequence, Penrose Conference

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According to Penrose Conference participants (1972) an idealized ophiolite sequence includes from the bottom upward, tectonized ultramafic mantle rocks, layered ultramafic to mafic cumulate rocks, isotropic gabbros, a sheeted dyke complex and a massive volcanic section dominated by pillow lavas and overlain by pelagic sediments.

Albanian ophiolites extending over 4 000 km² or 1/7th of Albanian territory link together the Dinaric and Hellenic ophiolites. They form two NNW-SSE sub-parallel belts, *the western ophiolite belt* and the *eastern ophiolite belt*, representing the most complete and coherent sequence of the Mediterranean ophiolites (Beccaluva et al., 1994; Shallo et al., 1995).

The Western belt ophiolite - It consist of high-Ti tholeiites within volcano-sedimentary series in the western peripheral parts and a volcanic sequence in the western part associated with lherzolite mantle suite and minor harzburgite and dunite tectonite and ultramafic and mafic cumulates (Shallo et al.1985; Beccaluva et al., 1994).

Westermost part ophiolites of the Rubiku-Vela area -

The volcano-sedimentary series comprises high-Ti tholeiitic pillow lavas of MORB affinity (Shallo et al.1985; Beccaluva et al. 1994) with scarce intercalations of argillitic-silicious-sericitic of sericitic-siliceous-carbonate shales and slates, and infrequently, hematitic radiolarian chert. The series is about 600 thick and northwest trending. The radiolarian assemblages of the chert specimens, intercalated with massive sulfide ores, indicate a Middle-Late Carnian age, possibly including the Early Norian (Hoxha, 2001).

At Rubiku and its vicinities, agglomerate and argillitic-siliceous shales topped by yellow-reddish hematitic manganiferous radiolarian chert overlie basaltic pillow lavas. Volcano-sedimentary series are underlain by gabbro, troctolite, ferro-gabbro, very scarce minor microdiorite as well as ultramafic injections and "wedges", ranging in thickness from few meters up to 100 m. It must be emphasized that ultramafics in many cases are changed to serpentines.

The greenschist-amphibolite sole Bajocian-Bathonian in age is widespread in the western ophiolite belt; it is located along the volcano-sedimentary-ultramafic contact with thickness from a few meters up to 120 m, with 2 to 4 m amphibolites and 10-120 m greenschists.

Western part ophiolite of Gziky-Kachinar area - The lithologica-startigraphical section consists of volcanic sequence overlain by Middle Jurassic hematitic shales and is underlain by gabbro, very scarce plagiogranite and ultramafic rocks. The volcanic sequence consists of an Upper basaltic pyroclastic succession averaging 400 m and a Lower basaltic pillows (mainly) and massive succession, about 600 m thick. The enormous resources of vanadium-bearing titaniferous gabbros of the Kachinar area located in their uppermost part, at the base of volcanics, should be noted.

The Eastern belt ophiolites - This belt is characterized by low-Ti tholeiites of basalt-andesite and andesite-dacite (rhyolite) series underlain by sheeted dyke complex, and quartzdiorite-plagiogranite and gabbros (ISPGJ-IGJN, 1983; Beccaluva et al.1994; Shallo, 1994) extending from Kyafe Mali in the north to Perlati in the south, offering the best prospects regarding copper-pyrite-zinc and precious metals mineralizations.

Due to some differences between the northern and the southern parts they will be described separately.

Northern Kyafe Mali-Spach area - It comprises two volcanic successions the lower and the upper one.

The Middle-Late Jurassic lower andesite-basalt pillow lava succession - It consists mainly of pillows (spilite) and occasional massive flows, about 1 000 m thick. They have glassy rim and quenched glass filling interstitial cavities.

Dykes, extending over than 30 km, from Repsi to Kyafe Mali, several centimeters to 2-3 m thick, consist of dolerite, microdiorite, andesite, dacite or rhyodacite and rare boninite dykes (Beccaluva et al., 1994; Shallo, 1994).

The Late Jurassic upper andesitic basalt and dacite pyroclastic succession, about 700 m thick, consists mainly of pyroclastic rocks and occasionally pillows. This succession is followed by a 1 meter to 10 m thick manganese-bearing hematitic radiolarian chert sequence of Late Callovian to Early Oxfordian age (Marcucci et al., 1994).

Southern Perlat-Kurbnesch area

The rocks of this area are internally disrupted, due to intense thrust tectonics. Tectonostratigraphy of the area consist of the Middle-Late Jurassic Lower andesite basalt pillow lava and rare massive flow succession, about 500 m thick and Late Jurassic Upper andesite basalt pyroclastic succession, about 300 m thick overlain by argillaceous-siliceous succession with radiolarian chert.

The volcanics are underlain by sheeted dyke complex, and quartzdiorite-plagiogranite and gabbros.

The Penrose Conference definition remains from the greatest achivments in geosciences.

From the comparison of Albanian ophiolite sections with the Penrose Conference definition only the sheeted dyke dominated segment (Chafe Mali-Reps), widely accepted formed above a subducting plate margin, is similar with it, the other parts seems to represent various stages of oceanic crust formation.

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VARIATIONS IN THE TRACE ELEMENT SYSTEMATICS OF THE MAFIC ROCKS FROM THE
ARCHEAN BELLETERRE-ANGLIERS GREENSTONE BELT, SE SUPERIOR PROVINCE,
CANADA: A PRODUCT OF CONTAMINATION, SOURCE VARIATION, OR BOTH?

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The Belleterre – Angliers greenstone belt (BAGB) is located near the southern margin of the Archean-aged Pontiac Subprovince and is one of the southernmost greenstone belts of the Superior Province. It is located north of the Grenville Front, and south of the Abitibi Subprovince within the Timiskaming region of Québec. Previous investigations have suggested that the BAGB represents a tectonically-juxtaposed arc and oceanic plateau environment that was rifted as it passed over a mantle plume (Barnes et al., 1993). Subsequently the BAGB is interpreted to have been emplaced as a thrust sheet above a midcrustal-level duplex consisting of tonalites and metasediments of the Pontiac Group (Sawyer and Barnes, 1994). U-Pb (zircon) geochronological investigations of the volcanic rocks within the BAGB have returned ages ranging from 2712–2682 Ma (Mortensen and Card, 1994, this study).

Lithochemical analysis of the tholeiitic mafic volcanic rocks from the BAGB reveals that they can be divided into two distinct populations based on their trace element concentrations: those which are characterized by an enrichment in the light Rare Earth Elements (REE) as well as pronounced negative Nb and Zr anomalies when plotted on primitive mantle-normalized diagrams (LREE population), and those which possess flat REE patterns with weakly negative to positive Nb anomalies (FREE population). The mafic intrusive rocks of the BAGB can be similarly classified into these two populations based on the same lithochemical criteria. The mafic lavas and mafic intrusions of the respective populations are interpreted to be consanguineous equivalents based on their similar degrees of alteration, metamorphism, and structural deformation, as well as lithochemical and geochronological similarities. Spatially, these two lithochemical populations of mafic rocks occur in close proximity to one another and thus two broad questions are raised: 1) What process(es) led to the formation of these distinctly different magmas? and 2) What mechanism allowed for them to be intruded and erupted in such close proximity to one another?

Field evidence from the study area suggests that the FREE (N-MORB-like) mafic magmatism preceded the LREE magmatism and that the LREE magmatism is spatially associated with a large unit of felsic, quartz and feldspar-phyric sediments and flows. Thus it is possible that the LREE magmas may have been formed via contamination of the FREE magma through assimilation of this felsic unit. However, simple mixing models using these two components does not accurately reproduce the composition of the LREE magma, particularly the pronounced negative Nb and Zr anomalies. Therefore if the LREE magmas were formed by contamination, it does not appear as though the contaminant is exposed at the present crustal level. The distinct trace element nature of the LREE magmas, however, may be a source characteristic rather than a function of crustal contamination. For instance, Zr, Hf, and HREE may have become fractionated from the LREE if majorite garnet was present in the mantle residue from which these magmas were derived (Xie et al., 1993). Furthermore, mantle metasomatism within a sub-arc mantle wedge can produce island arc-type lithochemical signatures which are similar to those of the LREE magmas. Further work is required in order to evaluate which of the processes described above or a combination thereof contributed to the formation of the LREE magmas.

The mechanism which is invoked for the emplacement of the two populations of magma will be largely dependent on the process by which they are inferred to have formed, thus a number of scenarios are possible. Similar spatial relationships between N-MORB and island arc-type mafic lavas have been documented in the Hemlo-Schreiber greenstone belt of the Superior Province by Polat et al. (1998). There the association is attributed to the accretion of oceanic plateau fragments with N-MORB geochemical signatures to an arc causing arc-trench migration of the magmatic activity through the accretionary wedge (Kusky and Polat, 1999). This would allow for the intrusion and eruption of island arc-type mafic magmas proximal to the accreted oceanic plateau. The impingement of a mantle plume on an island arc offers another

possible explanation for the spatial association of these two magmas (Kerrich et al., 1999, Barnes et al., 1993).

Significant questions remain regarding both the possible role of contamination and source variation in the formation of the two magma populations and the mechanism by which these magmas were emplaced into the crust. However, the identification of their lithogeochemical variations represents an economically significant observation in that Ni-Cu-PGE sulphide deposits have been identified in a number of the mafic intrusions belonging to the LREE population whereas the FREE intrusions appear to be devoid of mineralization.

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Plume magmatism and mantle convection: Revising the standard model

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Despite its elegant simplicity and its agreement with important first-order observations of the Earth's surface, the mantle plume hypothesis has become a target of focused criticism and the source of considerable debate (Anderson, 2001; Foulger, 2002; Hamilton, 2002). Its lack of acceptance can be attributed in part to the inability of plume adherents to articulate the role of plumes in an integrated mantle convection scheme that satisfies available geophysical, geochemical, and petrologic observations. In fact, there is little consensus among earth scientists today regarding the pattern of mantle convection. While the combined efforts of geophysicists and geochemists have greatly advanced our understanding of the structure of the Earth, vigorous debate still surrounds the central issue of where and how the mantle moves. Plate tectonic theory has enjoyed enormous success in explaining the origin and evolution of the Earth's crust, yet the theory that describes motion on the Earth's surface offers few insights regarding motion beneath it. Even the most basic questions still lack a satisfying answer: does convection in the mantle consist of large cells that continuously transport material across the upper mantle/lower mantle boundary at ~670 km depth, or do the upper and lower mantles evolve as separate chemical reservoirs? Indeed, the nature of convection in the Earth's interior remains one of the most important unsolved problems toward understanding the structure and evolution of our planet.

The mantle plume hypothesis is attractive because it can account for the profound compositional heterogeneities (in stable and radiogenic isotopes as well as major and trace elements) observed between and among ridge-related and intraplate magmas in the oceanic setting (Hoffman and White, 1982; Weaver, 1991; Hart et al., 1992). The plume theory was supported by scaled tank experiments that elucidate the nature of rising buoyant material (Whitehead and Luther, 1975; Olson and Singer, 1985; Griffiths and Campbell, 1991). These experiments, in turn, have led to the prediction and subsequent identification of flood basalts as the expression of melting of plume starting heads (Richards et al., 1989). However, the simple concept of a continuous plume conduit supplying material to a point source over an extended interval of time cannot successfully be applied to many (if not most) localities of intraplate magmatism. Furthermore, tomographic studies have failed to image definitively a low velocity conduit at hypothesized hotspot localities. Perhaps most significantly, a simple compelling model that satisfies the observations of each of the subdisciplines focusing on mantle convection has not yet been presented. Thus, the plume hypothesis has fallen out of favor among a growing contingent of earth scientists (see www.mantleplumes.org).

At the heart of our inability to assemble a complete picture of mantle convection is our lack of quantitative data of the critical parameters (viscosity and density) that control flow in the mantle. The mantle is cooled from above (driven by conduction in the uppermost mantle and hydrothermal circulation in the crust) and heated from below (powered by latent heat of crystallization in the core) as well as from within (powered by radioactive decay throughout the mantle). Because of the strong dependence of density and viscosity on pressure, temperature, and composition, we have few constraints to guide our development of whole mantle convection

models. In the absence of a quantitative description of the variables that control the dynamic behavior of the Earth, we are forced to rely upon the models that best fit the available first-order observations.

Geophysicists and geochemists are generally divided in their view of mantle evolution, despite the rich record of observational constraints derived from a wide variety of techniques. Geophysicists have determined that some down-going slabs penetrate the 670 km discontinuity, and thus generally hold to the view that convection in the mantle is single-celled and involves circulation that spans the entire silicate interior. In contrast, geochemists cite isotopic evidence indicating that long-lived mantle reservoirs have remained isolated from one another throughout much of Earth history and argue that the upper mantle and lower mantle convect independently with little exchange of matter between them. Are there two chemically distinct mantles, or does Earth's interior consist of just one poorly stirred reservoir? Recent measurement of the time scales required for chemical diffusion at the great pressures within the mantle bolster the geophysicist's argument that giant, mantle-wide convection cells exist, but that large, chemically-distinct regions can remain incompletely mixed within a 'plum-pudding like' mantle. Geochemists counter this position by noting the consistent extraction of (presumably) *shallow* material with uniform composition at spreading ridges and by contrasting this with the composition of material extracted at intraplate point-sources of magmatism, (presumably) from *deeply*-derived, upwelling plumes.

Here, I posit an integrated model for mantle circulation that reconciles the seemingly opposing viewpoints of geophysicists and geochemists. This model incorporates a recent analysis of hotspot volcanism (the 'plumelet' model, Ihinger, 1995) and offers a new approach to understanding circulation in the upper mantle. The plumelet model, in conjunction with established geophysical and geochemical constraints, suggests that convection in the mantle consists of three circulation 'cells' that are not entirely independent of one another. In brief, the integrated circulation model considers four spatially and chemically distinct reservoirs: the continental crust, the upper mantle, the lower mantle, and the region between the outer core and the lower mantle known as the D" boundary layer. Upper mantle convection is driven by the sinking of cold slabs in subduction zones; the return flow to spreading ridges is accommodated by material flow within a diffuse *shallow* zone extending laterally for thousands of kilometers beneath the lithosphere (Figure 1). The diffuse zone of return flow begins tens of kilometers below the rigid lithosphere, and extends to depths marked by the boundary with the lower mantle (at ~670 km), although most of the flow is accommodated in the upper ~100 kilometers of the zone (Figure 2). Material is removed from the upper mantle reservoir when: (1) partial melts rise above subduction zones to form the continental crust; and (2) portions of the down-going limbs (the subducted slabs) of some upper mantle cells sink to the deep D" layer. The continual process of continental crust formation has served to deplete the upper mantle of incompatible elements throughout Earth evolution and give it a distinctive 'depleted' character. Note that in this model, the oceanic crust is considered part of the 'upper mantle' reservoir, and that chemical fractionation associated with the formation of mid-ocean ridge basalts (MORB) does not serve to "deplete" the reservoir. Some subducted slabs do not penetrate into the lower mantle and are cycled directly back into the upper mantle circulation. Other slabs are lost from the upper mantle and descend through the lower mantle to the poorly-stirred D" layer, where they reside for hundreds of millions of years before returning to the upper mantle in the form of upwelling

plumes (thus contributing, at least temporarily, to the depletion of the upper mantle; Figure 3). In contrast, the lower mantle reservoir evolves primarily as an independent, closed system. Material in this reservoir does not undergo chemical processing and has retained a chemical character similar to that of the primordial, undegassed early mantle. Some material exchange both into and out of the lower mantle reservoir has occurred throughout Earth history via thermal entrainment due to the cycling of the hot and cold limbs of the upper-mantle/D" convection system passing through it. This integrated model of mantle convection, composed of: (1) one 'cell' confined to the upper mantle; (2) one 'cell' that exchanges material between the upper mantle and the D" layer; and (3) one 'cell' confined to the unprocessed lower mantle, is consistent with and reconciles existing first-order geophysical and geochemical observations.

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VOLCANIC SYSTEMS AND SEGMENTATION OF THE PLATE BOUNDARY IN SW-ICELAND.

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The Mid Atlantic plate boundary breaks up into a complex series of rift- and transform zones as it passes the Iceland hotspot. In this presentation we define the segmentation of the plate boundary in SW-Iceland and the southwest insular shelf and describe the active volcanic systems, their petrochemistry and their connection to the structural elements of the zones. The Holocene volcanic eruption sites in Iceland have been grouped into volcanic systems, units that often possess common structural and petrochemical characteristics. Volcanic production is generally highest in the central region of each system and many of them have developed a central volcano with the associated production of intermediate and acid rocks. High-temperature hydrothermal activity is often connected to the central part of each system. The plate boundary in southwest Iceland and the southwest insular shelf divides into four segments of similar length, each of which has relatively homogeneous characteristics. These are the Northern Reykjanes Ridge, the Reykjanes Peninsula, the Langjökull Volcanic Rift Zone and the South Iceland Seismic Zone. What is labeled as the Western Volcanic Zone in this presentation comprises the Reykjanes Peninsula and the Langjökull Volcanic Rift Zone. The South Iceland Seismic Zone connects the Western and Eastern Volcanic Zones.

The crest of the Northern Reykjanes Ridge (62.90°-63.80°N) has an axis that is about 65 degrees oblique to the spreading direction. It is a narrow belt of low elongated en echelon ridges, superimposed on an elevated shelf. Rock dredges (altogether 45) indicate that the ridges are primarily made up of basaltic hyaloclastites, basalt pillow lavas are found at >250 m depth. The ridges are thought to be constructional in nature and possibly all represent postglacial submarine volcanism. If so, productivity has been similar in all the ridges and the ridges are comparable in lateral extent to the volcanic systems on the Reykjanes Peninsula. The elevated central part of the ridges is probably analogous to the elevated central portion of the volcanic systems on the Reykjanes Peninsula. The ridges are in many cases flanked by depressions but their connection to the volcanism can not be elaborated because of insufficient data. Ten submarine eruptions are known to have occurred on the crest of the Northern Reykjanes Ridge during historical times (1100 years) and two active high-temperature hydrothermal areas have been identified. We concur with previous suggestions that Recent volcanism on the Northern Reykjanes Ridge is restricted to eight submarine volcanic systems.

The Reykjanes Peninsula (63.80°-64.19°N) forms the transition between the Reykjanes Ridge and the volcanic zones of Iceland. It has an axis that is about 30° oblique to the spreading direction. Volcanism has been vigorous during Postglacial time and is characterized by small eruptions and low topographical relief. The eruption sites group into five volcanic systems which are arranged in an en echelon fashion within the volcanic zone. Only basaltic rocks appear to have been extruded during Postglacial time and the productivity of rocks has been similar in all the systems, at 0.1 km³/km². The westernmost system, the Reykjanes system, is partly submarine

and is not clearly separated from the next system to the east. The volcanic system Hengill, at the junction of the Reykjanes Peninsula, the Langjökull Volcanic Rift Zone and the South Iceland Seismic Zone, has developed a central volcano during late Pleistocene with small amounts of intermediate to acid volcanism. The small Grimsnes system, to the east of Hengill, could be added as the sixth system of this segment of the Western Volcanic Zone. Fissure swarms or rift zones are connected to all the volcanic systems. These are prominent swarms of rifting structures such as fissures, and normal faults, a few kilometers wide and a few tens of kilometers long, sometimes with the structure of a shallow graben. Five large hydrothermal areas are found on the Reykjanes Peninsula, they are all situated near the central region of the volcanic systems. It is suggested that the intensity of the hydrothermal activity on the Reykjanes Peninsula is connected with the high strike-slip component of rifting.

The Langjökull Volcanic Rift Zone (64.19°-65.22°N), has an axis which is 60° oblique to the spreading direction. The region is characterized by considerable relief and relatively few large eruptions. Only 32 Postglacial eruptions have been distinguished in this region, all basaltic, as compared to ≥ 180 eruptions on the Reykjanes Peninsula, including Grimsnes. The presence of distinct volcanic systems on basis of Postglacial volcanic activity is more difficult to pinpoint in the Langjökull Volcanic Rift Zone than in the before-mentioned segments of the Mid Atlantic Ridge. This is because of the low intensity of volcanic and intrusive activity. It appears that Recent volcanic activity (and also late-Pleistocene activity) is restricted to three active volcanic systems in the Langjökull Volcanic Rift Zone. The productivity during Postglacial time in the southernmost system is similar as on the Reykjanes Peninsula, but lower in the two northernmost systems. High-temperature hydrothermal activity is associated with two of the volcanic systems, Prestahnúkur and Hveravellir, which also have developed intermediate and acid rocks during Pleistocene time. The region contains several fissure swarms, some of which are connected to the suggested volcanic systems.

The South Iceland Seismic Zone is a nascent transform-type zone connecting the Western and Eastern Volcanic Zones. The zone is nearly parallel to the plate spreading direction and the general sense of motion is left lateral. Most of the larger earthquakes of this branch of the plate boundary occur on N-S striking faults that are arranged side-by-side, transversely to the zone, and are associated with right-lateral faulting. The left-lateral motion is therefore accommodated by right-lateral motion on the faults and slight rotation of the blocks between them. This type of faulting has been termed “bookshelf faulting”, and may be a characteristic of immature transform plate boundaries. Similar set of bookshelf faults has been identified on the Reykjanes Peninsula, overprinting the tectonic fabric of the volcanic systems there, suggesting an alternating mode of tectonism along this highly oblique plate boundary.

A geochemical survey has recently been completed on the basaltic rocks belonging to the Brunhes magnetic epoch (< 0.78 Ma) in the Northern Reykjanes Ridge and the Western Volcanic Zone. A pronounced discontinuity at the western tip of the Reykjanes Peninsula is indicated by the data set with regard to some of the major and trace elements. The overall major and trace element chemistry of the basaltic rocks of the Reykjanes Peninsula is, however, similar to that of the Langjökull Volcanic Rift Zone. The new geochemical data do not show any systematic variations in chemistry with time in any part of the Western Volcanic Zone or the Northern Reykjanes Ridge. The geochemical data do not indicate that individual volcanic systems, as

distinguished by the grouping of eruption sites, can be clearly separated by any major or trace element components, or combinations of these. On the other hand groups of two to four systems can be distinguished from neighboring groups. For example, the four westernmost systems on the Northern Reykjanes Ridge can be separated from the three others to the east on the ridge, and the three westernmost systems on the Reykjanes Peninsula can be separated from the three easternmost systems on that peninsula. The three suggested systems in the Langjökull Volcanic Rift Zone have produced basalts of very similar composition.

Magnetic anomalies at 1 km altitude within the Western Volcanic Zone are typically of the order of 1000 nT in amplitude. With respect to the geological structure of the volcanic zone, the most conspicuous positive anomalies are of two types i) Elongated highs of width around 10 km, sometimes coincident with fissure swarms and ridges of the volcanic systems and showing an echelon pattern. However, the highs do not always correlate with local topography, and their sources may well reach 2-3 km depth. The zone of magnetic highs is often narrower than the zone of Brunhes epoch surface outcrops, for instance in the Reykjanes Peninsula. ii) Localized highs over volcanic edifices such as lava shields, tuyas and hyaloclastite ridges with pillow-lava cores. The size of the latter anomalies presumably reflects both the amount of normal-polarity crystalline rock present and its magnetization which depends on cooling rate and other factors. Localized anomalies associated with roots of central-volcano complexes are quite common in older regions of Iceland, but few if any anomalies of this type seem to be present in the Western Volcanic Zone.

The Langjökull Volcanic Rift Zone is marked by an elongated Bouguer gravity low of a few mGals while no obvious trends can be connected to the Reykjanes Peninsula. Individual volcanic systems cannot be distinguished in the areas where gravity point spacing is sparse but there are indications in areas of denser data spacing that at least some systems are associated with gravity lows. These lows may be caused by accumulations of hyaloclastites in subglacial eruptions during glacial periods. Localized gravity highs are largely absent; such highs in association with large central volcanoes and calderas are common in other parts of the volcanic zones and older regions in Iceland. This suggests relative scarcity of crustal magma chambers in the Langjökull Volcanic Rift Zone and the Reykjanes Peninsula.

Temporal variation of Hawaiian plume composition: Evidence from Hana Ridge (Submarine Haleakala Volcano), Hawaii

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The mantle source generating the Hawaiian Islands is chemically heterogeneous from volcano to volcano as is well documented in numerous studies. Moreover, growing evidence suggests that the source also varies over time within a single volcano [Takahashi and Nakajima, 2002; Tanaka et al., 2002]. New data from the submarine shield stage of Haleakala volcano on Maui also suggest temporal variation of the

Hawaiian plume source and contribute to new ideas on the structure, geometry, and composition of the plume itself.

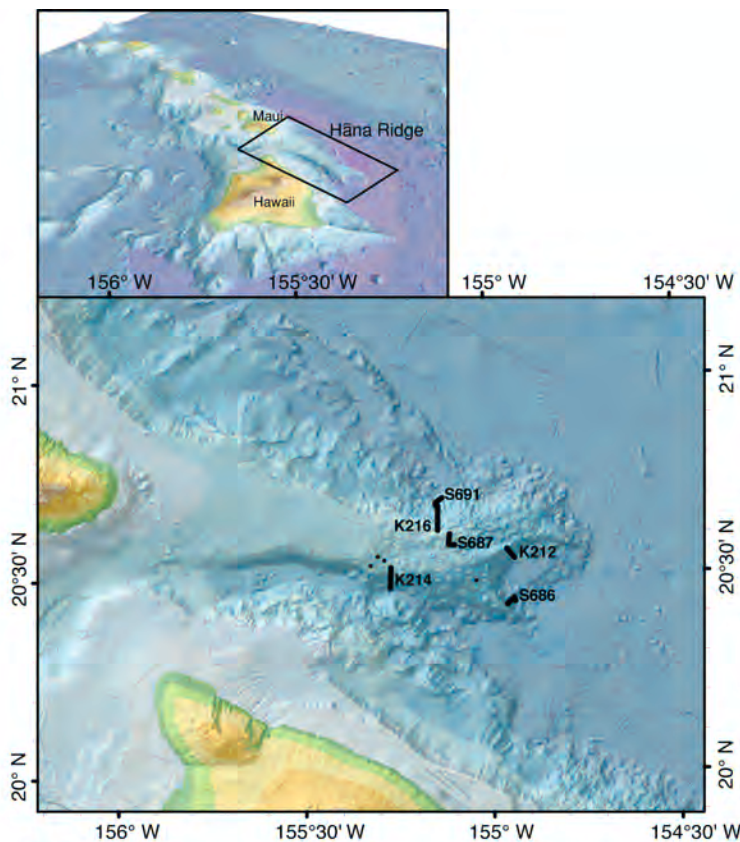


Figure 1. Perspective view of the Hawaiian Islands (top) showing the location and relative size of the Hana Ridge. Hana Ridge is shown (bottom) along with existing dredge sites (small dots; Moore et al., 1990) and JAMSTEC dive sites (Kxxx - ROV *Kaiko*, Sxxx - *Shinkai 6500*). Broad, smooth area uprift of sample sites is drowned coral reef.

Hana Ridge is the submarine portion of the east rift zone of Haleakala Volcano, Hawaii. At 140 km long, Hana Ridge is the longest submarine rift zone in the Hawaiian Island chain and has developed a complex morphology compared to other Hawaiian rift zones, such as Puna Ridge. The main ridge comprises three subparallel ridges related to distinct accretionary periods within the shield-building phase of Haleakala volcano. In order to investigate the

variability of Hawaiian plume compositions over time and the geochemical evolution of Haleakala shield-building, we sampled several sections of Hana Ridge on six dives with ROV *Kaiko* and *Shinkai 6500* submersible, both operated by JAMSTEC, in 2001 and 2002 [Johnson et al., 2002] (Figure 1).

All recovered rocks are tholeiitic basalts and more than half of them, those obtained in the deeper portions of the ridge, are picrites. This contrasts with the transitional to alkalic compositions from the subaerial shield stage of Haleakala (Honomanu stage)

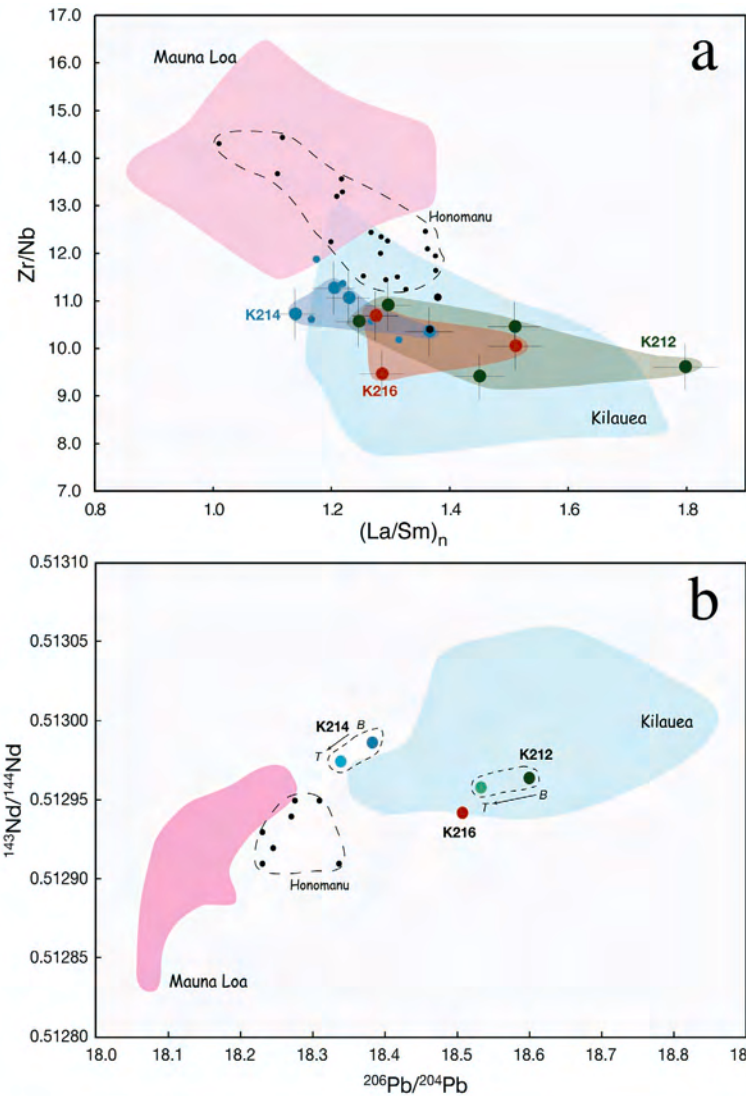


Figure 2. Variation diagrams for trace elements and isotopes in Hana Ridge samples analyzed to date [Johnson et al., 2002]. Fields for Kilauea and Mauna Loa are well separated in these plots. The Hana Ridge $(La/Sm)_n$ -Zr/Nb (a) and Nd-Pb isotope (b) data plot within the field for Kilauea basalts, while the Honomanu data (subaerial Haleakala) extend strongly into the Mauna Loa field. The clustering and relative positions of data points are consistent in both plots. B and T indicate bottom and top of dive track.

sparingly exposed in isolated outcrops [Chen et al., 1991]. Major and trace elements of the submarine Hana ridge rocks are similar to modern Kilauea and unlike subaerial Haleakala shield lavas. Each of the three subparallel ridges has a distinct trace element and rare earth element signature and their variations are geochemically self-consistent. This distinction carries over to radiogenic isotope compositions as well. The chemical variations defined by the sub-ridges plot within the general field for Kilauea lavas, but show consistent trends in trace element and isotope space from strongly "Kilauea-like" to transitional with "Mauna Loa-like" compositions; all subaerial Honomanu shield lavas plot within the Mauna Loa trace element and isotope fields (Figure 2).

Our results indicate that the mantle plume source for the Haleakala shield has changed over time from Kilauea-like compositions (high La/Sm , low Zr/Nb) in the submarine lavas to

Mauna Loa-like compositions (lower La/Sm , higher Zr/Nb) in the subaerial Honomanu shield lavas. Moreover, the subparallel ridges comprising the greater Hana Ridge show a trend from higher to lower La/Sm and $^{206}Pb/^{204}Pb$ with location. We infer that Haleakala shield volcano originally had tholeiite magma compositions whose source

material was similar to present-day Kilauea volcano and that the magma source became more Mauna Loa-like during growth of Haleakala volcano. Whether this temporal variation is consistent with that observed at Ko'olau volcano [Tanaka *et al.*, 2002] awaits further analysis. It is hoped that the data from this study, coupled with work on other Hawaiian shields, can improve our understanding of the structure and melting processes of the Hawaiian plume.

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What Can Seismology Say About Hot Spots?

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Seismology offers the highest-resolution view of mantle structure. In the decades since *Morgan* [1971] first proposed deep-mantle plumes, seismologists have used increasingly sophisticated methods to look for evidence of such structures, but so far they have had little success. This abstract outlines the relevant seismological methods for non-specialists and summarizes the current state of knowledge about structure beneath hot spots, to set the stage for the seismological component of this conference.

Factors Affecting Seismic-Wave Speeds

Direct thermal effect – If thermal plumes exist in the mantle, they would have lower seismic wave speeds than their surroundings. In the upper mantle, a 100 K temperature rise lowers the compressional-wave speed, V_P , by about 1%, and the shear-wave speed, V_S , by about 1.7%. In the deep mantle, this effect is several times weaker. The temperature anomalies proposed for plumes are about 200 to 600 K.

Indirect thermal effect – Temperature variations also cause variation in the depths of polymorphic phase boundaries in the transition zone between the upper and lower mantle. These are places where pressure causes certain minerals to change their crystal structure, and these changes are accompanied by jumps in density and seismic wave speed [*Anderson*, 1967]. Two such zones in particular, at depths of about 410 and 650 km, are global features and fairly easily detectable. A 100 K temperature rise would depress the “410-km” discontinuity by about 8 km, and raise the “650-km” discontinuity by about 5 km. (Both of these numbers are based on the assumption that olivine is the main mantle mineral, and are subject to significant uncertainty.) Thus a high-temperature anomaly would produce negative wave-speed anomalies at 410 km and positive ones at 650 km. The depths to these phase changes can also be measured directly using waves reflected from them (see **Receiver Functions**, below).

Chemical effect – If a plume has a different composition from the surrounding mantle, this alone will cause a seismic wave-speed anomaly. The sign and magnitude of the anomaly will depend on what minerals are involved. As a rule of thumb more buoyant materials have lower wave speeds, but an exception to this rule is mantle residuum – peridotite from which partial melt has been removed. Residuum is buoyant, but has a higher wave speed than its parent fertile peridotite.

Melting – The presence of even a small amount of melt in a rock can have a large effect on the seismic-wave speeds. Partial melting may reflect either thermal (high temperature) or chemical (low melting point) effects. The magnitude of the effect on seismic wave speeds depends

strongly on the geometric form of the melt bodies. Thin films on grain boundaries have the largest effect, and approximately spherical melt bodies have the smallest effect [Goes *et al.*, 2000].

Anisotropy – Seismic wave speeds and other properties of rocks vary with direction, and this can be as strong an effect as spatial heterogeneity. Olivine, in particular, becomes strongly anisotropic when flow causes crystal to align. Most studies of Earth structure ignore this effect, and their results probably are biased by this oversimplification. Studies dealing explicitly with anisotropy are becoming more common [Montagner, 2002].

Anelasticity – Many physical processes remove energy from seismic waves and convert it to heat, causing the waves eventually to die away. Most of these processes are thermally activated, so hotter regions are expected to exhibit stronger attenuation (high Q^{-1}). A side effect of anelasticity is to introduce a weak frequency dependence of the wave speeds, which must be accounted for in studies of Earth structure.

Seismic Tomography

The travel time of a seismic wave through the Earth gives an average of the wave speed along the ray path (but see **Bananas & Doughnuts**, below). If travel times are available for enough ray paths, passing through all parts of a region in many different directions, it is possible to unscramble the times to determine the three-dimensional wave-speed distribution. The term tomography, borrowed from medicine, is given to such seismic techniques. Seismic tomography is much more difficult than X-ray tomography, because the ray paths are curved and initially unknown, and in some cases the locations of the sources are poorly known. Three seismic tomography techniques are particularly useful in studying mantle structure:

Teleseismic Tomography – In order to study the structure immediately under an area, one can deploy an array of seismometers and record waves from distant earthquakes ($>\sim 2,500$ km away). Such waves arrive at angles within about 30° of the vertical, so crossing rays sample the structure down to depths comparable to the array aperture. The ray directions are not isotropically distributed, however; no ray paths are ever close to horizontal. Consequently, compact structures tend to be smeared vertically in images obtained by this technique [Keller *et al.*, 2000]. This smearing is unfortunate, because it generates artifacts that can be mistaken for real structures. It is possible to estimate quantitatively the severity of the smearing, however, and if due attention is paid to this error source, teleseismic tomography is the best technique available for studying the upper few hundred kilometers of particular regions.

Figure 1 shows results of one of the most detailed teleseismic tomography studies, of the structure beneath Iceland [Foulger *et al.*, 2001; Foulger *et al.*, 2000]. A strong low-wave-speed anomaly in the upper mantle, 200 to 250 km in diameter, extends to the deepest well-resolved depths, about 400 km. Significantly, though, the shape of the anomaly changes strongly and systematically below about 250 km, becoming tabular and parallel to the mid-Atlantic Ridge. This shape strongly suggests that the anomaly is related to plate-tectonic processes, a conclusion supported by whole-mantle tomography, which shows no continuation of the wave-speed anomaly beneath the transition zone [Ritsema *et al.*, 1999].

Whole-Mantle Tomography – There have now been thousands of seismometers deployed globally for decades, and millions of travel-time observations have accumulated and been used to derive three-dimensional models of the whole mantle. Some studies use enormous data sets

obtained from seismological bulletins such as that of the International Seismological Centre, but these data are subject to large and systematic observational errors. Others use data measured in more objective and consistent ways, usually using digitally recorded seismograms. Most whole-mantle models agree about the largest-scale anomalies (thousands of kilometers in size), but for a long time this was not so. The model that currently has the best resolution at depths of a few hundred kilometers is described by *Ritsema et al.* [1999]. This model shows, among other things, a strong low-wave-speed anomaly in the upper mantle beneath Iceland, which extends down to the transition zone but not to greater depths, confirming inferences drawn from teleseismic tomography of the region.

The resolution of whole-mantle tomography models is limited both by the ray distributions and by the state of computer technology. The smallest anomalies currently resolvable are 500 km or more in size. Furthermore, ray paths fall far short of sampling the Earth uniformly. Both earthquakes and seismometers are distributed irregularly over the Earth, and some places within the Earth are sampled poorly or not at all *e.g.*, the southern hemisphere, and particularly the south Pacific and Indian Oceans. The uneven ray distribution also systematically distorts anomalies in the Earth. As with teleseismic tomography, this distortion can be assessed quantitatively, but not by the general reader unless considerable information on this subject is given in the paper in question.

Surface-wave Tomography – Tomographic methods can also be applied to surface waves, low-frequency seismic waves that propagate in the crust and upper mantle and owe their existence to the presence of the free surface. The depth to which surface waves are sensitive depends on frequency, with low-frequency waves “feeling” to greater depths and therefore propagating at higher speeds. It is a rule of thumb that surface waves “feel” down to about a quarter of their wavelength. They also propagate at about 4 km/s, so this depth, in kilometers, is about $1/\text{frequency (Hz)}$.

Because of the distribution of earthquakes and seismometers, surface waves can often sample regions of the crust and upper mantle that body waves do not. They are also expected to be highly sensitive to plume heads, which are predicted to flatten out in the upper mantle, producing low wave speed regions that extend for thousands of km [*Anderson et al.*, 1992]. Body-wave and surface-wave data are often combined in whole-mantle tomography studies, such as that of *Ritsema et al.* [1999].

Bananas & Doughnuts

The statement above, that travel times are averages along ray paths, is a simplification. In reality, seismic waves “feel” the structure in a finite volume, and in fact *Dahlen et al.* [2000] have recently shown that travel times are most sensitive near a hollow surface around the ray, whose shape reminds them of certain snack foods. Figure 2 shows examples of the spatial distribution of sensitivity according to the new theory. Incorporation of frequency-dependent kernels into tomographic practice will significantly improve the quality of three-dimensional Earth models. The first results of such studies [*Montelli et al.*, 2003] show vertical low-wave-speed anomalies in the south Pacific, notably beneath Hawaii, Samoa-Tahiti, and Easter Island, but these are also regions of sparse data or clumps of data surrounded by regions lacking data (Figure 3), a

circumstance that would tend produce spurious structural features in tomographic images. The reality of these plume-like anomalies is a critical question at present.

Multiple ScS

Because they have limited resolution and can distort anomalies in complicated ways, tomographic results often are difficult to interpret. It would be much better if seismic waves sampled precisely a region of interest, and nothing else. Happily, nature occasionally arranges an experiment for us in just this way. For example, the seismic phase *ScS*, a shear wave reflected from the core-mantle boundary (CMB), when observed close to the epicenter of an earthquake, has a nearly vertical ray path through the entire mantle [Anderson and Kovach, 1964]. Such waves are ideally suited to looking for narrow vertical structures such as plumes.

On April 26, 1973, a magnitude 6.2 earthquake occurred in Hawaii, and the records from seismometers on Oahu show an usually clear train of multiple-*ScS* phases, reflected repeatedly between the Earth's surface and the CMB [Best *et al.*, 1975]. These waves are sensitive to structure in a vertical cylinder with a diameter of about 500 to 1000 km extending down to the CMB. They show no indication of a plume. The wave speed V_S in the upper and middle mantle inferred from arrival times is higher than the average for the southwestern Pacific [Katzman *et al.*, 1998], and the propagation efficiency (Q) is also high [Sipkin and Jordan, 1979; Sipkin and Jordan, 1980]. Figure 2 shows the sensitivity kernel for these *ScS* waves at different depths in the mantle. The location of a possible plume in the lower mantle might be far enough from Hawaii that these *ScS* waves would not sample it, but these observations argue strongly against a large region of unusually high temperature or extensive melting in the upper mantle beneath Hawaii. In particular, an upper-mantle anomaly similar to the one beneath Iceland appears to be ruled out.

Receiver Functions

When a compressional or shear seismic wave strikes a discontinuity in the Earth, it generates reflected and transmitted waves of both types. Because of this, waves from distant earthquakes passing through a layered medium such as the crust or upper mantle generate complicated seismograms containing many echoes. To interpret these records, seismologists process them to generate simplified artificial waveforms, somewhat inscrutably called *receiver functions*. These can be inverted to yield the variation of V_S with depth, and they are particularly sensitive to strong wave-speed discontinuities. Receiver functions are particularly powerful for studying the depths to the Moho and the "410-km" and "650-km" discontinuities, which may provide evidence about crustal thickness and temperature at these depths [Du *et al.*, 2002].

One of the most detailed receiver-function studies done to date took the form of a profile across the eastern Snake River Plain, the suggested track of a mantle plume now beneath Yellowstone, which lies at the northeastern end of the Plain [Dueker and Sheehan, 1997]. The results illustrate the complexity of structures revealed by receiver functions, and some of the difficulties of interpreting them. Several "discontinuities" are present, in addition to the major ones near 410 and 650 km. Even these two major features are not continuous, and the 410-km discontinuity appears to split in two near the left end of the profile. The depths to the 410- and 650-km discontinuities are expected to be negatively correlated if their topography results from temperature variations, but actually they are weakly positively correlated. The receiver functions thus provide no evidence of elevated temperatures

“Plume waves” (“Fiber waves”)

Zones of low wave-speed trap energy and act as waveguides, along which waves propagate efficiently for great distances. This is the principle behind fiber-optic communication. The narrow low-wave-speed anomalies expected for plumes would be ideal for supporting such waves (J. R. Evans, personal communication, 2002), but no such plume waves have ever been noticed. This failure might indicate merely that nothing excites plume waves efficiently (there are no earthquakes in the lower mantle), or it might mean that plume waveguides do not exist. Quantitative theoretical investigation of the excitation and propagation of plume waves would be highly worthwhile.

Summary

The main methods for studying Earth structure in a way that is useful in the search for plumes include seismic tomography, studying the transit times and attenuation of individual waves that penetrate the volume of interest, and the use of receiver functions to study topography on the boundaries of the transition zone. A potentially new and interesting approach is the search for “plume waves”. Whereas downgoing slabs in subduction zones and their effects on the transition zone have been easy to detect, the same cannot be said about plumes, heads or tails, and promising images often have not proved reproducible by later, more detailed studies. It will be interesting to follow what the next decade brings.

Figure captions

Figure 1: Compressional-wave speed (V_P) anomaly in the upper 500 km beneath Iceland, imaged by *Foulger et al.* [2001] using teleseismic tomography techniques. Inside the green surface, V_P exceeds the average value at a given depth by 0.5%. At shallow depths the anomaly is approximately cylindrical, but beneath 250 km it changes shape, and becomes tabular, with its long dimension parallel to the mid-Atlantic Ridge.

Figure 2: Frechet kernels giving the sensitivity of the travel time of the seismic phase ScS , a shear wave that reflects from the core-mantle boundary, to changes in the shear-wave speed at different depths, computed using the theory of *Dahlen et al.* [2000] for waves with dominant frequency 0.035 Hz. The geometry is appropriate for waves generated by the Hawaii earthquake of April 26, 1973 (star) and recorded at station KIP (triangle), as reported by *Best et al.* [1975] and *Sipkin and Jordan* [1980]. The maps have been corrected for convergence of the verticals, which otherwise exaggerates the size of deeper features; thus the distance scale, not than the geographic coordinates, correspond to reality in the deep mantle. The two shallowest maps have attenuated color scales.

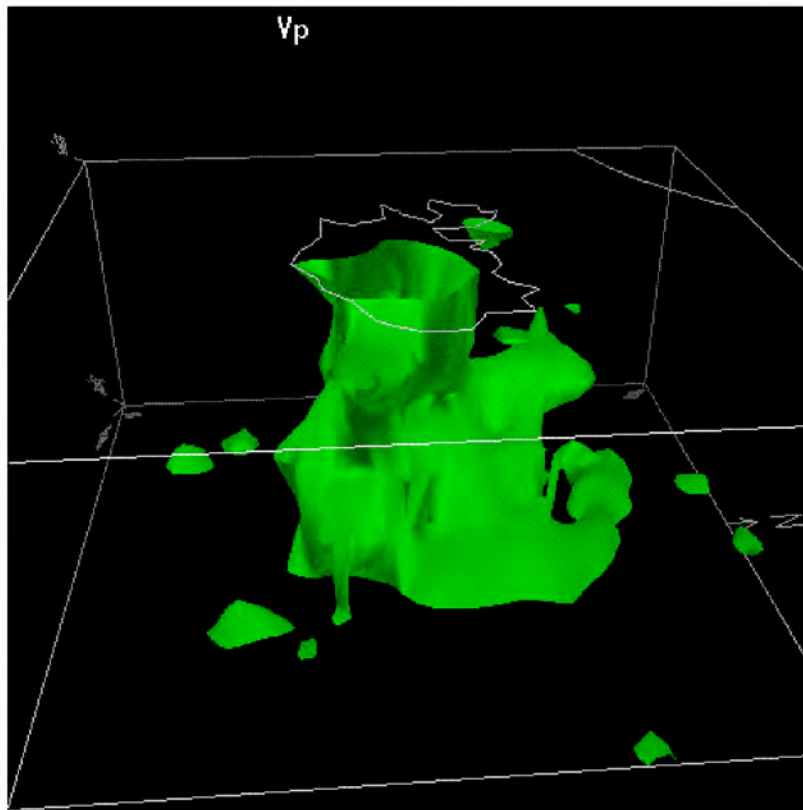
Figure 3: Sampling of the mantle by the travel time data set of *Bolton and Masters* [2001], which was used, in addition to high-frequency data from the ISC Bulletin, in the recent tomographic study of *Montelli* [2003]. Squares show surface projections of turning points of rays in the indicated depth ranges. P phases are shown on the left, and S phases on the right. In the central and south Pacific, coverage is sparse and observations occur in clumps that correspond closely to the plume-like anomalies reported by *Montelli* [2003].

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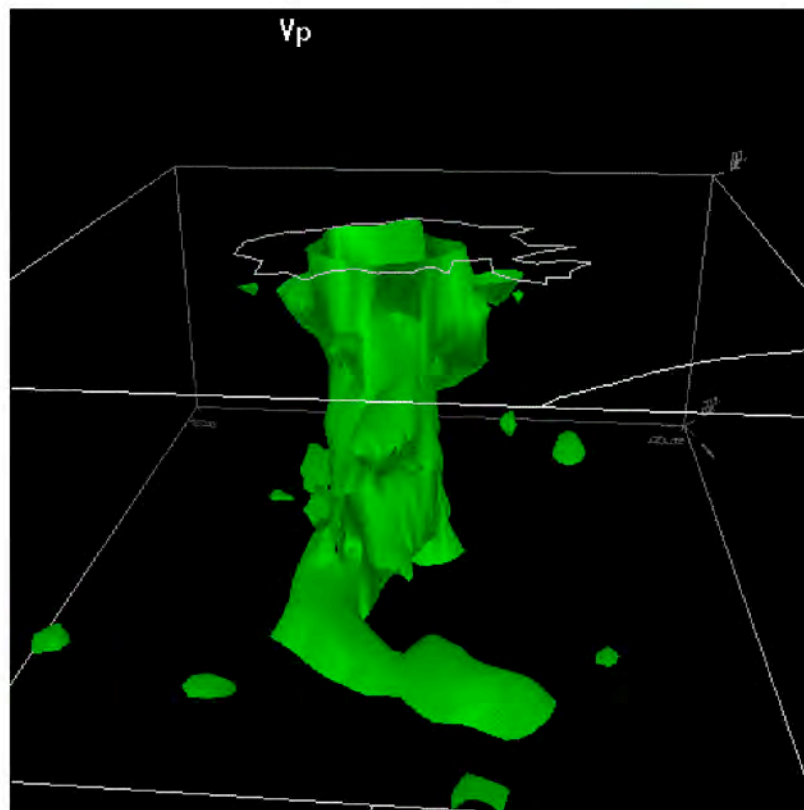
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a)



Looking
west

b)



Looking
south

Figure 1: Views of Iceland upper-mantle anomaly

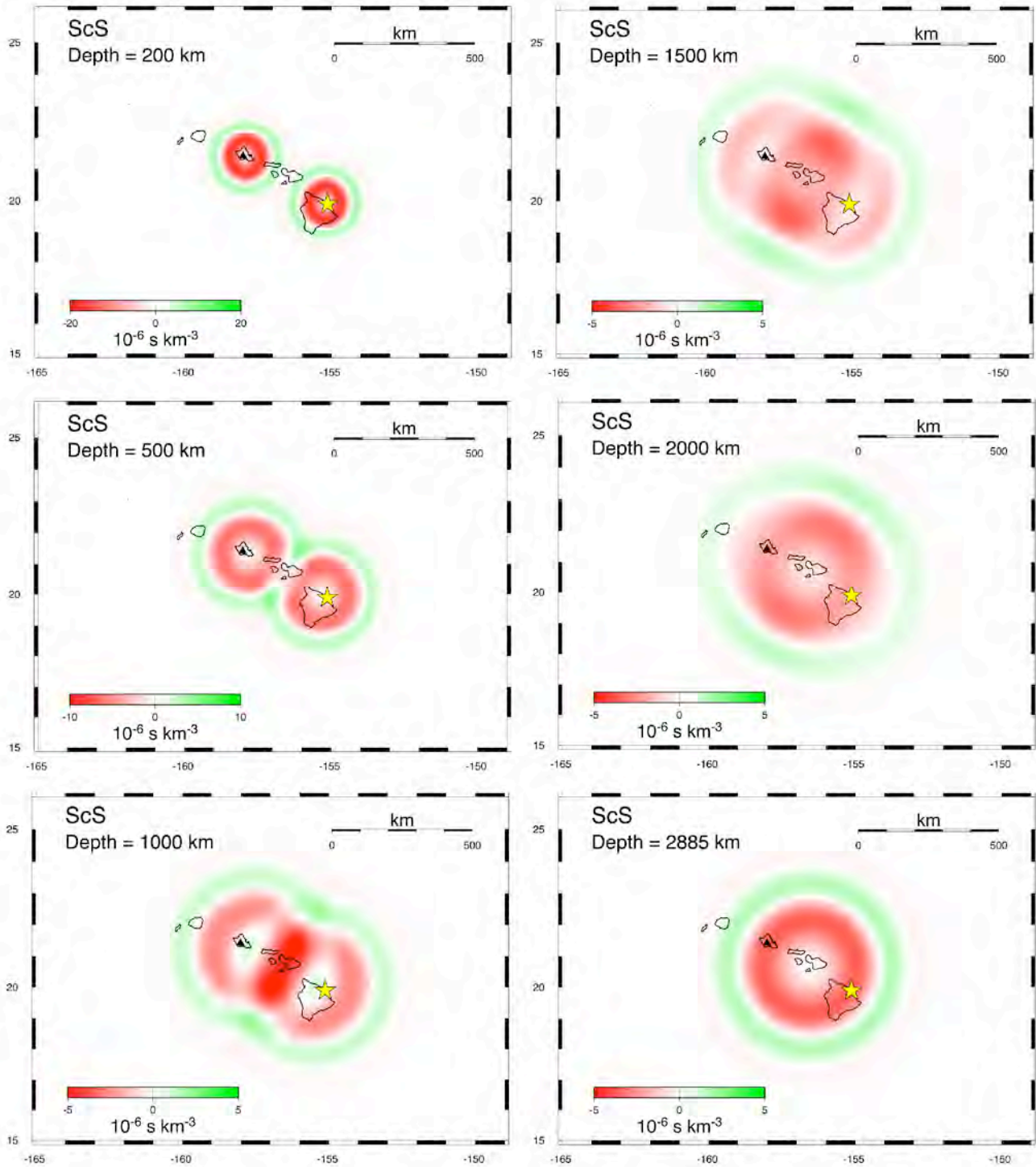


Figure 2: ScS travel-time data kernels

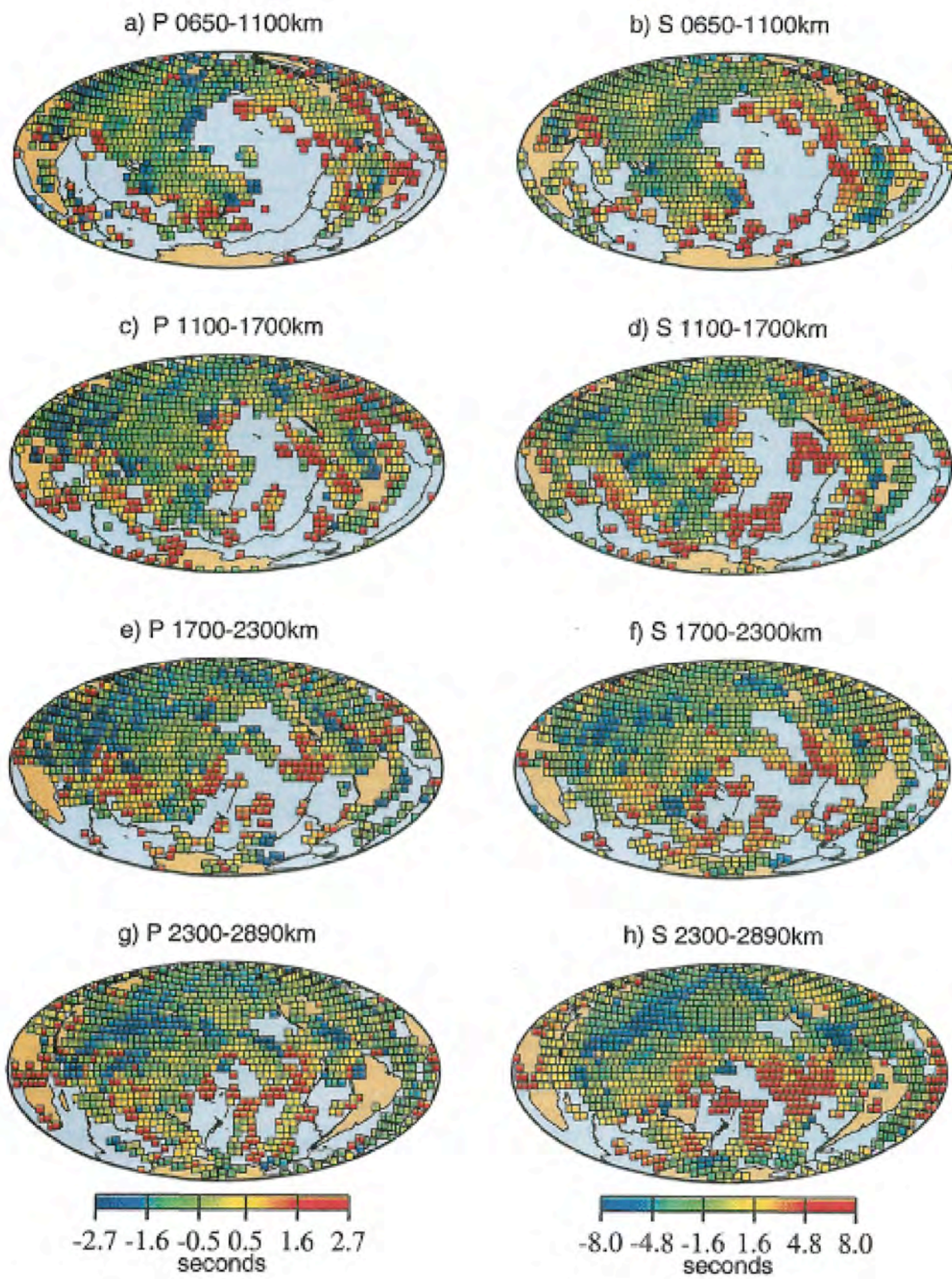


Figure 3: Geographic sampling of Bolton-Masters dataset

Upwellings on Venus: Evidence from Coronae and Craters. D. M. Jurdy¹, P. R. Stoddard², and A. Matias¹,
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Introduction: Venus' surface underwent global re-surfacing at about 300-500 m.y. ago, as estimated from impact crater counts [1]. Although our sister planet lacks apparent plate tectonic features, it has experienced tectonic and volcanic activity, some quite recently. Hundreds of coronae adorn its surface. These nearly circular features, ranging in size from 100 to 2600 km, may be caused by localized upwellings and thus be analogous to Earth's hotspots. The current deformation of Venus' surface has been attributed to a swell-push force, defined as the gradient of the geoid height [2]. Two likely areas of current geologic activity, Atla Regio and Beta Regio are marked by prominent geoid highs. We compare these two features using their coronae and craters to establish the nature and relative timing of their formation.

The Regios: Atla and Beta Regios are both marked by pronounced topographic and geoid highs. Each lies at the intersection of multiple rifts - the chasmata system. Venus' chasmata system can be fit by great circle arcs at the 89.6% level [3]. When corrected for the smaller size of Venus, the length of the chasmata system measures [3], within 2.7% of the 59,200-km length of the spreading ridges determined for Earth by Parsons [4]. These, likely surface expressions of mantle upwellings, represent the most recent evidence of tectonic/volcanic activity on the planet. We examine the distribution, style, and attitude of coronae, and the location, and modification of craters with respect to these two geoid highs, noting the strong correlation of Earth's hotspots with its geoid highs [5].

Corona Classification: Corona formation on Venus could be caused by rising diapirs. Unlike Earth, Venus shows no evidence of horizontal motion, resulting in juxtaposition of its coronae of all ages. Also, there is little erosion to modify features. DeLaughter and Jurdy [6] have classified 394 coronae based on the morphology of the interior, terming them domal, circular, and calderic. They proposed that these differing styles reflect different stages in the evolution of a corona, from domal (youngest, perhaps still active) features, progressing through increasing degrees of collapse to the calderic coronae. Comparison of elevations of these features shows the domal coronae average higher elevations, and calderic at lower elevations, with circular in between. Jurdy and Stoddard [7] attempted to relate types of coronae to dif-

ferent chasmata in the Beta-Atla-Themis (BAT) region, the most volcanically active region of Venus. They suggest that the chasmata may also be of different stages of activity.

Craters: Impact craters on Venus are used to determine relative age and degree of modification of a region. Venus hosts approximately 940 craters, of which about 158 are tectonized, and 55 embayed, but only 19 planetwide are both tectonized and embayed [1]. Of these, four are near the crest of Atla, and one on its flanks, while Beta is home to three on its flanks and one on its crest. In addition, Beta and Atla have high relative percentages of craters that have been tectonized or embayed (Figure 1a, 1b). Although the

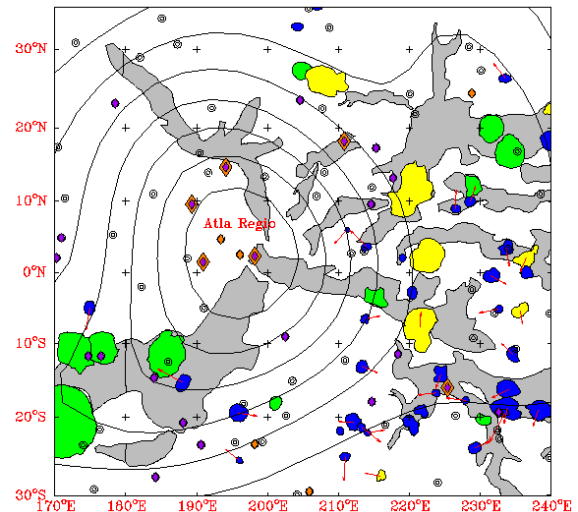


Figure 1a: The Atla Regio region. Geoid is indicated by 10-meter contours. Craters: small circles – unmodified; orange diamonds – embayed; purple diamonds – tectonized; purple within orange – tectonized and embayed. Grey regions indicate rifts. Corona color scheme same as in Figure 1. Red arrows indicate dip directions, where attainable.

overall global distribution of craters approximates random, areas with high concentration of altered craters clearly are more active. Most of the crest of Atla may still be active; otherwise some recent craters should be pristine. Indeed, Crater Uvayasi (2.3N, 198.2 about 38km diameter), one of only about 50 with parabolic dark halos - thought to be the youngest of all craters [8] - shows significant modification, both tectonic disruption and embayment by lava. We may therefore infer that Atla, with a higher

percentage of altered craters, (8/17 or 47%, within three 10 meter geoid contours of the summit) than Beta (4/14 or 29%, in the same range) is a more active or recent feature than Beta. Both are younger than "average" terrain ((158 T + 55 E)/940 = 23%). Besides the evidence of tectonic/volcanic forces modifying craters, there exists a slight but systematic deficit of about 20-30 craters close to the chasmata [9].

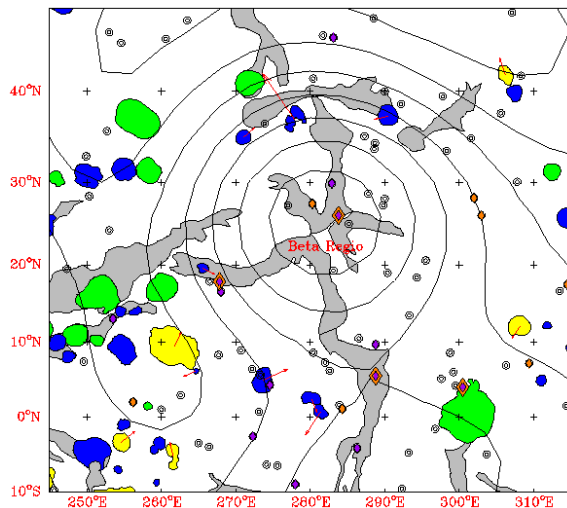


Figure 1b: The Beta Regio region. Symbols are the same as in Figure 1a.

Both Atla and Beta are ringed by many coronae, but neither has coronae at or near their crests even within 2 contours of their geoid highs (Figure 1a, 1b). Coronae do occur in many rift segments, yet none occurs at or near these intersection points. Perhaps just as remarkable, Atla has a partial ring of four domal coronae, all between 4 and 5 geoid contours from the crest, while Beta has a partial ring of 6 or so calderic coronae between three and four contours from its crest. In both instances, the rings parallel geoid contour lines. These are the nearest coronae of their type to the crests. If corona formation is contemporaneous with the uplift process at Atla and Beta, and if the domal are younger than the calderic coronae, then Atla should be more recent or active than Beta, in agreement with the modified-crater analysis. If these dating schemes are accurate, and uplift of Atla post-dates formation of calderic coronae, then we would expect the latter to be tilted away from the center of uplift at Atla. Conversely, if calderic coronae formed concurrently with the uplift of Beta, (or domal formed concurrently with Atla uplift) then we would not expect to see such a pattern of dips. To test this, we determined the average dip of each corona (red arrows in Figure 1a, 1b). These determinations are at best very rough, but there is a suggestion that the calderic coronae dip away from Atla's crest, espe-

cially for those immediately west of the uplift. Neither the domal features at Atla nor the calderic features at Beta show such a pattern.

Conclusions: We use two independent schemes for assessing the relative timing of the uplift of Atla and Beta Regios. Both schemes, percentage of craters altered, and style of corona, suggest a younger age for the uplift of Atla. Another test of this result, orientation of the dip directions of coronae near each uplift, is consistent with the younger age for Atla. On the basis of dark halos, Basilevsky and Head [10] have also come to the conclusion that Atla younger than Beta. Our model is shown in Figure 2.

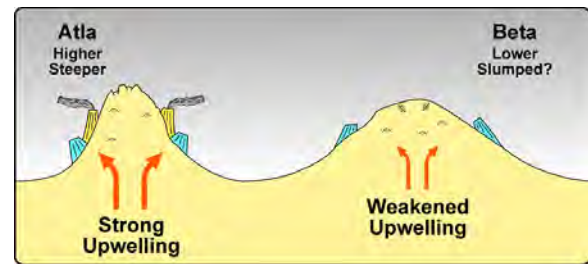


Figure 2: Schematic showing relationship between upwelling mantle and Atla and Beta Regios. Atla, presumably still active, has new, domal coronae (yellow) forming atop older, tilting calderic coronae (blue). Atla's crest should be in extension. Beta, no longer active, has begun to slump, tilting the calderic coronae towards the crest of the regio, which should be in compression.

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Edge Driven Convection and Iceland

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One of the alternative hypotheses for hotspot volcanism is Edge-Driven Convection (EDC). A small-scale convective instability forms at any step or discontinuous change in thickness in a thermal boundary layer [e.g., Elder, 1976]. The EDC hypothesis envisions that this instability will form at boundaries between stable cratons and oceanic, or young continental, lithosphere [e.g., King and Anderson, 1995; 1998]. Because this is a relatively weak instability, large lateral variations in temperature or fast plate-scale flow can overwhelm EDC instabilities. (These same factors would also overwhelm most plume instabilities.) In the Central Atlantic, seismic tomography supports the EDC hypothesis—seismically fast anomalies (presumably cold downwellings) are observed at the edge of the South American and West African cratons (Figure 1) just where we would expect the downwelling limbs of EDC to be located. King and Ritsema [2000] conclude that many, if not all, of the off-ridge Central and South Atlantic hotspots could be explained by EDC.

The question remains, “Can an EDC mechanism explain the excess volcanism at Iceland (and the Azores)?” The answer is not as clear as we would like, yet it is premature to throw out the EDC hypothesis. The biggest obstacle to considering the EDC hypothesis for Iceland is that there are no seismically fast anomalies beneath the Greenland and Scandinavian cratons. In order to properly evaluate whether this is strong enough to rule out EDC, we need to understand the time evolution and the effect of the width of the ocean basin on the planform of EDC. We know that the North Atlantic began opening later than the Central and South Atlantic. Based on the temporal evolution of the EDC instability, it is possible that a downwelling broad enough to be observed seismic tomography may not yet have formed. The North Atlantic is the narrowest part of the Atlantic basin. A detailed parameter study reveals that an EDC instability at one or both of the cratons bounding the North Atlantic could up well at the Mid-Atlantic ridge.

In the central and southern Atlantic, which are much wider than the North Atlantic, hotspots occur off the ridge axis. This is in agreement with the study of basin width and EDC. Furthermore, EDC predicts that these upwellings should be weaker because, the of the time since the beginning of spreading in the southern and central parts of the Atlantic and, only one EDC instability contributes to each of these hotspots whereas EDC instabilities from both sides contribute to ridge-centered hotspots.

To understand the origin of Iceland (and the Azores), it is important to explain not only the excess volcanism on the Mid-Atlantic ridge, but also to understand the North Atlantic geoid and topographic anomalies. The spherical harmonic degree 3-12 expansion of the geoid from the GEM-T2 model is contoured in Figure 2. I have purposely removed the degree 2 term of the expansion because the large $l = 2, m = 0$ term obscures the lateral variation in the geoid. There is a broad geoid high in the North Atlantic extending from north of Iceland, along the Mid-Atlantic ridge, to south of the Azores. For reference, the star at the southern tip of the white oval represents the Azores and the one at the northern tip of the white oval represents the Jan Mayen hotspot. The star just to the south of Jan Mayen represents the Iceland hotspot.

The North Atlantic geoid high one of several prominent geoid highs on the planet. Others include the geoid high in the western Pacific associated with the Solomon and New Britain subduction zones [c.f., Hager, 1984] and the geoid high in the southern Indian Ocean, associated with the uplifted plateau of southern Africa [Lithgow-Bertelloni and Silver, 1998; Gurnis *et al.*, 2000]. The previously cited studies propose that the African/Indian geoid high and the elevated

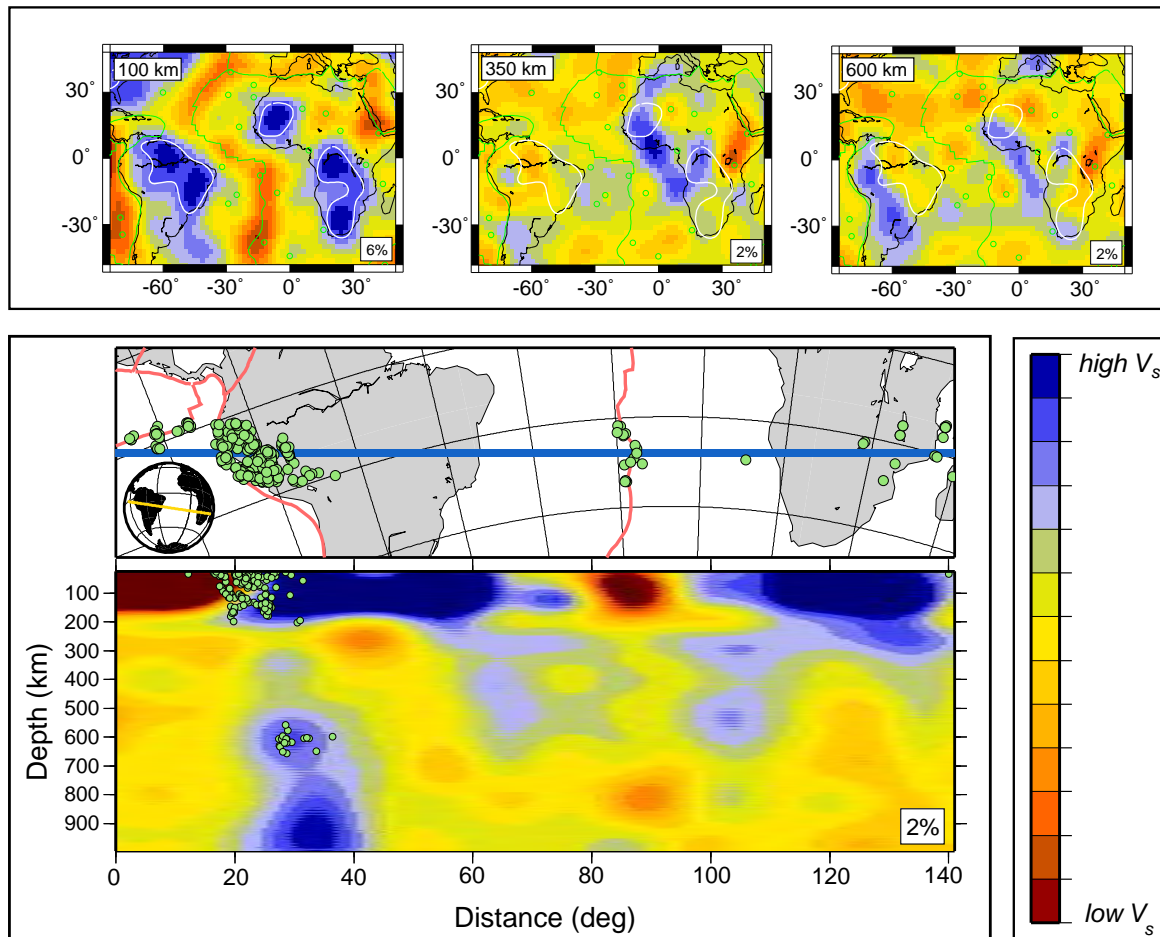


Figure 1: (A) Horizontal cross-sections through seismic tomography model S20RTS [Ritsema *et al.*, 1999] at depths of 100 km, 350 km, and 600 km. Relatively high velocity and low velocity regions are indicated by blue and red colors, respectively, with an intensity that is proportional to the amplitude of the velocity perturbation from the PREM. Green lines represent plate boundaries. White lines circumvent regions in the mantle where the seismic velocity at a depth of 100 km is larger than in the PREM by 4% or more. These regions roughly outline the location of Precambrian cratons. (B) A 140 deg wide cross-section through S20RTS across South America and southern Africa. The green circles indicates locations of earthquakes in the Harvard CMT catalog.

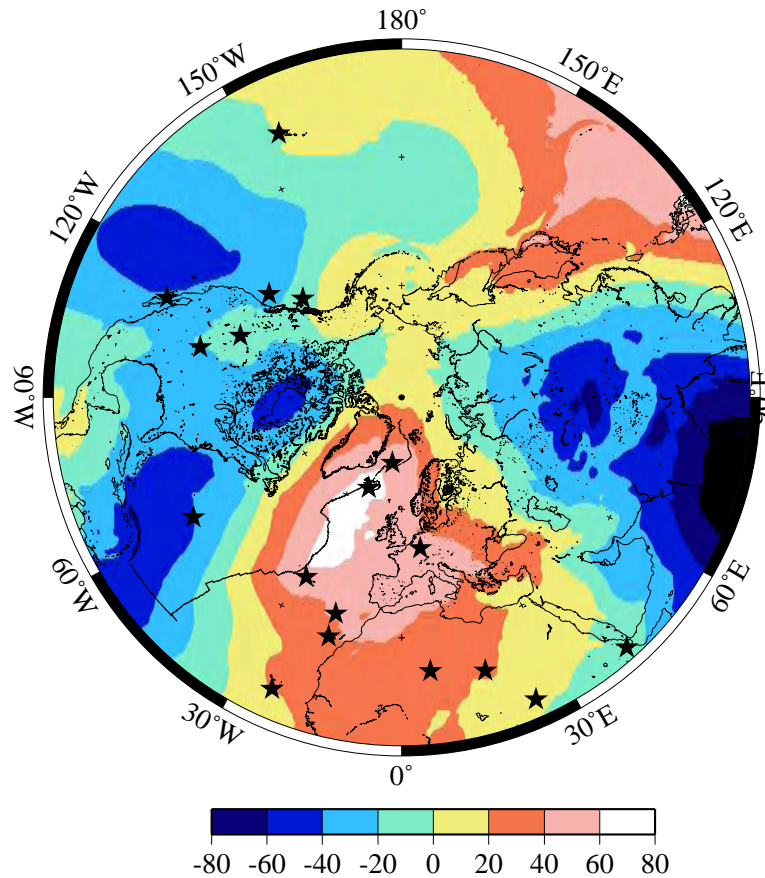


Figure 2: The North Atlantic geoid from GEM-T2. The boundary between the North American and European Plate is shown by the black line. The solid stars are the locations of hotspots.

plateaus in southern Africa are caused by mantle flow driven by a large, slow seismic velocity anomaly in the deep mantle, sometimes referred to as ‘the Great African plume.’ In hotspot swell studies, features this wavelength are removed by filtering. Interestingly, Iceland is one of three significant geoid anomalies that remain unexplained after accounting for subducted slabs and glacial isostatic adjustment [Simons *et al.*, 1997; Mark Simons thesis].

Seeking a consistent explanation for the southern Indian/African and the North Atlantic geoid highs, it is clearly tempting to relate the North Atlantic geoid high to a deep sourced mantle upwelling beneath Iceland. Note that the scale and morphology of ‘the Great African plume’, and presumably the deep upwelling that supports the North Atlantic geoid and topographic anomalies is completely different than what is envisioned in the plume hypothesis. Within the field of Figure 2 there are many other hotspots including: Bermuda, Azores, Canary Islands, Cape Verde, Madeira and Yellowstone (solid stars in Figure 2). There are no comparable features in the geoid that can be associated with these hotspots (except for the anomaly along the Mid-Atlantic Ridge which encompasses both Iceland and the Azores). Geoid anomalies associated hotspots are typically on the order of 1-5 meters [Crough, 1983, Monnereau and Cazenave, 1990; Sleep, 1990] whereas the North Atlantic geoid anomaly is on the order of 50 meters. This would argue that either there is more than one mechanism for hotspot volcanism or that there is no direct relation between the North Atlantic geoid and topographic anomalies and the excess volcanism at Iceland.

Most of the gravity and topography swells associated with the Hawaiian hotspot can be explained by compositional, as opposed to thermal, density anomalies [McNutt *et al.*, 1998]. Further, the heatflow anomaly at Hawaii is better explained by a hydrothermal circulation model than it is by residual heat from a mantle plume [Harris *et al.*, 2000]. After accounting for hydrothermal circulation and compositional density effects, the remaining anomalies that could represent the thermal plume are very small. Hawaii and Iceland are often considered the type examples of plumes. If the swell and geoid at Iceland are the results of a deep mantle plume, then it is truly a plume with no equal on the planet.

One can envision several experiments to distinguish between a deep and shallow source for the North Atlantic surface anomalies. Most obviously, the magnitude geoid and topographic anomalies generated by EDC can be compared with the North Atlantic anomalies. It is difficult to produce the magnitude of geoid and topographic anomalies with EDC calculations; however, work is still underway. Geoid and topographic swells from temperature-dependent plume calculations with realistic parameterizations are compared with observations. Most of the surface anomalies from these calculations are related to upper mantle plume/lithosphere thermal anomalies as opposed to deep anomalies. From this we conclude that it may not be possible to completely resolve upper versus lower mantle sources for surface anomalies. (In the sense that the upper mantle source in this case, the plume influenced lithosphere 'lithospheric erosion,' is related to the deep mantle instability.) However, the spatial extent of the Great African plume, which is thought to be responsible for similar-scale geoid and topographic anomalies in the southern Indian ocean, is quite different from the plume structures envisioned in the plume hypothesis. By comparing the pattern of seismic tomography in the upper and lower mantle with the pattern of the residual geoid from Simons global geoid analysis, we can attempt to isolate the source of the anomalies by matching spatial patterns. A preliminary investigation is promising and more results will be presented.

My working hypothesis is that the volcanism has an upper mantle (EDC) origin while the topography and geoid has a deep mantle origin (like the proposed South African/Southern Indian ocean geoid and topographic anomalies).

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MANTLE WEDGE FLOW AND THERMAL MODELS FOR THE CENTRAL MEXICAN SUBDUCTION ZONE

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In the last years, the models of the mantle wedge flow in the subduction zones, including the thermal models are applied in order to explain the source of the arc volcanism.

In this study the flow in the mantle wedge is modeled using Navier-Stokes equations for an incompressible fluid with isoviscosity and creep diffusion law for olivine. Then the corresponding thermal models were developed for the Central Mexico subduction zone. Both thermal models, one for the isoviscosity and another for the creep-diffusion law for olivine, suggest that the subducting plate beneath the volcanic arc scarcely reach the melting conditions. Strong temperature gradients are observed in the model with the creep-diffusion viscosity, suggesting that the mantle beneath the volcanic arc is melted, being a source of the volcanism in the area. The model reveals also the conditions for the serpentine (and associated minerals) stability in the tip of the wedge. The location of the serpentine in the wedge may indicate a down dip limit of the intraplate earthquakes.

Magmatic processes in the oceanic lithosphere: characterization of the ultramafic and mafic materials from the Holocene volcanic centers of Bandama and La Caldera de Pinos de Gáldar (Gran Canaria, Canary Islands)

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Introduction

Mantle and crustal xenoliths carried to the surface by basaltic melts in intraplate oceanic islands represent one very important information source of the deeper parts of the oceanic lithosphere. Thus, our research is based on the study samples of peridotite xenoliths and olivine-pyroxene-amphibol crystals and cumulates from pyroclastic deposits of Bandama and La Caldera de Pinos de Gáldar volcanic centers (Gran Canaria, Canary Islands) and samples of lava flows associated to these volcanoes. We have combined data from the volcanological study and the electronic microprobe and microthermometric analysis of minerals and melt and fluid inclusions, and the bulk-rock analysis of these samples to define some characteristics of magmatism which gave rise to these volcanoes.

Gran Canaria island is at the centre of the Canary Archipelago and this is associated to a mantle plume. Geophysical data indicate the occurrence of an oceanic-type basement underneath the Gran Canaria island modified by massive intrusions with a Moho depth of 13 kms. Magmatic activity of this island began in the Miocene with an episode of submarine volcanism and their subaerial volcanism have been divided over three main episodes: Old cycle (Miocene, 14.5-7 Ma), Roque Nublo cycle (Pliocene, 5.5-3 Ma) and Recent cycle (Plio-Quaternary, 3 Ma-Present). Taking into account the hawaiian formation model for intraplate volcanic islands, the Old cycle of Gran Canaria contains magmatic materials associated to the shield building and alkaline declining stages, and the Roque Nublo and Recent cycles represent the rejuvenation stage. Thus, Bandama and La Caldera de Pinos de Gáldar Holocene volcanic centers are one of the latest eruptions associated to the rejuvenation stage of Gran Canaria.

The Holocene volcanic center of Bandama is made up of a volcanic caldera (La Caldera de Bandama) and a strombolian cone (El Pico de Bandama). Bandama eruption began in the area of La Caldera with a strombolian episode which produced deposits of fall and a basanite lava flow. Later, the magma in the conduit mixed periodically with groundwater, producing both phreatomagmatic eruptions with base surge deposits and strombolian eruptions with fall deposits. Although the volcanism episodes were to continue in the area around La Caldera, most of these manifestations were centred around the NW area of the fissure, producing the strombolian cone known as the Pico Bandama, abundant deposits of fall in the surrounding areas and an intracanyon flow. The genesis of La

Caldera is posterior to El Pico and was produced as a result of the last effusive-explosive eruptions and the eventual collapse of the volcanic cone located to the SE of the fissure.

On the other hand, the Holocene eruption of La Caldera de Pinos de Galdar (2,830 years b.p.) is characterized by an strombolian cone, several lava flows of basanite composition and fall pyroclastic deposits. This volcano and other neighbour volcanic cones (Valleseco, Montañón Negro –2,970 years b.p.--, Hondo de Fagagesto –2,210 years b.p.--, Sao and Los Berrazales) are almost contemporaneous and they form a NW-SE structural lineation of 9 km long.

Mineralogical and geochemical characteristics

Lava flows of Bandama show basanite composition (SiO_2 41.8% and sum of the alkalis 3.8%) and porphyritic texture with phenocrysts and microcrysts of olivine, pyroxene and oxide (spinel and ilmenite) with sizes under 2 cm. Microprobe analysis of these minerals reveal olivine Fo_{70-89} (Ni/Ca ratio: 0.3-2.2) and clinopyroxene Wo_{40-52} , En_{27-54} (augite-diopside). Spinel appear as subordinate minerals and as inclusions in phenocrysts and have Cr_2O_3 content between 36.4 and 40.2%. These analysis show variations of composition between the cores and rims of some phenocrysts and reveal different crystal generations. In addition, the lava flows contain some ultramafic and mafic xenoliths and cumulates

On the other hand, the lava flows of La Caldera de Pinos de Galdar have basanitic composition (SiO_2 43.2% and sum of the alkalis 5%), show porphyritic texture and these flows contain olivine (Fo_{81-90}) and augite-diopside (Wo_{44-55} , En_{28-50}) phenocrysts, subordinated spinel ($\#\text{Cr}_{35-65}$) and magnetite ($\#\text{Cr}_{0.4-3.3}$), and scarce mafic and salic cumulates.

Phreatomagmatic eruptions of Caldera de Bandama form base surge and explosive breccia rich in lithics (such as phonolites, tephrites, alkaline basalts, etc.) deposits with spectacular volcanic-sedimentary structures (imbricated channels, sand waves, bomb impacts, etc.). However, the strombolian volcanism is characterized by the presence of fall deposits, containing peridotite xenoliths and olivine-pyroxene-amphibole megacrysts and magmatic cumulates with sizes under 7 cms. Electronic microprobe analysis of these minerals show:

1) phenocrysts and megacrysts of olivine Fo_{77-89} (Ni/Ca: 0.5-2.2 and spinel with Cr_2O_3 : 34-36.3%), clinopyroxene Wo_{34-52} , En_{33-51} (spinel with Cr_2O_3 : 20.8-23.4%) and kaersutite (destabilized to fassaite -- Wo_{54-57} , En_{31-34} --, rhönite, olivine and subsaturated melt).

2) ultramafic cumulates (dunite, werhlite, clinopyroxenite with olivine and clinopyroxenite) with olivine Fo_{80-87} (Ni/Ca: 0.6-2.1 and spinels with Cr_2O_3 : 14-31.2%) and clinopyroxene Wo_{39-51} , En_{36-55} (spinels with Cr_2O_3 : 10-16.1%).

3) ultramafic xenoliths (dunite and lherzolite) with olivine Fo_{83-89} (Ni/Ca: 1-20 and spinels with Cr_2O_3 : 13.2-32.8%), orthopyroxene Fo_{84-85} (spinels with Cr_2O_3 : 31.3%) and clinopyroxene Wo_{40-44} , En_{48-50} .

The pyroclastic deposits of La Caldera de Pinos de Galdar display mafic and ultramafic xenoliths and cumulates with sizes below 15 cms and these are characterized by:

1) ultramafic xenoliths: A) dunite (olivine I: Fo_{87-91} , olivine II: Fo_{89-93} ; spinel I: $\#Cr_{95}$ and spinel II: $\#Cr_{50-96}$), B) lherzolite (olivine I: Fo_{91} ; olivine II: Fo_{91-94} ; orthopyroxene: Fo_{81-96} ; clinopyroxene: Wo_{41-47} , En_{50-57} ; spinel I: $\#Cr_{90-91}$ and spinel II: $\#Cr_{89-96}$) and C) harzburgite (olivine I: Fo_{88-93} ; olivine II: Fo_{82-97} ; orthopyroxene: Fo_{91-96} ; clinopyroxene: Wo_{40-47} , En_{48-58} ; spinel I: $\#Cr_{86-87}$ and spinel II: $\#Cr_{67-88}$).

2) mafic and ultramafic cumulates: A) olivine clinopyroxenite (olivine: Fo_{81-83} ; augite-diopside: Wo_{47-56} , En_{38-45} ; spinel: $\#Cr_{40-57}$), B) clinopyroxenite (olivine: Fo_{81-83} ; augite-diopside: Wo_{46-55} , En_{39-51} ; spinel: $\#Cr_{13-30}$), C) hornblende clinopyroxenite (olivine: Fo_{77-91} ; augite-diopside: Wo_{42-55} , En_{38-56} ; spinel: $\#Cr_{15-44}$), D) clinopyroxenite with apatite and Fe-Ti oxides (augite-diopside: Wo_{50-53} , En_{39-44} ; magnetite: $\#Cr_{<2.1}$; ilmenite: Ilm_{55-66} , $Geik_{16-34}$, Hem_{10-17}), and E) clinopyroxenite with apatite, titanite and Fe-Ti oxides (augite-diopside: Wo_{50-56} , En_{25-44} ; magnetite: $\#Cr_{<1.8}$; ilmenite: Ilm_{52-86} , $Geik_{6-34}$, $Hem_{<19}$; hematite: Hem_{96-100} , $Geik_{<4}$, $Pyroph_{<3}$; and Fe sulphides). Clinopyroxene megacrysts (Wo_{50-54} , En_{35-42}) containing crystal inclusions of apatite and hematite (Hem_{98-100} , $Geik_{<1}$, $Pyroph_{<1}$).

Melt inclusion study in olivine from fall deposits and lava flows of Bandama volcanic center show SiO_2 content: 35-44%, sum of alkalis: 4.1-7% and high values of S and Cl (<5,500 and <980 ppm, respectively). The melt inclusion compositions are different from the lava whole rock and the interstitial melt. The melting temperatures (T_m) of melt inclusions range between 1,060 and 1,260°C.

Microthermometric study of melt and fluid inclusions in olivine and clinopyroxene of Bandama volcanoes shows pure, or almost pure, CO_2 trapped in the gas bubble. These carbonic fluids reveal a wide range of T_h of CO_2 (-39 to 31°C) in liquid, indicating minimum depths of mineral formations (at T_m : 1200°C) between 12-18 Km for olivine-pyroxene- amphibole megacrysts, 4.5-27 Km for olivine-pyroxenes cumulates, 10.5-25 km for ultramafic xenoliths, and 7.5-33 Km for olivine phenocrysts of lava flows. The depth ranges observed indicate magma reservoirs and mineral genesis within the upper mantle and in the lower and upper crust (Fig. 1).

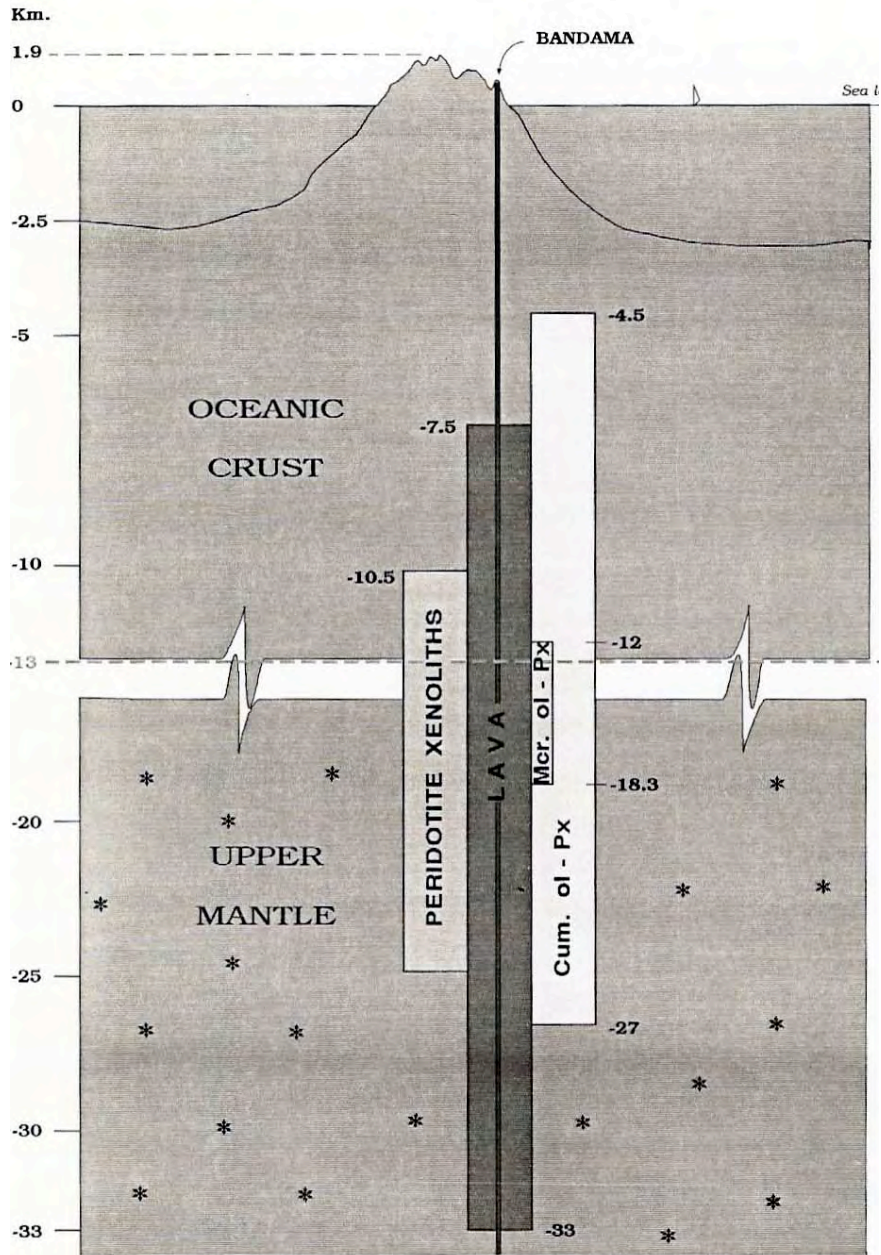


Figure 1. Interpretative scheme of depth calculated by CO₂-rich fluid inclusions trapped in olivine and pyroxene from lava flows, xenoliths, cumulates and megacrysts of the Bandama volcanic center (Gran Canaria, Canary Islands).

Discussion

The olivine, clinopyroxene and amphibole which make up the crystals (phenocrysts and megacrysts), cumulates and xenoliths of Bandama and La Caldera de Pinos de Galdar volcanoes display different textural, mineralogical, microthermometric and

chemical composition characteristics. Thus, the lava flows have basanite composition (ultramafic, low-evolved magma), cumulates and megacrysts were formed in more-evolved magmas (mafic and ultramafic) and the upper mantle xenoliths have variable peridotitic compositions and textures (ultramafic). In addition, numerous studied minerals reveal generally reaction rims with basanite melts, documenting prolonged magma contact (mineral diffusion zones). Therefore, we can conclude that these studied minerals were originated under variable magmatic conditions and chambers.

We propose a model of the multi-stage magma ascent and storage beneath Gran Canaria island (Fig. 1). Therefore, the basanitic magmas which gave rise to Bandama and Pinos de Galdar volcanoes ascended from upper mantle to the surface trapping xenoliths, cumulates and megacrysts of different depths and genesis. The basanite ascending magmas stagnated in several reservoirs en route to the surface between -33 km to 1,5 km, trapping ultramafic xenoliths, mafic and ultramafic cumulates, and olivine-pyroxene-amphibole megacrysts from the upper mantle and the upper oceanic crust (-27 to -4.5 km.).

VOLCANIC FEATURES OF THE CENTRAL ATLANTIC OCEAN: TECTONIC AND MAGMATIC MODELS

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Similar to other ocean basins, the central Atlantic Ocean (roughly latitude 10° to 50° N) contains many (more than a hundred) volcanic seamounts grouped in chains, clusters, and individual features. Although the Atlantic Ocean crust has been forming continuously from the Early Jurassic to the present with tholeiitic rift basalts, most of the seamounts were created during the Middle to Late Cretaceous as alkaline basaltic or bimodal volcanoes, some of which have continued activity into recent times.

Wide-ranging homogeneous upper (?) mantle horizons produced low-Ti, intermediate-Ti, and high-Ti basalts of the Central Atlantic Magmatic Province, which are apparently all between 201 and 198 Ma in age. CAMP magmas formed distinct but overlapping suites of sub-parallel dikes that are related to the geometry and extension of Pangaeian rifts. CAMP dikes do not define a co-magmatic swarm that radiates around a common geographic center, despite many such assertions in the literature. Basaltic magmas flowed obliquely within the dikes as they followed propagating fractures for hundreds of km, but they were not interconnected between suites and swarms. There is no evidence for domal uplift over a “plume center,” and there is no “tail trail” of Jurassic hotspot volcanoes or ocean ridge that proceeds away from any plume center. Rather than becoming dormant and then resurfacing much later with alkaline seamounts, it is more reasonable to believe that mantle convection responsible for CAMP basalts is related to the subsequent and continuing production of Atlantic Ocean crust. Basaltic melts evolved after Pangaeian rifting in the Early Jurassic, progressing from an early magma enriched by continental and older subduction-slab components into a relatively depleted ocean mantle melt.

In the Cretaceous after 125 Ma, geometrically discrete, chemically less-depleted sections of the deeper mantle produced pulses of alkaline magmas to feed numerous and widely-separated volcanoes, within the continent margins as well as the ocean. These magmatic events started when entire plates experienced major tectonic shifts during the Cretaceous, related to a change in plate motions as mantle convection and lithospheric rifting moved northward and southward into the Laurasia and Gondwana super-continents. Sub-lithospheric alkaline magmas collected during this event and moved rapidly to the surface along extensional fracture zones and fracture intersections, in the oceans as well as in adjacent continents within the same plates. The few linear seamount chains in the central Atlantic basin, which may or may not be time-progressive, are all linked to much older continental linear structures that propagated into the ocean crust.

The geographic patterns, compositions, and styles of igneous activity of the seamounts contrast greatly with ocean crust magmatism at the mid-ocean ridge. A

deep mantle plume cannot be responsible for both the initiation of Early Jurassic sea-floor spreading and the younger alkaline seamounts. All of the magmas are probably derived from the upper mantle and can be related to mantle convection, depth of melting, lithospheric structures, stress patterns, and plate tectonism.

Plume- Lithosphere Interactions: Cases of Afar (Africa), and Pacific Hotspots

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The origin of hotspots is usually related to the surface manifestation of deep mantle plumes, which are originating in thermal boundary layers. Three boundary layers can be considered: asthenosphere, transition zone, D''-layer. And it can be shown that there are different kind of plumes.

The detection of plumes in the mantle from geophysical and geochemical data is still controversial and trigger vigorous debates. It remains unclear how plumes are formed, their origin at depth, and whether they act independently from plate tectonics. We may learn about the role of plumes in mantle dynamics by studying their interactions with lithosphere and crust below ridges and the way in which they perturb the flow pattern in the uppermost mantle.

Several regional tomographic studies of seismic velocity and anisotropy around several hotspots were obtained during the last 2 years. Their lateral resolution is smaller than 1000km and they enable to make qualitative intercomparison between Afar (Horn of Africa Program), Azores (COSEA project) in the Atlantic, La Reunion in the Indian Ocean and Pacific provinces hotspots. These models demonstrate that there is not only one family of plumes but several ones. Some plumes are confined in the uppermost 200km but a few can originate in the transition zone and may be at the Core-mantle Boundary for superplumes.

Seismic anisotropy which is a good marker of deformation processes and mantle flow pattern, shows that the interaction between a plume and a ridge below the lithosphere can occur over distances larger than 1000km, via sublithospheric channels. The existence of LACs (Low Anisotropy Channels) below the Pacific plate seems to be intimately related to the active hotspots in Central Pacific and indicate a future reorganization of plate boundaries. Another important consequence of the interaction between plume, lithosphere and ridge is the triggering of secondary convection in the asthenosphere, which can also give rise to babyplumes. In conclusion, the concept of plume is not univocal but might refer to different geological objects.

Finite frequency tomography reveals a variety of plumes in the mantle.

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Our understanding of the Earth’s dynamics relies mainly on the degree of knowledge of the deep Earth structure. Seismic tomography is the only tool available to date able to map the three-dimensional structure of the Earth’s interior. It provides a snapshot of the present mantle convection. Even though there is a general agreement on the average, spherically symmetric structure of the Earth, the real fate of sinking plates as well as the origin and geometry of the upwelling regions are still subject of open debate. Hotspots are probably the most intriguing geophysical object. They are approximately fixed with respect to plate motion, providing us with an absolute reference frame. Morgan (1972) proposed that hotspots are due to plume-like upwelling from the lower mantle, but seismic tomography studies have been so far unable to clearly detect such deep plumes.

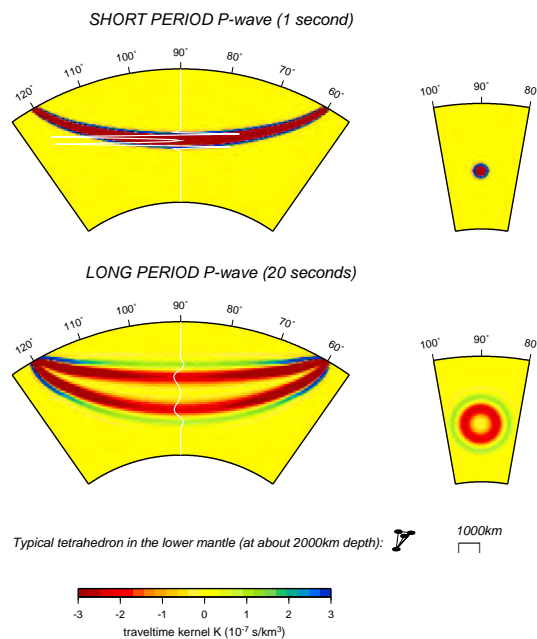


Figure 1: Sensitivity kernels for a direct P wave at an epicentral distance of 60° with two dominant period: 1 s (top), 20 s (bottom). The size of a tetrahedron in the lower mantle is shown for reference below the 20 s kernel.

a wave with 20 s dominant period. The sensitivity is significantly different. A broadband P traveltime is sensitive to anomalies in a hollow banana-shaped region surrounding the unperturbed path, with the sensitivity being zero on the ray. Because of the minimax nature, surface reflected PP waves show a

Our first global finite frequency tomography of compressional waves show distinct conduits rising from the deep mantle (Montelli et al., 2003) What was the problem then with the previous tomographic studies?

Almost all global P-wave tomographic models have been so far obtained by applying the approximation of ray theory. Waves propagate as rays only in the high-frequency limit of the elastodynamics equations of motion. The travel time is only influenced by the Earth’s properties along an infinitesimally narrow path that follows Snell’s law. This simplifies the mathematics, but it is quite far from the physical reality where rays have a given thickness depending on the frequency content of the propagated wavefield. The traveltime of a finite-frequency wave is sensitive to velocity structure off the geometrical ray within a volume known as the Fresnel zone. Classical ray theory predicts that even a small heterogeneity on the raypath would influence the traveltime. But physics teaches us that small scale objects do not really influence the propagation of waves. They only do when their scale length is comparable to the width of the Fresnel zone. Figure 1 shows the sensitivity region for a P wave at 60° epicentral distance. The top kernel is the sensitivity region of a wave with 1 s dominant period, while the bottom kernel represents the sensitivity of a wave with 20 s dominant period. The sensitivity is significantly different. A broadband P traveltime is sensitive to anomalies in a hollow banana-shaped region surrounding the unperturbed path, with the sensitivity being zero on the ray. Because of the minimax nature, surface reflected PP waves show a

much more complicated shape of the sensitivity region, with the *banana-doughnut* shape replaced by a *saddle-shaped* region upon passage of a caustic. Not surprisingly, the introduction of such complicated sensitivity has consequences for the final tomographic images. A small size heterogeneity would affect the 1 s wave arrival times but would not be seen in the travel time of the broadband P wave (however, it would influence the amplitudes). Mantle plumes are narrow and therefore the most affected by an inappropriate modeling of finite-frequency effects. Because of their size, plume tails could partially be hidden in the region of insensitivity around the unperturbed ray. Wavefront healing, neglected by classical ray theory, but properly accounted for in our finite-frequency modeling enhance the capability to detect such Earth's structures.

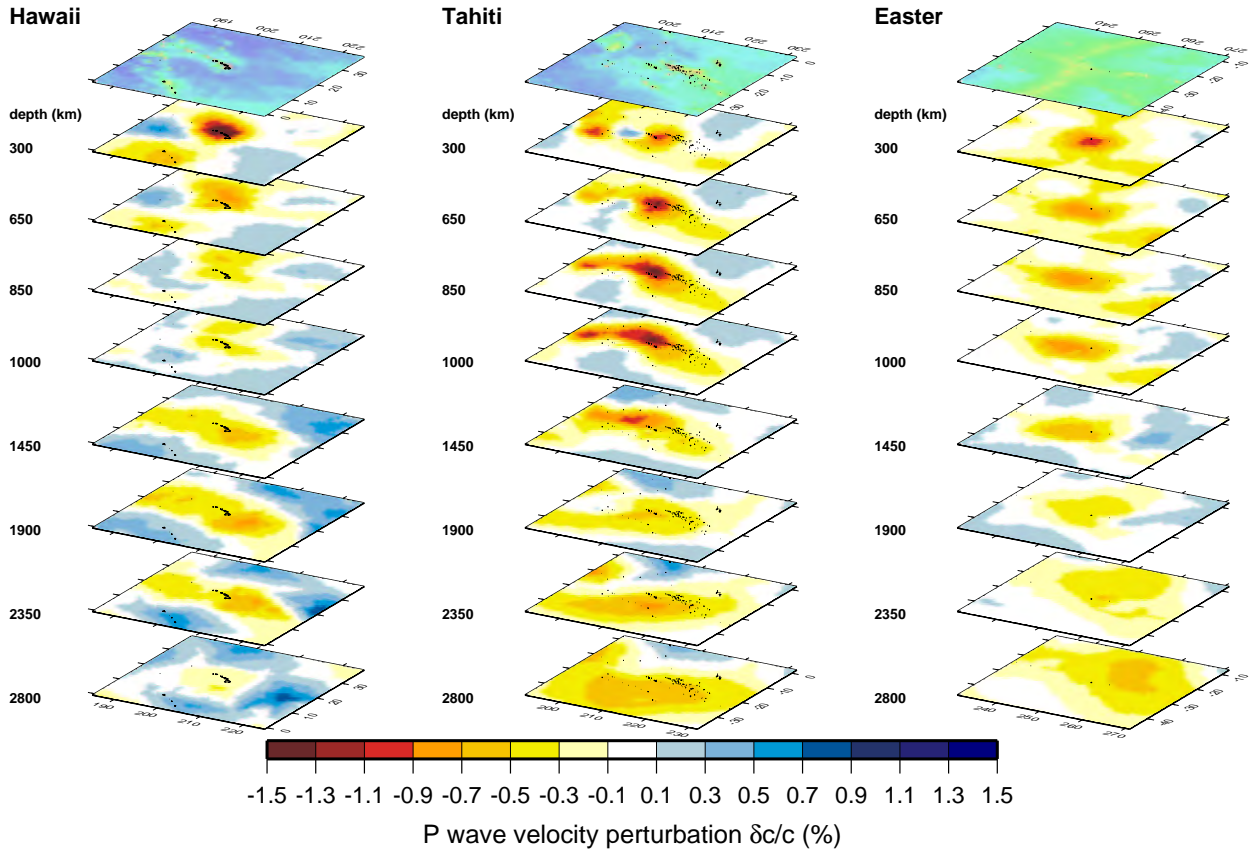


Figure 2. Cross section of the joint inversion (no pP arrivals included here) velocity model as a function of depth at different hotspot locations: from left to right Hawaii, Tahiti, Easter Island.

We present the results of an inversion of finite frequency P, PP and pP waves with a dominant period of 20 s, whose travel time sensitivity kernels are modeled by using the recently developed formalism derived by Dahlen et al. (2000); combined with short period P and pP extracted from the ISC data set (Engdahl et al. 1998) modeled by using standard ray theory. Inverting a combination of low and high frequency waves allows us to properly constrain long wavelength heterogeneity with the kernels, while using the high-frequency data to constrain smaller-scale structure. The velocity structure is sampled using an irregular distribution of points to form a Delaunay mesh (Watson 1981, Watson 1992, Sambridge et al. 1995). Node spacing is adapted to the expected resolving length of our data and ranges from about 200 km in the upper mantle to about 600 km in the lower mantle. This

flexibility gives an additional improvement in the tomographic images. As a result our tomographic images provide, for the first time, unambiguous evidence that at least 5 hotspots originate in the deep lower mantle: Hawaii, Easter Island, MacDonal, Samoa, Tahiti (Figure 2) and suggest that few others, such as Kerguelen and Cape Verde might be connected to the core-mantle boundary.

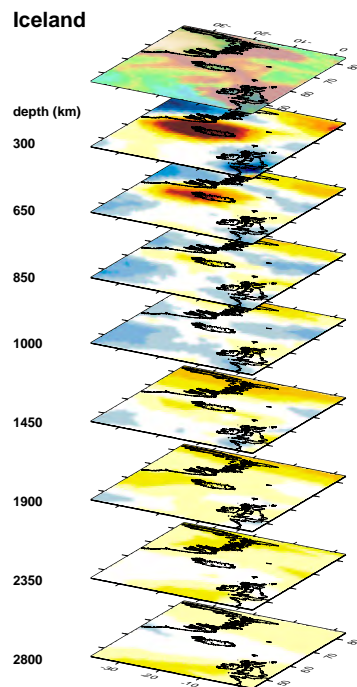


Figure 3: Cross section of the joint inversion velocity model as a function of depth below Iceland.

Major hotspots which do not seem connected to a deep lower mantle plume include Afar, Ascension, Galapagos, Kilimanjaro, Madeira, Reunion, Tristan. These all seem to originate in the mid mantle. Iceland seems of a shallow depth (Figure 3). Indications that Iceland is not a deep-rooted anomaly were already presented by Ritsema et al. (1999) and a shallow origin was argued from indirect evidence by Foulger & Pearson (2001), Foulger et al. (2001) and Foulger (2003). The results of our inversion confirms these observations and clearly contradict the finding of Bijwaard & Spakman (1999), who proposed a plume extending all the way to the core-mantle boundary.

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What *Really* Happened in the Pacific?

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The proper way to consider seamount volcanism on the Pacific plate is to examine all seamounts at once, as if the ocean were entirely drained away. This can be done by considering the modern chart of bathymetry derived from satellite-based altimetry (Smith and Sandwell, 1997), which was not available to Jason Morgan when he proposed mantle plumes (Morgan, 1971, 1972a, 1972b). This chart, based on a uniform remotely-sensed sampling of the entire plate, reveals thousands of seamounts in the Pacific arranged in bewildering arrays of alignments, non-alignments and orientations, and they are far from being regularly distributed. The Pacific plate contains huge plateaus without trailing seamounts; trains of seamounts that have no affiliated plateaus; trains that changed volume and rates of propagation through time; trains that contain curving, splayed, imbricate and cross-trend ridges; and trains that terminate at transform faults. It contains huge clusters of seamounts that are not trains at all. Only a few of them are grouped in the linear, concentric island chains so central to the plume hypothesis. The idea that linear island chains alone provide an adequate picture of mantle geodynamics beneath the plate thus must be seen as a misapprehension.

Extensive dredging of seamounts by many workers demonstrates that the geochemical category of “ocean-island basalt (OIB)” is misleading. Varieties of alkalic basalt or related differentiates occupy the summit of almost all seamounts that rise more than about 2-km above the seafloor, and islands themselves account only for the very small percentage of them that actually erupted in entire edifices. Even so-called E-MORB that erupt directly on ridge axes are merely alkalic olivine basalts very much like those of tall seamounts both near and away from spreading ridges (e.g., Engel et al., 1965; Batiza et al., 1989, 1990; Natland, 1989; Davis et al., 1995; Cousens, 1996; Niu et al., 2002; Niu and O’Hara, 2003). Mixing between depleted and enriched magma strains may explain much of the geochemical diversity of abyssal tholeiites. Islands are actually only the tips of the tallest of seamounts, thus if our interest were in deciphering the origins of “tall-seamount basalts” we would be nearer the mark than if we restrict our attention to the compositions, temporal sequences, and geometrical relationships of basalts that can be sampled only on foot. We should include all petrologically similar basalts in our purview, whether they erupted on islands or below sea level, young, old, tall or small seamounts, *or* spreading ridges. Since OIB-like lavas occur on so many features that are not age-progressive linear island chain, why should such geochemistry *require* a deep mantle plume anywhere else? What can we imagine the typical internal structure of plumes to be when some volcanoes within chains have no geochemical consistency from one to the next and the chains themselves have opposite geochemical trends through comparable stages of volcanism?

The Pacific plate, the largest on Earth, spans a history of more than 165 million years (e.g., Pringle, 1992), from an infancy when it was very small and entirely

surrounded by spreading ridges (Nakanishi, 1993). Beginning in the Jurassic, it grew in all directions until some of its edges necessarily intersected subduction boundaries during mid- and later Cretaceous times. Today this giant among plates is more than half bounded by subduction boundaries and linking transform faults. The current trend of the plate's motion is to the WNW relative to those subduction boundaries, as it has been for at least 47 Ma (Sharp and Clague, 1999).

The modern concept and acceptance of mantle plumes depends strongly on assumptions of parallelism (concentricity) of the traces of linear volcanic chains on the Pacific plate and their apparent fixity with respect to each other since 47 Ma, but only since that time. Prior to that, the picture is much more complicated. For example, a substantial body of radiometric-age data extending back to the mid-Cretaceous exists for the Line Islands seamounts, which Morgan (1971; 1972a, b) proposed to be coeval and concentrically age-progressive with the Emperors portion of the Hawaiian-Emperor chain. However, radiometric-age data summarized by Davis et al (2002) show NO age progression within the Line Islands seamounts (Figure 1). The northern end of the chain is also several times wider than the track of the Emperors, and has ridges and seamounts arranged in two principal orientations, NNW and WNW. Nearly synchronous volcanism occurred along much of the length of the chain at about 90 Ma and occurred again at about 70 Ma. The most recent volcanism was in the middle of the chain at about 37 Ma. The WNW Crosstrend ridges date from about 70 Ma (Davis et al., 2002), and are parallel to the much younger, non-hot spot Puka Puka Ridges west of the present-day East Pacific Rise (Sandwell et al., 1996; Lynch, 1999, Janney et al., 2000), and to the current direction of Pacific plate motion. Samples from these ridges include varieties of amphibole-bearing potassic mafic lava not found on emergent Pacific linear chains, but which are well known from, e.g., African rift valleys (Natland, 1976; Davis et al., 2002).

Early Cretaceous and Jurassic ocean crust, which extends from the Mid-Pacific Mountains to the western Pacific trenches, is so covered with large seamounts and their aprons (Menard, 1959) that an actual sample of ridge-related basalt has been difficult to obtain by drilling. Among these are hundreds of Mesozoic guyots capped with drowned reef platforms (e.g., Hamilton, 1956; Winterer et al., 1993; Haggerty and Premoli Silva, 1995). Based on isotope geochemistry, Koppers et al (in press) group these into several short seamount trains, but these are not age-progressive, and coeval trains dating from 130-90 Ma have sharply different trends. Lavas from these seamounts are mainly variably enriched alkalic basalts and related differentiates that are similar isotopically to lavas from the modern Polynesian linear island chains (Winterer et al., 1993; Koppers et al., in press).

The Mid-Pacific Mountains number several widely spread, very large, and elongate Mesozoic ridges that *splay* from ENE to E (Figure 1), thus which are not parallel. Such large, splayed ridges have no counterparts among the active linear chains of the Pacific, although less accentuated splaying occurs in the Tuamotus. Toward the west, seamount groups from the Wake to the Magellan seamounts have no clear or geographically persistent alignments. Instead, most are collections of large individual seamounts, guyots, and seamount-guyot clusters (Vogt and Smoot, 1986) that erupted throughout the same large region from 130-90 Ma (Winterer et al., 1993; Pringle et al., 1995; Koppers et al., in press). No similar congregation of large, isolated, and scattered volcanoes has formed anywhere else on the Pacific plate since that time.

We infer that orientations and distributions of seamount ridges on the Pacific plate are to first order controlled by tectonic stresses acting across the plate (Jackson and Shaw, 1975; Favela and Anderson, 2000; Smith, in press), and that volcanic alignments occur along fractures that are generally orthogonal to the direction of least principal stress (e.g., Fiske and Jackson, 1972; Nakamura, 1978). Since spreading ridges change orientation and geometry in response to consumption of ridge axes at trenches by migration of triple junctions along continental margins (e.g., Lonsdale, 1991), we assume that the stress field within a plate must change in complementary fashion, and that changes in directions of least principal stress will guide changes in patterns of lithospheric fracture that allow seamount volcanism to occur (Hieronymus and Bercovici, 2000, 2001). Major bends in linear volcanic chains reflect changes in the balance of stress orientations, not wholesale changes in plate motion, and splaying of ridges indicates either a non-uniform stress regime, or one that rotated through time (cf. Carey, 1958).

We thus construe three general periods with different stress regimes in the history of the Pacific plate.

1) When, during its earliest days in the Jurassic and Early Cretaceous, the Pacific plate was surrounded by ridge segments and was near the center of the huge world ocean, there were no major stress alignments within it. In this respect, it was much like that of the present-day Antarctic and African plates (Hamilton, 2002). Within-plate volcanism thus assumed the scattered arrangement predicted by the models of Hieronymus and Bercovici (2000) for the condition of *no tectonic stress*, and the large Magellan and Wake seamount clusters formed. Nonetheless, near the eastern boundaries of the plate, which were marked by migrating triple junctions, and which adjoined plates that were probably disappearing into subduction zones, complex and shifting patterns of ridge reorganization dictated formation of long, splayed, near-axis seamount ridges.

2) By about 90 Ma, the growing middle-aged Pacific plate achieved its first persistent stress regime with the formation of subduction boundaries along its western or northwestern margin. The plate was no longer static but began to move over the asthenosphere and into the mantle. The precise arrangement of those subduction boundaries and the overall direction of subduction are uncertain, but this imparted a general yet not fully stable component of tension across the plate. This stress combined with others produced the initial NNW Gilbert-Marshall, Musician, Line and Emperor Seamount ridges, orthogonal to the overall direction of least principal stress, and which still could vary somewhat from place to place. The Line Island seamount chain, being near ridge axes, thence to plates subducting into trenches in the eastern Pacific, sustained a more variable stress regime, thus its great width and dual orientations of ridges.

3) By 47 Ma, the Pacific plate was huge, and the Gondwanan dispersal of southern continents was shifting into a pattern of major continental collision. Plates east of the East Pacific Rise axis were growing smaller as the approaching Americas rolled them back at trenches. Nearly half of the boundaries of the Pacific plate now were also trenches spanning from the Aleutians to New Zealand. In addition, northward migration of the Indian plate and Australia caught a major portion of the westerly moving Pacific plate between the northeast corner of the Tonga Trench and the Aleutians. The plate could no longer shift laterally in response to whatever was occurring along its eastern spreading boundaries. A very consistent and possibly stronger stress regime therefore

developed across the Pacific plate with a NNE direction of least principal stress. The change in stress orientation may have taken up to 10 million years, during an interval marked by little or no volcanic productivity at the western end of the Hawaiian chain. Since that time, the predominant alignment of both linear island chains and Puka Puka-type ridges, from the Kodiak-Bowie chain in the Gulf of Alaska to the Louisville Ridge south of the Antarctic convergence, has been orthogonal to this direction. The expression of this stress regime nearest the Tonga Trench is the voluminous, elongate, WNW-trending, post-erosional volcanic sequence in the Samoan Islands, which is along the tensional crest of a bend in the plate dipping toward the transform portion of the trench (Natland, 1980).

Three additional hypotheses seem necessary to explain seamount volcanism on the Pacific plate. The first is that enriched mantle sources of variable geochemical provenance are distributed in a shallow layer at the base of the lithosphere. This geometry is required wherever volcanism has occurred simultaneously along very long ridges, as, e.g., the Samoan post-erosional volcanic rift zone (Natland, 1980), or the Puka Puka ridges in the eastern Pacific (Janney et al., 1999). Perhaps dispersed veins, schlieren, or blobs of enriched material, such as have been invoked to explain alkalic basaltic summits of near-ridge seamounts in the northeastern Pacific (Cousens, 1996), and that are readily subject to partial melting, produce melts that migrate and accrue at the base of aging and thickening lithosphere (e.g., Anderson, 1989, 1995; Niu et al., 2002). They may concentrate there because of shear dilatancy (Holtzman et al., 2003) at the base of the nearly impermeable lithosphere. The magmas are later tapped when regional stresses cause that lithosphere to fracture. The geometry of a zoned plume may instead be represented by lateral or vertical heterogeneity within a layered mantle. Enriched material attached to the base of the lithosphere may later become involved with volcanism a great distance away. This hypothesis says nothing about the ultimate depth of origin of the enriched material, the ages of its diverse isotopic components, or how they were transported vertically, only that it accrues through time at the base of the lithosphere. Isotopic similarities between Mesozoic seamounts of the far western Pacific to their likely backtracked locations near the modern chains of the Southwest Pacific (Staudigel et al., 1991; Koppers et al., in press) suggest persistence of this process in the same part of the upper mantle for more than 130 million years.

The second hypothesis is that some concatenation of stresses is likely required to localize the modern linear island chains, as for example plate-bending near the Samoan Islands. Although the general pattern of stresses acting across the Pacific plate appears to have a consistent orientation, the stresses sum from all boundaries of the plate, thus their magnitude within the plate cannot be uniform. The underlying asthenosphere itself could also bulge locally in response to plate-tectonic stresses, and develop concentrations of partial melt at the base of the lithosphere, where most basalts erupted on island chains appear to originate. The lithosphere may contain internal zones of weakness inherited from prior stress fields or local concentrations of more readily fusible and thus weaker materials. These will act to concentrate stress and initiate fractures (Lawn, 1993). The lithosphere will contain regions of great thickness and strength that will stop propagating fractures in their track or deflect them. Local hydraulic overpressure resulting from inequities in the distribution of ponded magma and magma buoyancy may drive fractures from underneath. Transfer of material from the base to the top of the lithosphere will

modify the stress field and produce new stress concentrations that will become preferred locations of new volcanism (e.g., Hieronymus and Bercovici, 2001). Nevertheless, the general effect of a strongly directional stress field, one that developed through time on the Pacific plate, will be an equilibrium tendency (not always ideally satisfied) for parallel fractures to develop and to propagate in tandem at the rate of plate motion over the asthenosphere. On the other hand, once the lithosphere is fractured, the resulting lines of weakness should easily be reactivated by shifts in the stress regime, as in the Line Island seamounts (Davis et al., 2002) and the younger Cook-Austral chain (McNutt et al., 1997).

Finally, the large sizes of Pacific plateaus and some seamounts, seamount ridges, and linear island chains, require one or more of three things: 1) concentrations of mantle with high fertility, which is fundamentally an aspect of heterogeneity of the bulk composition of the mantle (e.g., presence of fertile peridotite, garnet pyroxenite, and/or eclogite); 2) differences in the size of master fractures through the lithosphere, effectively determining the ease or efficiency with which melt can rise through it – a valve effect; or 3) locally more vigorous convective turnover of mantle beneath the plate; the basalt-releasing conveyor belt moves faster. The latter, of course, is presumed in the plume hypothesis, but usually without consideration of whether 1 or 2 might be important. Even so, any such turnover in the upper mantle need not involve deep-mantle material (Sandwell et al., 1996).

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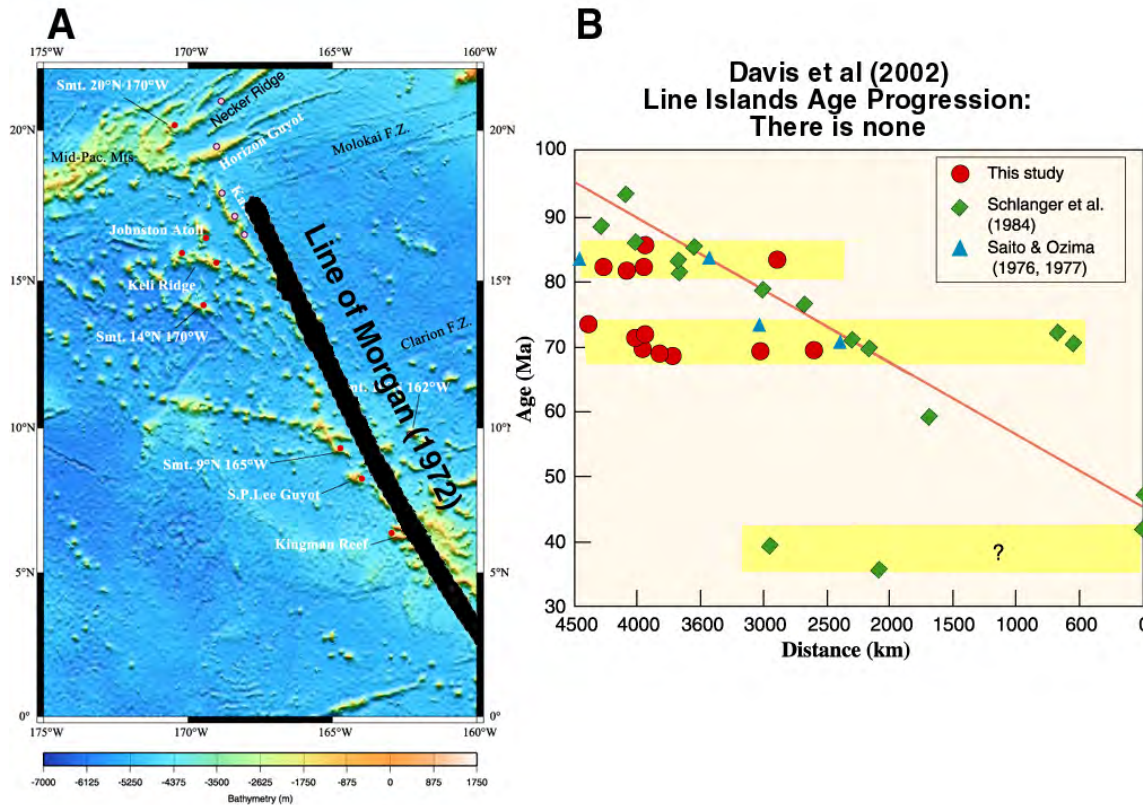


Figure 1

A. Bathymetry and B. age progression of the Line Islands seamounts, from Davis et al (2002). Bathymetry is taken from Smith and Sandwell (1997). Locations of 68-73 Ma volcanism are red dots; locations of 81-86 Ma activity are pink dots. The black line is the hypothetical age-progressive trend line of Morgan (1972a, b) paralleling the Emperor Seamounts. The line was selected from a scanned image of his figure, and is plotted here to the scale shown. The principal Line seamount trend and Line cross-trend ridges are evident in the bathymetry, as is the pattern of ridge splaying of the Mid-Pacific Mountains to the north. Ages in B, from sources cited in the figure, are plotted against distance from the Line-Tuamotu bend. “The diagonal red line represents rate of volcanic propagation (9.6 ± 0.4 cm/yr) proposed by Schlanger et al (1984) as evidence for a hot spot trace. New ages indicate two major episodes of volcanism more than 10 million years apart. Ages of Schlanger et al (1984) from the southern Line Islands suggest another episode of volcanism ~ 40 Ma) in this region.” (Davis et al., 2002, p. 17, caption to their Figure 8).

What's going on at Iceland?

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Icelandic volcanic rocks present the peculiar geochemical feature that the extent of geochemical enrichment deduced either from trace elements or isotopes, and conventionally attributed to mantle source heterogeneity, correlates with the degree of differentiation of the rocks. Whereas at most other island chains, ranges in extent of depletion or enrichment are grounded in compositions of primitive basaltic lavas (olivine tholeiite, alkalic olivine basalt, basanite, olivine nephelinite) that are interpreted to be little removed in composition from direct and successively smaller partial melts of the mantle, the full isotopic variability at Iceland occurs within tholeiitic basalts and associated differentiates (see **Iceland Petrologic Conundrums Figure 1, below**). Not only this, but the average differentiated basalt, usually a ferrobasalt, is more enriched than primitive olivine tholeiite, and the typical silicic eruptive – usually rhyolite – is often more enriched than ferrobasalt. This should not be the case if, as usually thought, all these lava types belong to a common, shallow, tholeiitic liquid line of descent. Instead, the general case at Iceland appears to be that differentiated lavas are *not* so related to primitive basalt, even at the same eruptive centers, even though almost all belong to tholeiitic magmatic lineages.

Recent geochemical studies also indicate that the most primitive Icelandic tholeiites derive from mantle sources that include recycled ocean crust in their bulk composition (Chauvel et al., 1999; Breddam, 2002). Many of the most nearly picritic Icelandic tholeiites are like this, and have geochemical attributes usually ascribed to a “depleted plume component” in the source region (Fitton et al., 1996; 2003; Kempton et al., 2000). The general notion is that ancient ocean crust descends through the entire mantle following subduction, and then is much later entrained by ascending buoyant material derived from near the core-mantle boundary and transported back up through nearly the entire mantle before contributing to partial melting (Halliday, 2002). Enriched Icelandic basalts thus are derived, ultimately, from different but also deep mantle source components entrained in the plume.

However, based on detailed comparisons to drilled abyssal gabbro (Natland and Dick, 2001; 2002) and the variability of basalt compositions along other spreading ridges, I now propose instead that almost all Icelandic volcanic rocks are derived by differing extents of partial melting of eclogite derived from lithologically variable recycled ocean crust that includes basalts, dikes and abyssal plutonic rocks as diverse as troctolite, olivine gabbro, gabbronorite, oxide ferrogabbro, and trondhjemite. Icelandic picrites are derived in the main by large-scale or even entire remelting of abyssal gabbro cumulates, and thus have higher Nb/Zr, Nb/Y, and Nb/U than typical MORB, but MgO contents of <12%, a value typical of abyssal olivine gabbro; ferrobasalts are derived from less extensive melting that selectively extracts constituents of oxide gabbro and gabbronorite from the recycled crustal assemblage; silicic rocks are comparable to the small-degree partial melts of eclogite that have been produced in experimental studies and to tonalite/trondhjemite veins in abyssal gabbros. Alternatively, they may have been produced by extended crystallization differentiation in the thick Icelandic crust, by low-pressure silicate liquid immiscibility, or by partial fusion of varieties of old, deep continental crust stranded in small masses beneath Iceland. Wherever

these silicic materials are in the crust or upper mantle, their high concentrations of, e.g., Rb, Th, and U, predict local radiogenic ingrowth that is preferentially extracted and concentrated in small-degree partial melts. Consequently, rhyolite occupies the position of the geochemically most-enriched volcanic material at Iceland, having Sr-, Nd-, and Pb isotopic characteristics elsewhere assigned either to a FOZO or a HiMu mantle component (see **Iceland Petrologic Conundrums, Figure 2, below**). Mixing with or selective assimilation of rhyolite by primitive basalt, which is well documented in many field associations on Iceland, accounts for the variable degrees of enrichment of the basalts. Taking into account the effects of mixing with rhyolite, the residual isotopic heterogeneity of Icelandic basalts is little different from, and no greater than, that of typical basaltic associations combining N- and E-MORB elsewhere along spreading ridges. On the East Pacific Rise between Clipperton and Siqueiros Fracture Zones, E-MORB ($K_2O = 0.5-1.1\%$) comprises about 6% of basalt erupted both along the ridge axis and on nearby seamounts (based on analyses in the Lamont Petrology Data Base), thus should be present in a similar proportion in both extrusive and intrusive portions of subducted masses of ocean crust. Preferential extraction of such material transformed to eclogite may explain some of the general isotopic and trace-element enrichments of Icelandic basalt compared with typical depleted MORB.

Most basalt of Reykjanes and Kolbeinsey Ridges, respectively south and north of Iceland, has parental characteristics (low Na_8 , low Ti_8 , high Fe_8) indicating greater extents of partial melting of mantle peridotite than that of typical MORB. A difficulty is that Icelandic basalts themselves have *higher* average Na_8 than those of the adjacent submarine ridges (see **Iceland Petrologic Conundrums, Figure 3, below**), something not predicted by melt-column models for a place with thicker crust but the same mantle materials beneath. The low Na_8 , Ti_8 , and generally low incompatible trace-element concentrations of basalts from Reykjanes and Kolbeinsey Ridges alternatively suggest partial melting of a more refractory (more nearly harzburgitic) mantle source than for typical MORB. Concentrations of incompatible trace elements and isotopic compositions shift gradationally toward Icelandic compositions as the ridges shoal near Iceland, indicating transitions in both directions from predominantly peridotitic to predominantly eclogitic source materials.

Following Foulger et al (2002 and in press), a source for Icelandic basalts in ocean crust of Caledonian age trapped in the Iapetus suture and since re-exposed by continental rifting can account for all of the principal petrologic and geochemical variability of Icelandic volcanic rocks, and their non-exceptional eruptive temperatures, without recourse to a deep mantle plume. Both higher average extents of melting of such readily fused material compared with peridotite, and the steep dip of the crustal material in the fossil suture contribute to the thicker crust at Iceland than elsewhere along the Mid-Atlantic Ridge. The sutured eclogite was trapped between two continental Archaean cratonic keels, the mantle of which is characterized by more refractory peridotite than that beneath most spreading ridges. Detached or abandoned remnants of this ancient peridotite are what now contribute to the distinctive petrologic characteristics of basalts from Reykjanes and Kolbeinsey Ridges, including their relatively high $^3He/^4He$ ratios ($R_A = 10-15$). Even higher $^3He/^4He$ among some (not all) primitive Icelandic tholeiites does not correlate with other isotopic or trace-element attributes of those basalts, thus it is not tied to basalt-rhyolite mixing. Its origin remains enigmatic, but it may derive from rupture of fluid inclusions in olivine-rich cumulates (cf. Natland, 2003) present either in the recycled eclogitic crust, its adjacent trapped abyssal peridotite, or more likely the ancient cratonic mantle that enclosed these

rocks from both sides. The high R_A is indicative of the age of lithospheric dunite, not a deep mantle reservoir. The mantle source beneath Iceland is zoned, but not because of a plume.

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Icelandic Petrologic Conundrums Figures

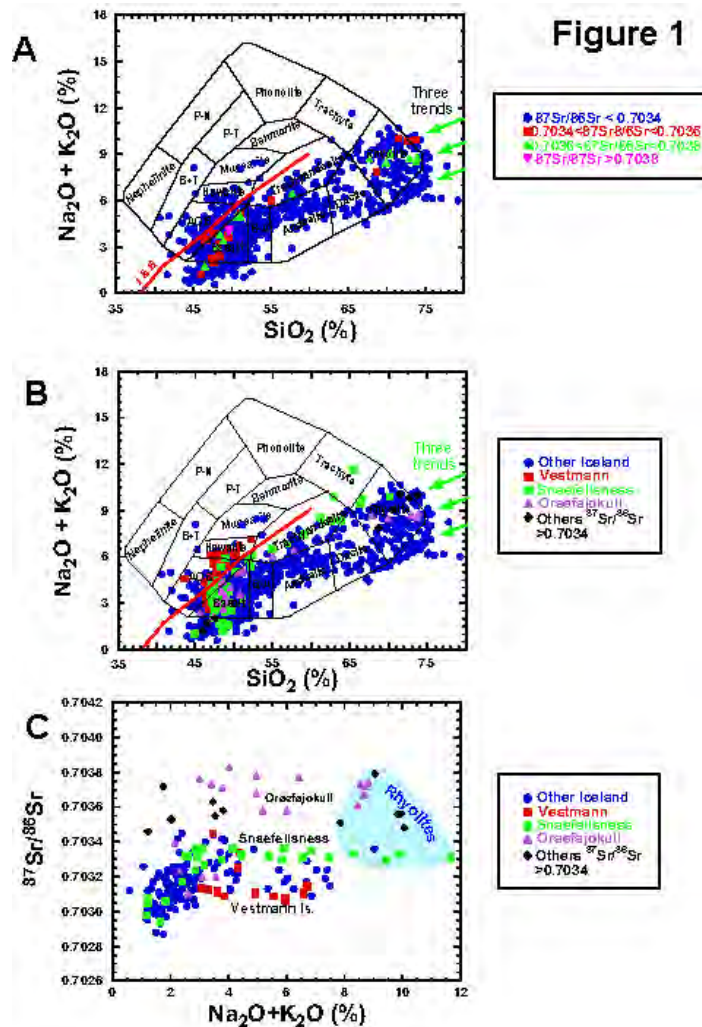


Figure 1

Conundrum 1: The isotopically most enriched lavas from Iceland belong mainly to tholeiitic, not alkalic, differentiation sequences, and the most enriched of all are rhyolites.

A. Alkalis-silica classification diagram (Cox et al., 1979) for Icelandic lavas compiled from the GeoRoc data base, plus more recently published data. The transition line of Irvine and Baragar (1971) between alkalic basalts (plotting above the line) and tholeiites (plotting below) is in red. Almost all rocks are tholeiitic-series basalts and differentiates. Samples for which $^{87}\text{Sr}/^{86}\text{Sr}$ has been measured are shown with different colors, as given in the key. Almost all, including those with highest $^{87}\text{Sr}/^{86}\text{Sr}$, are tholeiites.

B. alkalis-silica diagram for the same Icelandic lavas plotted in A but with data for mildly alkalic off-axis locales (Vestmann Islands, Snaefellsness, and Oraefajokull) shown using different colors, as given in the key. Although most basalts from these places are mildly more alkalic than from elsewhere in Iceland, they include some tholeiites.

C. Total alkalis versus $^{87}\text{Sr}/^{86}\text{Sr}$ showing that samples with highest $^{87}\text{Sr}/^{86}\text{Sr}$ are rhyolites and intermediate lavas with high total alkalis, whereas tholeiitic basalts (mainly with $\text{Na}_2\text{O} + \text{K}_2\text{O} < 4\%$). The most enriched lavas from off-axis localities, indicated in the key, are intermediate differentiates and rhyolites. Note the consistent yet distinctive isotopic values for each off-axis locale.

Interpretation. Intermediate compositions are basalt-rhyolite hybrids, and are dominated by the isotopic signature of rhyolite. Mixing, documented by many workers, occurs in the shallow Icelandic crust.

More primitive basalts are variably contaminated with rhyolite, thus their MANTLE isotopic signature is probably restricted to values of $^{87}\text{Sr}/^{86}\text{Sr}$ below those of the Vestmann Islands (<0.7031), and may be no higher than for typical MORB.

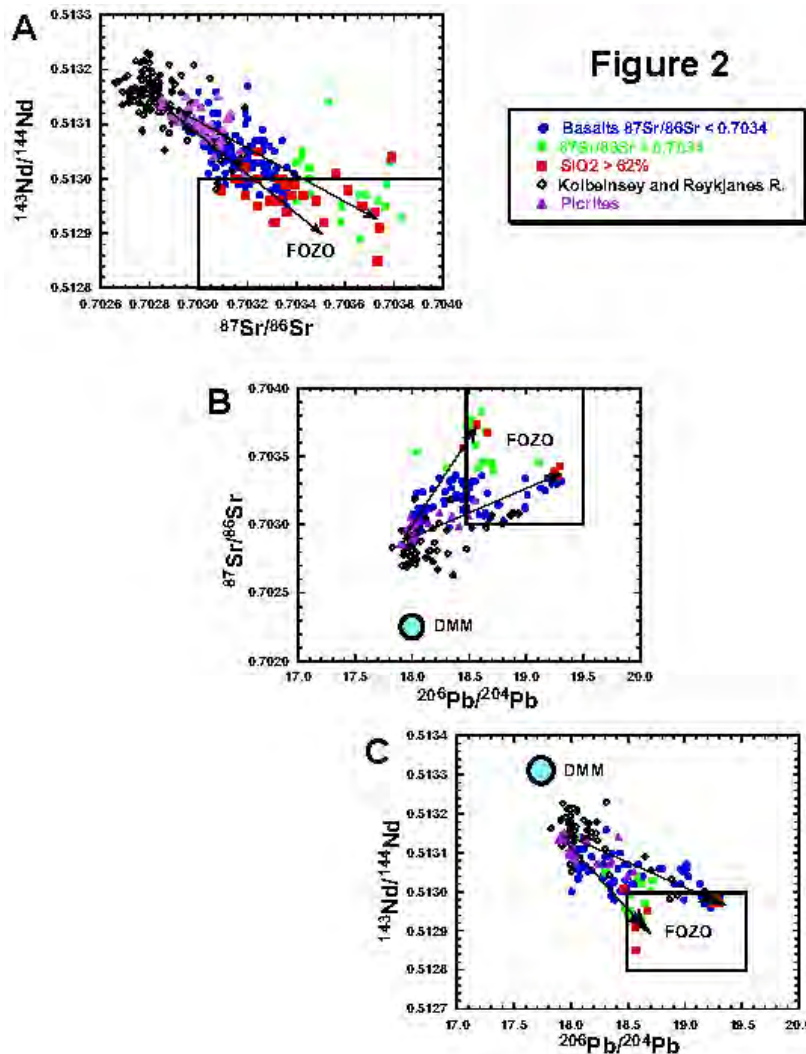


Figure 2.

Conundrum 2. The “enriched component” at Iceland, with FOZO characteristics, is rhyolite and rhyolite-contaminated basalt, not primitive mantle-derived basalt.

Some isotopic systematics of basalts from Iceland, Reykjanes Ridge and Kolbeinsey Ridge. A. $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{143}\text{Nd}/^{144}\text{Nd}$. B. $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$. C. $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$. Data are from GeoRoc and the Lamont Petrology Data Base (PetDb) plus some more recently published. Locations of depleted MORB mantle (DMM) and FOZO are from Bell and Tilton (2002). See keys for symbol explanations. On all diagrams Icelandic dacites and rhyolites ($\text{SiO}_2 > 62\%$) plus other lithologies with $^{87}\text{Sr}/^{86}\text{Sr} > 0.7034$ occupy the region of FOZO.

Interpretation. Icelandic basalts and many from Reykjanes and Kolbeinsey Ridges fall mainly along two potential mixing lines (with arrows) between rhyolite and Icelandic picrite, the least contaminated among the latter having isotopic values approaching those of depleted N-MORB.

Figure 3

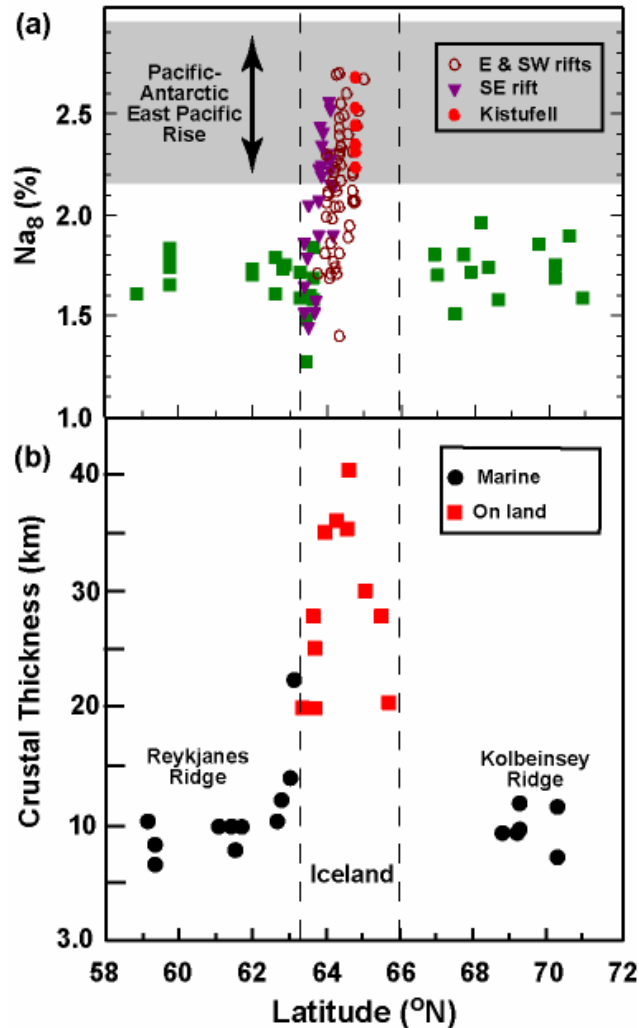


Figure 3

Conundrum 3. Melt-column models (e.g., Klein and Langmuir, 1987; Langmuir et al., 1992) inversely correlate Na_8 with crustal thickness. Iceland contradicts this. Values of Na_8 diverge at Iceland from those of adjacent ridges, but on the average increase, precisely where crustal thickness most sharply increases beneath the northern and southern Iceland margins.

(a) From Foulger et al (ms). Parental soda (Na_8) in basalt glass v. latitude. Data from Kolbeinsey and Reykjanes Ridges (green squares) are from the Lamont petrological database (PetDB). Icelandic compositions (see key) are from Breddam (2002) and Meyer et al. (1985). The range for the Pacific-Antarctic East Pacific Rise is from Castillo et al. (1998). (b) Crustal thickness vs. latitude, with data from a compilation of seismic experiments in Iceland and the North Atlantic.

Interpretation. Melt-column models invoking a homogeneous mantle source do not apply to Iceland. The mantle source is significantly more fertile with respect to a basaltic melt fraction beneath Iceland than beneath the adjoining ridges. The model suggested by Foulger et al (ms) is that the extra crustal thickness is provided by extensive melting of basaltic and gabbroic ocean crust that was once caught in the Caledonian suture, which persists beneath Iceland in the eclogite facies. Refractory peridotitic sources probably including material once in subcontinental lithosphere supply most of the parental basaltic magma beneath Reykjanes and Kolbeinsey Ridges.

“Plume-Ridge Interactions” as a Consequence of Ridge Suction

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Geological processes are consequences of Earth’s thermal evolution. Plate tectonics, which is driven by the cooling lithospheric plates atop the mantle, explains geological phenomena along plate boundaries such as volcanisms and earthquakes at ocean ridges and within subduction zones. Mantle plumes, which are considered to result from cooling of the earth’s deep interior, explain geological phenomena taking place away from plate boundaries such as intra-plate volcanisms, seamount chains etc. In this context, we may say that mantle plumes and plate tectonics are genetically independent of each other. However, when the ascending plumes approach lithospheric plates, interactions between the two inevitably result. Such interactions are most prominent near ocean ridges where the lithosphere is thin and the effect of plumes is best revealed. “Plume-ridge interaction” has been a hot topic in recent years, and much effort has been expended in this area [1-22]. While all these existing models differ in detail, they have several common assumptions: (1) mantle plumes are deep-sourced and necessarily hot, hotter than MORB mantle; (2) plume materials, relatively to MORB source, are enriched in volatiles and incompatible elements; (3) plumes dynamically “invade” MORB mantle; and (4) dispersion of plume materials in MORB mantle or mixing between the two distinct singular materials prior to or during melting gives rise to the “geochemical mixing” in erupted lavas.

Here we offer some new perspectives on plume-ridge interactions: (1) plumes and MORB source share a common two-component mantle [19,23-25]: *Easy-to-melt* or *Enriched dikes/veins (E)* dispersed in the *Difficult-to-melt* or *Depleted peridotitic matrix (D)*. (2) Plumes are likely to have variably greater *E/D* than MORB source whose *E/D* varies as well [19,24-28]. (3) Not all plumes necessarily come from excessively hot deep interiors although plume materials could well be intrinsically hotter because of greater abundances of heat producing elements (e.g., Th, U, K), depending on both the history and size of the “plume material domains”. (4) While ocean ridges are mostly passive features in terms of plate tectonics, they play an active, *NOT passive*, role in the context of plume-ridge interactions. This active role is simply a ridge suction force that drives asthenospheric mantle flow towards ridges because of material needs to form the ocean crust at ridges and lithospheric mantle in the vicinity of ridges: > 99.9% ocean crust is formed at ridges, and ~ 50% total thickness of oceanic lithosphere is created in the first ~ 17.5 m.y. (i.e., $t_{1/2} = [0.5 \cdot 70^{1/2}]^2$, assuming an oceanic plate reaches its full thickness, say, ~ 95 km, after ~ 70 Ma [29]). This ridge suction force *must* increase with increasing plate separation rate because of increased material demand per unit time [19]. For example, in the first one m.y., the mass flux towards the ridge to form the crust (assuming 5 km thick for simplicity) and lithospheric mantle due to heat loss per unit along-ridge length (km) is Φ (km^{-3}/Ma) = $25R_{1/2}$ (where the proportionality 25 has the unit of km^2 , and $R_{1/2}$ is half-spreading rate, km/Ma). As the seismic low-velocity zone atop the asthenosphere has the lowest viscosity that increases rapidly with depth [30,31], the ridge-ward asthenospheric flow is largely horizontal beneath the lithosphere in the direction against the motion of the overlying plate. It follows that the asthenospheric flow is necessarily decoupled from its overlying oceanic lithospheric plate, and the degree of the decoupling increases with increasing spreading rate (Figure 1).

The above concept is fully supported by observations. (1) The systematic lava compositional variations along the Foundation hotline towards the Pacific-Antarctic Ridge result from progressive decompression melting of ridge-ward flowing plume materials (Figure 2). (2) The similar geochemical observations in lavas along Easter Seamount lavas from Salas-y-Gomez Islands to the Easter Microplate result from the same process. (3) The increasing ridge suction force with increasing spreading rate explains why the Iceland plume has asymmetric effects on its neighboring ridges: both topographic and geochemical anomalies extend < 400 km along the slower (20 to 13 mm/yr northward) spreading South Kolbeinsey Ridge, but > 1500 km along the faster (20 to 25 mm/yr southward) spreading Reykjanes Ridge. (4) The spreading-rate dependent ridge suction force also explains the first-order differences between the fast-spreading East Pacific Rise

(EPR) and the slow-spreading Mid-Atlantic Ridge (MAR). Identified mantle plumes/hotspots are abundant near the MAR (e.g., Iceland, Azores, Ascension, Tristan, Gough, Shona and Bouvet), but rare along the entire EPR (notably, the Easter hotspot at $\sim 27^\circ\text{S}$ on the Nazca plate). Such apparent unequal hotspot distribution would allow a prediction of more enriched MORB at the MAR than at the EPR. However, the mean compositions between MAR-MORB and EPR-MORB are similar in terms of incompatible element abundances, and are identical in terms of Sr-Nd-Pb isotopic ratios (Figure 3). This suggests similar extents of mantle plume contributions to EPR and MAR MORB. We consider that the apparent rarity of near-EPR plumes/hotspots results from fast spreading. The fast spreading creates large ridge suction forces that do not allow the development of surface expressions of mantle plumes as such, but draw plume materials to a broad zone of sub-ridge upwelling, giving rise to random distribution of abundant enriched MORB and elevated and smooth axial topography along the EPR (vs. MAR).

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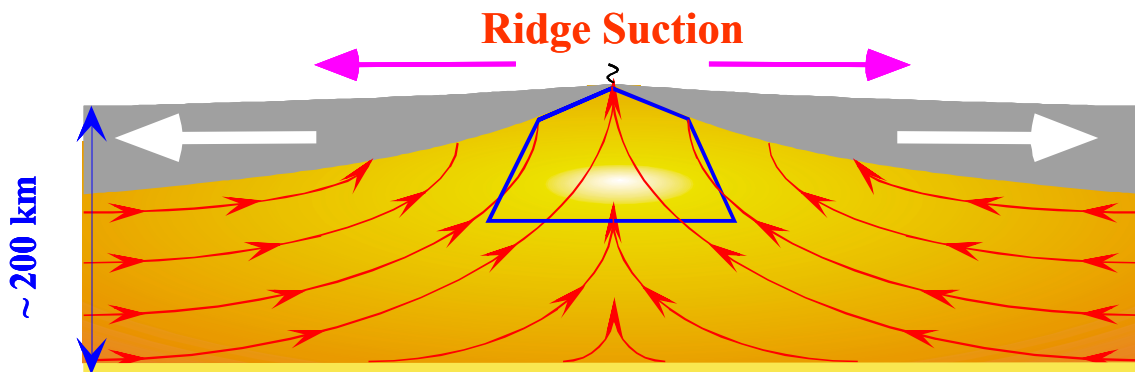


Figure 1. The regions of asthenosphere beneath ocean ridges have the lowest pressure in the entire asthenospheric mantle that drive asthenospheric flows (i.e., ridge suction). This suggests that the spreading lithospheric plates are necessarily decoupled from the sublithospheric flow. This is an important concept needed to understand mantle flows in the context of plate tectonics. The actually “affected” depth or depth range is unconstrained, but probably coincides with the seismic low-velocity zone, say, down to 200-250 km. The outline in blue represents region of decompression melting for MORB.

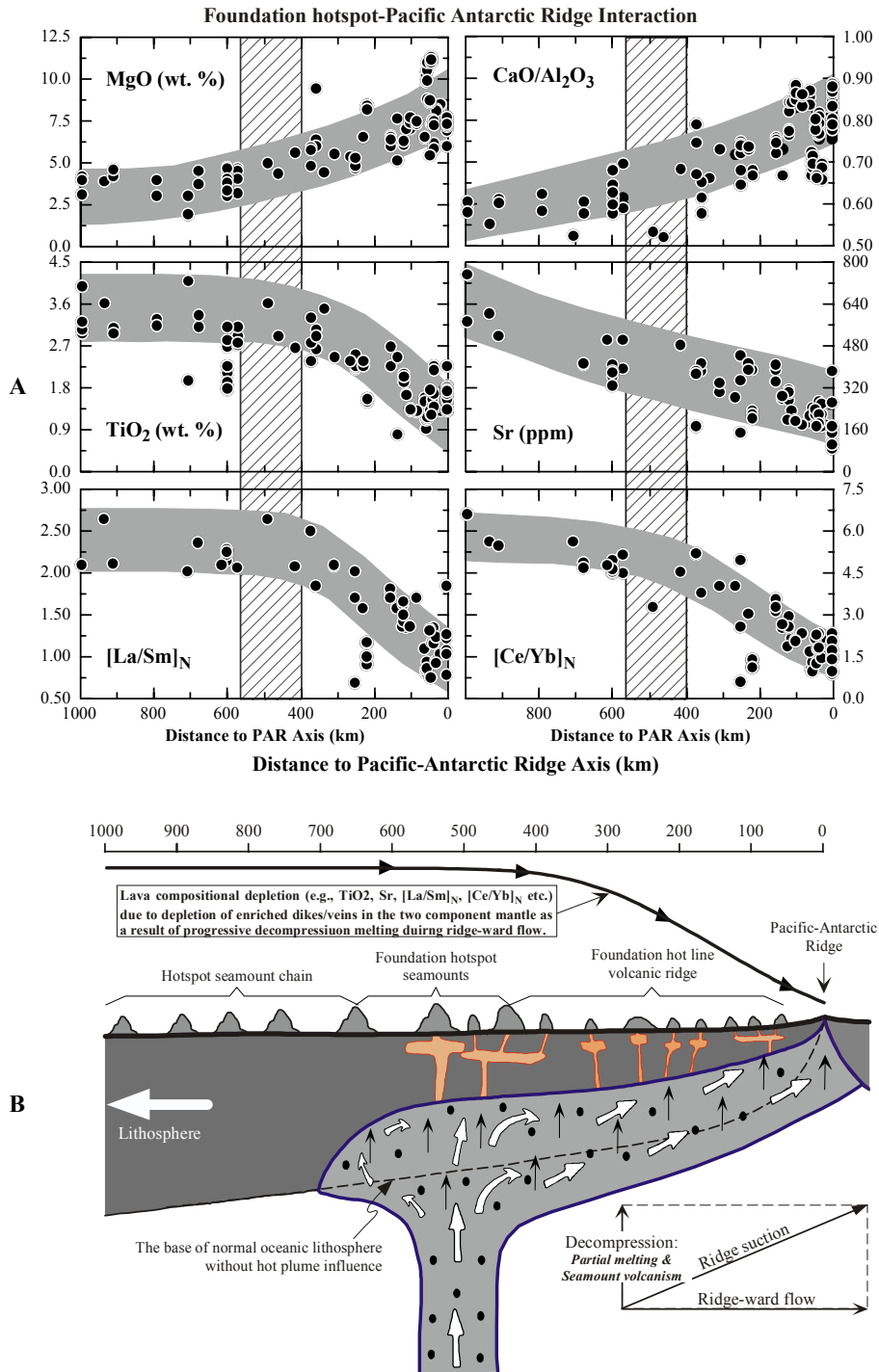


Figure 2. *A*. Lava geochemical systematics along the Foundation hotline volcanic ridges (~ 37°S) as a function of distance to the Pacific Antarctic Ridge. While scattered, the most data define systematic trends as highlighted by the shaded bands. *B*. Cartoon illustrating that the observed lava geochemical variation is the consequence of progressively melting a two-component plume material. Ridge suction requires the “hot” and “wet” plume material to flow towards the ridge with an upwelling component that causes decompression melting of the flowing plume material. The enriched dikes/veins (*E*) with low solidus temperatures are progressively depleted during the ridge-ward flow, thus leading to progressive melting of the more depleted matrix (*D*), and producing more depleted lavas towards the ridge. The geochemical data are from [16,17] with highly evolved samples (andesites, dacites and rhyolites having SiO₂ > 55 wt. %) excluded.

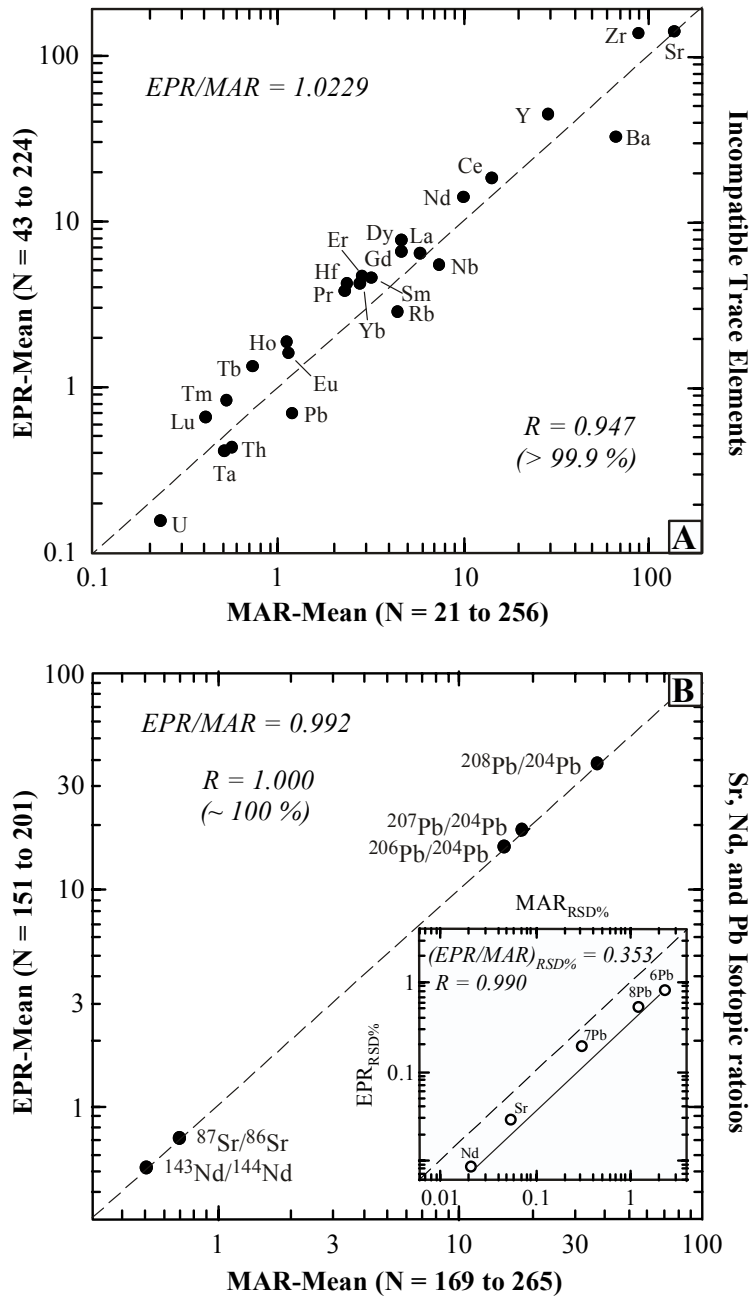


Figure 3. Comparison of mean abundances of incompatible elements (**A**) and Sr-Nd-Pb isotopic ratios (**B**) between MORB from the fast-spreading EPR (23°S to 23°N) and the slow-spreading MAR (55°S to 52°N) using the recently available global MORB data base [33]. Note the statistically significant correlation with a nearly unity (1.0229) slope in **A**, suggesting similar “plume material” contributions to the two ocean ridge systems. In **B**, the mean Sr-Nd-Pb isotopic ratios are statistically identical, which reinforces that “plume material” contributions are identical at the EPR and MAR. The correlated smaller variability ($RSD\% = 1\sigma/\text{mean} \times 100$) of EPR MORB isotopic ratios plotted in the inset reflects a well-known effect of greater extents of melt homogenization in EPR MORB. The $RSD\%$ for incompatible element abundances are not correlated, thus not shown, which is largely due to inhomogeneity in data quality (analyzed by different means in different laboratories with variable precisions and accuracy) in the literature, whereas Nd-Sr-Pb isotopes are all determined by TIMS normalized to international standards in all laboratories. Note that logarithmic scales are used to show all the details.

What's driving what?

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Abstract:

What's driving what: An active asthenosphere driving a passive lithosphere, or the other way around? Important clues are obtained from looking at the origin and movement of oceanic plateaus and their transformation into accreted terranes, their impact on arc volcanism, the opening of back arc basins, etc. By and large these processes tend to be more consistent with the notion that plate mechanics and the mechanics of plate interactions control what we call plate tectonics, not asthenospheric processes.

This is supported by (a) At convergent plate boundaries - the way oceanic plateaus influence processes when they dock at plate boundaries (e.g., volcanism, mountain buildings); (b) At divergent plate boundaries - the general migration of ridges relative to the asthenosphere suggest that associated so called "hot spots" must also be moving relative to the asthenosphere; (c) Back arc basins - the extensional episodic opening of back arc basins within compressional/convergent settings suggest that it is the strength and cohesiveness of the plate that control this divergent process. The asthenosphere is responding passively; and (d) At transform boundaries - big bends, and bending of major plate boundary faults systems (e.g., San Andreas at the transverse ranges, Dead Sea transform at the Lebanon Mountains) suggest that it is their weakness that controls the behavior of the boundary, not the underlying asthenosphere.

Distinguishing local from deep sources using high-resolution age-mapping of oceanic-hotspot volcanism?

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Introduction

The temporal, spatial and geochemical distribution of hotspot volcanism has long been a key to investigating the processes controlling hotspot-lithosphere interaction and the hypothesis of deep mantle plumes. High-precision dating is therefore of first-order importance when seeking to understand the long-term processes controlling the history and distribution of hotspot volcanism. Our aim here is to draw attention to the issue of plume theory developing much faster than the accumulation of real data due to the prohibitive cost of ship time and post-cruise analyses. Recent studies indicate that data remain far too scarce to provide robust ocean-wide understanding of histories and distributions of hotspot provinces. We support this proposition with the example of recent ⁴⁰Ar/³⁹Ar data for rocks from the Foundation Seamount Chain (O'Connor et al. 1998, 2001, 2002) (Figs. 1 & 2). Our results show how a detailed understanding of the long-term history of time progressive volcanism along seamount chains and their surrounding structures can begin the process of distinguishing long term (i.e., deep?) plume-hotspot behavior from local lithospheric control.

Short-lived (local) versus long-lived (deep?) control of Foundation hotspot volcanism

The main trend in the Foundation age data is one of linear migration of midplate – often geochemically enriched – volcanism at a rate of 91 ± 2 mm/yr along the Foundation Chain for at least the past 22 Myr. Such time progressive volcanism supports the conventional model of the Pacific plate drifting over a narrow, stationary plume of hot mantle material upwelling from depth. Furthermore, similarity between rates of migration of volcanism along the Hawaiian and Foundation chains supports a stationary Foundation versus Hawaiian mantle plume, at least for the past 22 Myr.

However, our dredge-sampling covered volcanic elongated ridges (VER) flanking the Foundation Chain at different stages of development (Fig. 2). The transition from a narrow line of seamounts to a broad region of volcanic elongate ridges (VERs) about 5 Myr ago was assumed initially to be the result of interaction between the Foundation plume and the encroaching Pacific-Antarctic spreading-center. Some of our data support this notion by showing that volcanism along morphologically distinct VERs can develop occasionally as rapidly formed continuous lines of coeval volcanism extending from a region of intraplate volcanism to the Pacific-Antarctic spreading center. However, a significantly more dominant trend is for coeval, yet structurally disconnected, segments of Foundation Chain VERs to develop in a series of *en echelon*, NE-SW elongate ‘zones’ of coeval hotspot volcanism. These elongate zones developed at intervals of approximately 1 Myr while maintaining a basically steady-state orientation and size as the Pacific-Antarctic spreading center migrated continually closer to the Foundation plume hotspot. Although such VER development was controlled in part by local factors (e.g. location of nearest spreading center segment, lithospheric stress), long-lived attributes of the Foundation plume hotspot (e.g. size, orientation, periodicity) appear to have played a significant role.

The key to testing this notion is the fact that the Foundation Chain represents a rare, possibly unique, case of a hotspot trail crossing a fossil microplate. Prior to encountering the Selkirk Microplate the Foundation Chain formed as broad zones of scattered, synchronous Foundation volcanism – similar to those identified west of the present Pacific-Antarctic spreading center (Fig. 2). However, once the significantly older microplate lithosphere began capping the plume hotspot about 14 Myr ago, the chain narrowed abruptly into a line of discrete seamounts, only broadening again about 5 Myr ago when sufficiently young lithosphere once again drifted over the plume hotspot. Foundation hotspot volcanism can therefore be prevented across elongate hotspot zones if the capping tectonic plate is too thick for plume melts to penetrate to the surface. (O'Connor et al., 1998, 2001, 2002). The lack of a seamount chain connecting the Foundation and the Ngatamato chains (McNutt et al., 1997) can be similarly explained, so supporting the notion that the Pacific plate has drifted a distance of at least 3400 km over a Foundation plume-hotspot during the last ~34 Myr. We infer from this information that Foundation Chain development was controlled primarily by tectonic plate migration over broad zones of hot plume material of fundamentally constant size and orientation created with an apparent periodicity of about once per Myr (O'Connor et al., 2002).

Creation of broad zones of synchronous Foundation magmatism at regular ~1 Myr intervals leads us – in combination with recent numerical plume modeling (e.g., Larsen and Yuen, 1997; Larsen et al., 1999) – to propose that the Foundation Chain is the product of a stationary plume pulsing hot masses against the base of the Pacific plate from depth with an apparent periodicity of once per Myr (O'Connor et al., 2002). Assuming the validity of the hypothesis of deep mantle plumes (Morgan, 1971), our model for Foundation Chain development has implications for future investigations of Pacific midplate volcanism. We propose that plume-hotspots such as Foundation, spreading on impact with the lithosphere, influence very wide areas such that apparently unconnected hotspot volcanism can be produced simultaneously across wide swaths, often crosscutting seamount chains. Thus, variations in the age, structure and stress patterns of tectonic plates drifting over (pulsing?) mantle plumes might control if, where and how hotspot volcanism develops on the Pacific plate. This modified plume-hotspot theory might also explain widespread scattered midplate volcanism (e.g., VERs) revealed by satellite altimetry mapping as well as randomly distributed reheating events warming and raising Pacific lithosphere (Smith and Sandwell, 1997) – given that many other mantle plumes are similarly pulsing large masses of hot plume material (not necessarily with the same periodicity or mass) into broad regions impacting the base of the Pacific lithosphere.

Conclusion

While we find evidence for a link between local plate tectonic processes (lithospheric architecture, stress, rifting) and the distribution of hotspot volcanism we also see evidence for long-term underlying episodic/periodic 'plume-hotspot' control. Thus, in the case of the Foundation hotspot we believe that we can distinguish between second-order lithospheric and first-order 'plume-hotspot' processes controlling the history and distribution of volcanism. This insight would not have been possible without an unusually extensive dredge-sampling and post-cruise analytical program. For example, the conventional wisdom that the broad region of volcanic elongate ridges near the Pacific-Antarctic spreading axis are primarily the product of plume-ridge interaction would still prevail – especially considering the focus of so many resources on active spreading-ridge research.

Inferring plume behavior from localized studies of oceanic volcanism inevitably produces a 'snap-shot' of what could well be a long-term dynamic mantle process. We believe therefore that the possibility of distinguishing local from long-lived (deep?) processes controlling the history

and distribution of hotspot provinces provides the opportunity of 1) better testing current plume-hotspot theory and 2) merging new multidisciplinary thinking with the acquisition of real data from selected volcanic provinces. In short, developing and testing old – and especially new ideas and models – requires significantly more detailed sampling and age/geochemical analyses.

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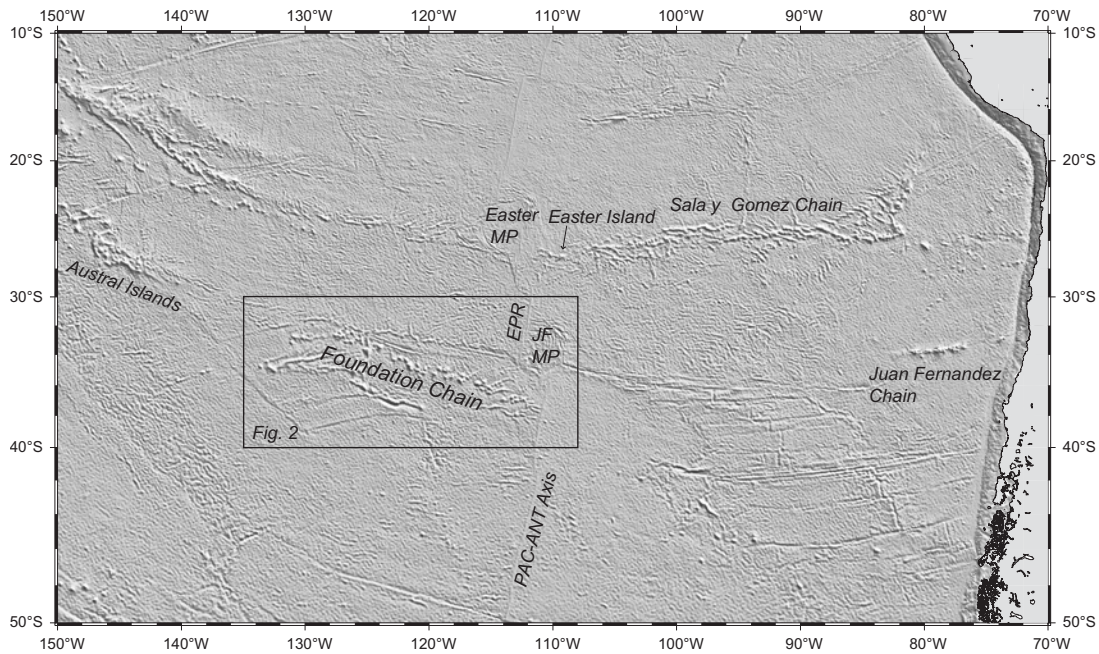


Figure 1. Predicted topography (Smith and Sandwell, 1997) of SE Pacific seafloor showing the location of the Foundation Chain. MP = microplate; JF = Juan Fernandez; EPR = East Pacific Rise. Figure modified after O'Connor et al., 1998

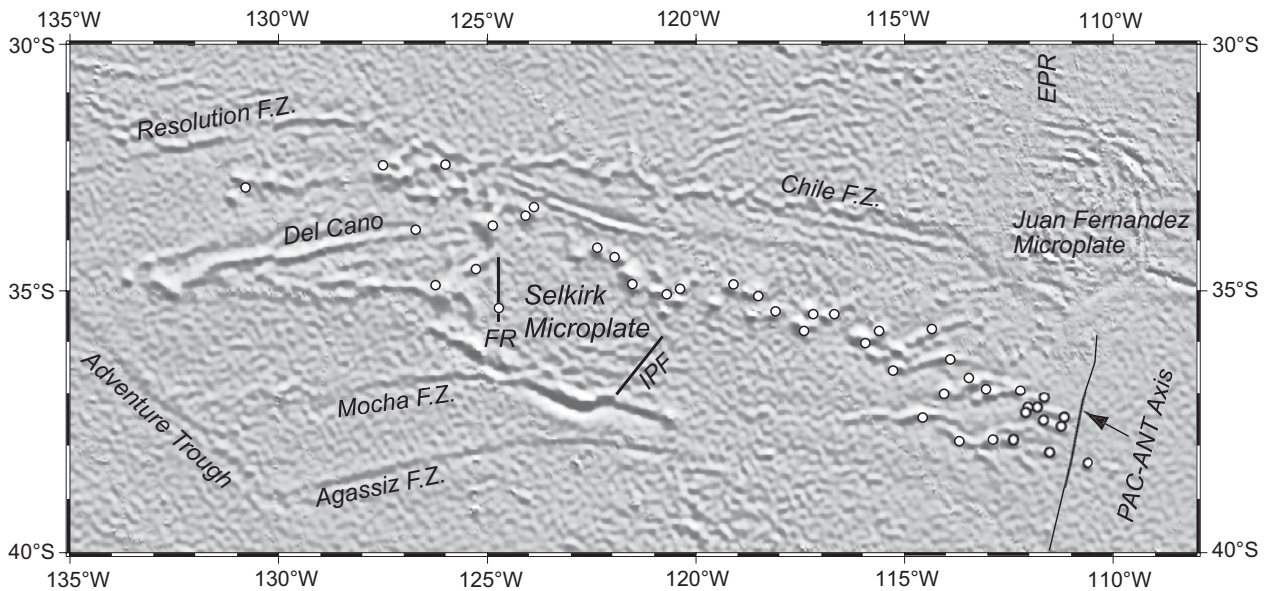


Figure 2. Predicted topography of the Foundation Chain (Smith and Sandwell, 1997). *F. S. Sonne* and *N/O Atalante* dredge sites are indicated by black rimmed white dots. $^{40}\text{Ar}/^{39}\text{Ar}$ ages, details of sample information and analytical date are in (O'Connor et al., 1998, 2001, 2002). IPF = inner pseudo fault and FR = failed rift of Selkirk microplate (Mammerickx, 1992). Figure modified after O'Connor et al., 1998

MANTLE PLUMES: FERTILE? FECUND? PHANTASMAGORICAL? OR SIMPLY FANTASTIC? PIMMS ANYONE?

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Places of Irruption of Much Magma (PIMMs) like Iceland, Hawaii and many other postulated surface expressions of hypothetical mantle 'plumes' may be much less 'productive' than MORs - the PIMM source is not *required* to be volumetrically more 'fertile' than MORB mantle (not true of LIPs, however). Eruptives at PIMMs typically have lower eruption temperatures than MORB, hence it may be misleading to think of PIMMs as either 'hot spots' or 'heat spots'.

At any fixed temperature within the earth's upper mantle, depleted (residual) peridotite is always less dense (higher Mg#, less spinel or garnet) than fertile peridotite and much less dense than eclogite. If any *solids* are going to ascend adiabatically *within* the upper mantle, they should be depleted peridotite rather than fertile peridotite, and certainly not eclogite. 'Solutions' invoking higher temperature in the fertile peridotite or eclogite are prone to encounter the solidus before buoyancy sets in; those invoking patches or veins of fertile peridotite, pyroxenite or eclogite entrained within ascending residual peridotite require large volumes of ascending and spreading peridotite, for which there is little direct evidence; solutions invoking plumes from the lower mantle 'pushed' through the transition zone encounter negative buoyancy because of the effect of lower Mg# on the phase changes, and pay an enthalpy and temperature toll as olivine is formed at the transition zone/upper mantle boundary.

The model of a 'plume' of hot, fertile peridotite floating through depleted asthenosphere and possibly lithosphere also, and *partially melting because of its adiabatic decompression*, leaves unanswered at least four questions:- Why do trace element features apparently require small mass fractions of partial melting, when the local abundance of irrupted products apparently requires high mass fractions of partial melting or alternatively an exceptionally large source region? Why is there so little evidence of hot, low density residual peridotite spreading away from the surface expression of a 'plume'? How is the need for buoyancy in the model to be reconciled with the density of fertile peridotite? Where does the enthalpy of melting of basalts come from and how is it concentrated, given that conduction of heat through the solid will be slow and would proceed down-temperature, out of the allegedly hot plume?

Partial melting in the upper mantle produces solid residue and partial melt liquid, *both* less dense than the fertile peridotite they replace. If the enthalpy source is external, both products are also as hot as or hotter than the unmelted fertile peridotite, further enhancing buoyancy. An alternative model envisages ascent and adiabatic decompression of mantle materials commencing *because they have already partially melted*. This could result from a heat flow out of the lower mantle and an intersection between the geotherm and a cusp on the mantle solidus at the transition zone-upper mantle boundary. The liquids might escape by cracks and conduits (with much modification en route) to form

PIMM volcanoes while the solids of very different rheology might rise (with further partial melting) somewhere else than at the site of the PIMM.

Small mass fractions of nearly isobaric partial melting of a vast and diversified source region are permitted; that source region automatically replenishes itself by gravity controlled subsidence of upper mantle into the partial melting zone, facilitating fertile mantle and eclogite participation; the relative density problems of fertile 'plumes' vanish; enthalpy is concentrated by flow of partial melt, not by flow of heat; location of the PIMM is controlled by tectonics and is not required to be fixed; and the peridotite most directly associated with the PIMM would be hot ultramafic cumulates formed adjacent to the conduits from the ascending melts - a mass of the same order of magnitude as, rather than ~ 10 times greater than, that of the erupted magma at a PIMM.

The petrogenetic environment underpinning PIMMs would then be one of progressive partial crystallisation throughout the upper mantle, of an originally ultramafic partial melt, capped by substantial low pressure partial crystallisation in a thick volcanic superstructure. This is *radically* different from the regime underpinning MORs, where picritic or olivine basaltic parental liquids form at high levels by adiabatic decompression melting continuing and advancing until close to the Moho, capped again by substantial low pressure crystallisation in a thinner volcanic superstructure.

**GEOCHEMICAL AND ISOTOPIC VARIABILITY OF PLIO-
QUATERNARY MAGMATISM IN ITALY : PLUME VS. SHALLOW
MANTLE PROCESSES**

BY

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ABSTRACT

Plio-Quaternary magmatism in Italy exhibits an extremely variable composition, which spans almost entirely the spectrum of magmatic rocks occurring world-wide (Fig. 1). Petrological and geochemical data provide a basis for distinguishing various magmatic provinces, which show different major element and/or trace element and/or isotopic compositions. The Tuscany Province (14-0.2 Ma) consists of magmas generated through crustal anatexis and of mantle-derived calcalkaline to ultrapotassic rocks; mantle-derived rocks in this province contain mantle xenoliths, but display very radiogenic Sr isotope signatures and LILE/HFSE ratios close to upper crust. The Roman and Neapolitan provinces (0.8 Ma to present) consist of dominant potassic to ultrapotassic, which still possess crustal-like geochemical and isotopic signature. The Aeolian Arc (1 Ma to present) consists of calcalkaline to shoshonitic rocks. The Sicily Province contains young to active centers (notably Etna) with a tholeiitic to Na-alkaline affinity. Finally, volcanoes of variable composition occur in Sardinia and, as seamounts, on the Tyrrhenian Sea floor. Magmas in the Aeolian arc and along the Italian peninsula have a subduction-related geochemical character, whereas the Sicily and Sardinia Provinces display intraplate signatures. Intraplate and orogenic volcanics coexist on the Tyrrhenian Sea floor.

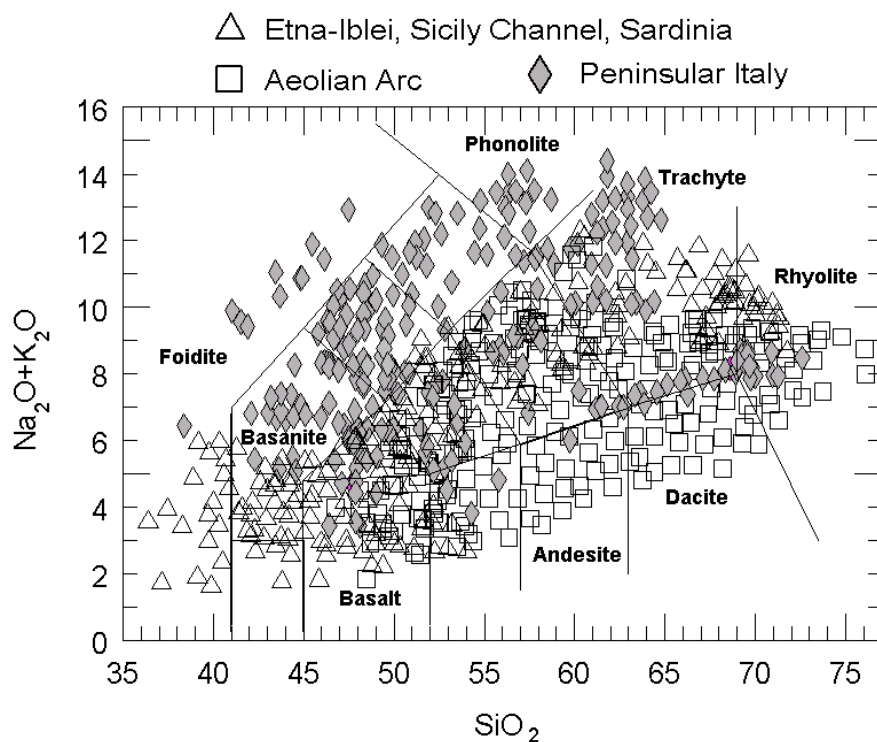


Fig. 1. Total alkali vs. silica diagram for Italian Plio-Quaternary magmatism. Note the extreme compositional variability that covers the entire spectrum of igneous rock compositions occurring worldwide.

Sr, Nd and Pb isotope ratios of mafic magmas ($\text{MgO} > 4 \text{ wt\%}$) from the various provinces are variable, and compositions akin to MORB, EMI, EMII, HIMU and FOZO are found. These geochemical and isotopic complexities reveal that the upper mantle beneath Italy consists of compositionally distinct domains, covering both orogenic and anorogenic characteristics. Explaining these diversities has profound petrological and geodynamic implications.

Sr vs. Nd isotope variations display a curved trend between MORB and upper crust, highlighting interaction between mantle and upper crustal reservoirs. On the other hand, Pb vs. Nd and Sr isotopic variations display more complex patterns, which reveal mixing between HIMU and EMI and between HIMU or FOZO and upper crust (Fig. 2). These complexities have been suggested to be related to a zoned mantle plume beneath the Tyrrhenian sea. A role for both plume and upper crustal materials brought into the mantle by subduction processes has been also suggested (Gasperini et al., 2002; Bell et al., 2003). The alleged occurrence of "carbonatites" in the Italian peninsula has been considered to strongly support the plume hypothesis.

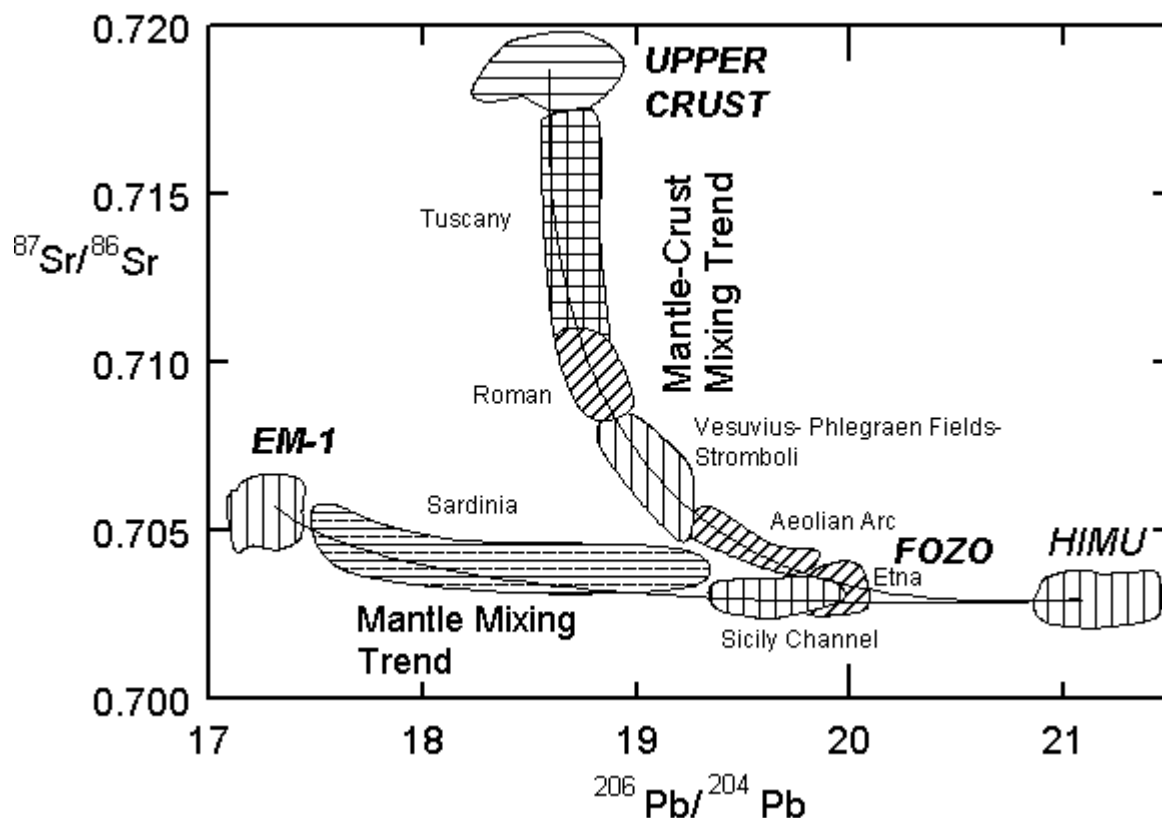


Fig. 2. $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagram for Plio-Quaternary mafic Italian volcanics. The isotopic variations can be interpreted as resulting from mixing between mantle and upper crust and between HIMU and EM1 mantle reservoirs.

However, the mafic rocks with high $^{87}\text{Sr}/^{86}\text{Sr}$, low $^{143}\text{Nd}/^{144}\text{Nd}$ have incompatible trace element patterns that resemble very closely the upper crust. This implies a genesis in an upper mantle contaminated by crustal material. However, the close geochemical similarity with upper crust suggests recent contamination and argues against a plume hypothesis. Moreover, rocks with FOZO-HIMU isotopic compositions are very widespread in the Mediterranean region suggesting that this represents a resident rather than allochthonous (i.e. plume) mantle material. Finally, scrutiny of geochemical data clearly indicates that the so-called "carbonatites" actually represent silicate potassic rocks affected by secondary replacement by carbonates from sedimentary wall rocks.

The complex petrological and geochemical variations of magmatism in central-southern Italy are best explained by assuming that various metasomatic events affected the upper mantle during the Apennine orogenesis (Peccerillo, 1999). The pre-metasomatic mantle rocks had a heterogeneous composition, as it is expected for a zone affected by several ancient orogenic cycles (e.g. Hercynian, Alpine etc.). The superimposition of multiple metasomatic events over a compositionally variable pre-metasomatic mantle is able to generate an extremely heterogeneous upper mantle, whose geochemical and isotopic variability is inherited by the erupted magmas.

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Is hot spot magmatism, like Hawaii, coming from shallow mantle?

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Earth's magmatism occurs primarily at hot spots (Hawaii), subduction zones (convergent margins like Japan) and mid-ocean ridges (divergent margins like the Mid-Atlantic Ridge). In all these cases the Earth's mantle partially melts to generate magma. The isotope geochemistry of the magmas, and hence the source of the magma, erupting at mid-ocean ridges (MORs) is distinct from those of hot spots. Currently two different models explain the origin of these magmas. This study proposes a test (of the Anderson model) by comparing the geochemistry of the magmas with a plate motion parameter that should correlate with the depth of the source mantle.

We have looked for a possible relationship between the absolute migration speed of MORs and the radiogenic isotope geochemistry of MOR basalt (MORB). A positive correlation between ridge migration velocity and degree of isotopic enrichment in MORB may bolster the idea of a global, shallow, buoyant, enriched mantle reservoir, because more rapidly migrating ridges should tap shallower mantle. Preliminary results show no strong correlation between present ridge migration rates and the radiogenic isotope composition of MORB. However, the global MORB database is ever expanding and more data points are becoming available. Also, there are other reference frames for calculating migration rates that might be explored. It is our intention to retrieve more data for analysis and use other reference frames to determine whether a global, shallow, buoyant, enriched mantle reservoir exists.

Petrological Constraints on Potential Temperature

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The existence of an active volcanic center or "hot spot" has commonly been taken as evidence of a hot diapir or "plume" ascending from the core-mantle boundary. Because many of these centers are located near or on volcanically active ridges, the origin of both types of volcanism must be considered together.

Determination of the potential temperature (T_p) is fundamental to correct modeling of the magma generation process, and would also address the issue of whether or not enhanced volcanic activity at a ridge (Iceland) or remote from a ridge (Hawaii) is driven by a hot diapir originating at the core-mantle boundary, with consequent implications for mantle dynamics. Large disagreements about T_p at Hawaii and along spreading ridges have persisted for decades. In Hawaii, olivine-controlled fractionation is well-documented and various authors have suggested that the parental magmas for this fractionation lie either toward the olivine-rich (high T_p) or olivine-poor (low T_p) end of this trend. Picritic glass shards on the ocean floor from the Puna Ridge of Kilauea (Clague *et al.*, 1995) are compositionally between the two extremes and require a minimum T_p of about 1420°C (Gudfinnsson and Presnall, 2003). No comparable picritic glass compositions have been found at Iceland or anywhere else along a spreading ridge. See Gudfinnsson *et al.* (this conference) for a discussion of the origin of the most magnesian volcanic glasses from Hawaii and Iceland.

The T_p along spreading ridges, which includes Iceland and close proximity to a number of hot spots, has been controversial since the beginning of experimental studies of MORBs. From 1967 to 1988, petrologists were about evenly divided between those favoring low-pressure (0.7-1.1 GPa, low T_p) and those favoring high pressure (1.5-3.0 GPa, high T_p) magma-generation models. Beginning with the papers of Klein and Langmuir (1987, 1989) and McKenzie and Bickle (1988) and extending to the present time, the earlier controversy has been forgotten and fractional polybaric melting over a wide range of T_p (about 1240-1510°C in the model of Klein and Langmuir) has been widely accepted. However, Kelemen *et al.* (1997) found that the trace element data for abyssal peridotites are consistent not only with near-fractional melting, but also with a range of batch plus fractional melting scenarios. Both Asimow *et al.* (1999) and Walter (in press) concluded that a model involving a combination of fractional and batch melting best fits the data. In addition, Walter (in press) concluded that the pressure range could be small and centered at about 1 GPa (T_p of about 1240°C). Green *et al.* (2000) have argued that both hot spot and MORB magmas are generated at a T_p of about 1430°C. Presnall *et al.* (2002) have recently presented a model that includes melting over a wide range of pressures from 0.9 to 7 GPa, but emphasizes the dominant role of low pressures (0.9-1.5 GPa, T_p of 1240-1260°C) and mantle heterogeneity in producing the major-element systematics of MORBs. The evolution over the last 36 years from a low vs high T_p controversy to widespread acceptance of a large range of T_p , and now to revival of the original debate over high vs low T_p is remarkable. An

extension of this debate will be presented that integrates experimental phase equilibrium data with global seismic tomography and provides strong support for a low and relatively uniform T_p in most parts of the ocean basins. This requires mantle heterogeneity and dominance of athermal processes in the formation of ridges and "hot spots" near or on ridges, and is consistent with the plate tectonic model of Anderson (2002).

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Flood basalts (LIP) of continental margin west of India: inferences of long duration history

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Deccan flood basalts of west of India and their offshore extensions would cover $\approx 3 \times 10^6 \text{ Km}^2$, overlie Precambrian granites and/or Mesozoic sediments and considered as one of the Large Igneous Provinces (LIPs) of the world. The basalts are $\approx 2 \text{ km}$ thick (maximum) and $\approx 10^6 \text{ Km}^3$ in volume. They are prolific in important peri-cratonic rift basins; the Kutch Basin, Kathiawar Basin and Bombay Offshore Basin of the western continental margins of India. The Reunion hotspot outpoured them during the Plume head interaction with northward moving Indian plate (Duncan, 1993) and in short, 1 to 3 million years duration $\approx 65 \text{ Ma}$ (Courtillot et al., 1988 and Duncan and Pyle, 1988). Trail of the Reunion plume manifested as linear ridges covered by sediments along the shelf margin basin (Krishna et al., 1994., Chaubey et al., 2002 and Gopala Rao et al., 2003). It has been traced further south through the Chagos and Saya-De-Malha Banks and Mascarene Plateau up to the Reunion Island. However, the plume role in continental rifting and vice versa, mode and duration of emplacement of volcanism are important aspects of evolutionary history of the LIPs.

Recent Industry Drill wells and marine geophysical investigations have revealed that basalts occur on the entire shelf (Kothari et al., 2001., Rathore et al., 1997 and several others). The seismically imaged trail part of Mangalore coast $\approx 12^\circ\text{N}$ consists of four closely spaced parallel sub-surface ridges each of 12 to 15 km wide at basement level and $> 3 \text{ km}$ high unlike in other areas. They are associated with positive free-air gravity, $> 50 \text{ mGal}$ and magnetic anomalies, interpreted as volcanic constructs and mark an phase of intense volcanism. Considering the northward relative motion of the Indian Plate, the plume would reach $\approx 12^\circ\text{N}$ in about 4 to 5 Ma later to the main phase of Deccan volcanism. Yet another later intense volcanism has produced the voluminous Chagos volcanic Bank in further south. The radiometric, Kr-Ar and Ar/Ar ages of basalts of Ratnagiri coast at 18°N (drill well R8-1) (Gopala Rao, 1996 unpublished report), Padua Bank at 12°N (Kothari et al., 2001), Cochin coast at 10°N (K-K-1 drill well at 8°N) (Sastri, 1981) and Chagos Bank at 4°N (ODP site 713) (Backman and Duncan et al., 1987) are ≈ 63 , ≈ 60 , 55-58 and 48 Ma respectively. Further younger, 42 Ma age basalt are also reported in Gulf of Cambay (drill well MT3) (Gopala Rao, 1996, unpublished report).

The intrinsic observations of the onshore basalts studies are 1) Deccan volcanism of northwest India $\approx 65 \text{ Ma}$. 2) the acidic and intermediate differentiates (trachytes and rhyolites) and basic (diorites and monzodiorites) magma plugs, 60 to 61.5 Ma age of the Deccan volcanic province of the west coast and Saurashtra area. Their isotopic, Sr, Nd and Pb compositions are similar to Deccan trap volcanics. The intrusions occurred from the under-plated gabbroic layer as a result of decompression partial melting and shallowing into the crust 4 to 5 Ma later to Deccan volcanism (Sethna, 2003). Younger volcanism has been reported in the shield area of the northwest India (Sahasrabudhe, 1963; Hooper, 1990). and 3) Kennet and Widiyantoro (1999) from seismic tomography studies have inferred a low velocity region north of the Cambay Rift of Saurashtra at sub-

crustal depth that had developed at least 3 million years preceding the Deccan volcanism. The sequence of events of Saurashtra, northwest India possibly indicates a magma source that could incubate at sub-crustal depth and give rise to repeated volcanism subsequently. Their recurrence perhaps is related to the tectonics of the region. The geochemical and isotopic similarities of the later volcanism and the Deccan volcanism of northwest (Saurashtra) India imply a magma source at sub-crustal depth feeding the later volcanism at Saurashtra.

The Laxmi and Laccadive Ridges are interpreted as continental slivers rifted from the Indian mainland (Naini, 1981; Chaubey et al., 2002 and Gopala Rao et al., 2003). Series of Seaward Dipping Reflectors (SDRs) and their feather edges occur along the western flank of the ridges in the north and south. Thus, the ridges in ~3.2 km water depth separate the plume trail of the shelf margin basin in the east from the SDRs and oceanic crust of the eastern Arabian Sea in the west. The plume appears to have remained for long time in intra-continental setting. Thus their mode of emplacement of volcanism is quite different in both cases.

The SDRs along the western flanks of the ridges interpreted as rift related volcanism that are developed during the initial continental break up where as the volcanic constructs of the shelf margin basin are plume generated. In such case we need to consider two different sources for the volcanism. Then the question is do the SDRs prevail all the rifted continents? Even under unfavorable conditions? Even though the SRDs have been identified in the north and south of the passive volcanic margin west of India where intense phases of plume volcanism had occurred, such SDRs are not found on the margin of central west coast of India. It does imply their occurrence depends on supply of magma and kinematics.

Chaubey et al, (1998) had reported two phases; through 68 Ma to 52 Ma (magnetic isochrons 28 to 24) and 30 Ma to Present (magnetic iso-chron 11 to present) seafloor spreading punctuated by a pause in spreading in the Arabian Sea. Perhaps plate tectonics and seafloor spreading could be one of the potential sources for initiating the tectonism and/or magmatism along the margin. The aspects have relevance to the long duration history of evolution of the Large Igneous Province and plume related volcanism.

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Tharsis Rise, Mars: Is there room for a plume?

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The Tharsis Rise is an area of extensive volcanism containing the largest shield volcanoes in our solar system. A number of investigators have suggested that the sustained volcanism, geoid and topographic anomalies that comprise the Tharsis Rise to be the result of a mantle plume. Harder and Christensen (1996) presented a calculation for convection in a Mars-sized body that resulted in a single plume. However, their calculation evolved through stages of several plumes down to a single plume and took greater than the age of the solar system to develop into a single plume.

Efforts to remove the isostatic contribution to the geoid and isolate the dynamic contribution have shown that while much of the long wavelength signal can be explained by the crust, there is a significant mantle component (Kiefer et al., 1996; Whitesell and King, 2001). Furthermore, dynamic models suggesting Tharsis is largely supported by convection (Kiefer et al., 1996; Harder and Christensen, 1996; Harder, 2000; Kiefer, 2001) can justify the young ages of the Tharsis shield volcanoes. Thus, there is reason to believe that a mantle plume or plumes may exist beneath Tharsis.

Research conducted thus far has consisted of varying the Rayleigh number and rate of internal heating in an isoviscous rheology and activation energy in a temperature-dependent rheology. The geoid and topography over isoviscous plumes in a Mars-sized body are greatly reduced with increasing Rayleigh number and internal heating. Additionally, calculations over

temperature-dependent plumes show that after a thick, strong lithosphere forms, the geoid and topography from a plume become even smaller. While our exploration of geoid and topography over plumes suggests that there could be a plume under Tharsis, the plume hypothesis has a number of shortcomings: Why is there only one (possibly two) plume(s) on Mars?; Why has the plume remained stable over a long period of Martian history?; What heat source in the core could have produced enough heat to maintain the plume till present? Is it a coincidence that the plume is so closely spatially related to the crustal dichotomy? Because of these questions which the plume model does not adequately address, we are exploring the hypothesis that the Tharsis volcanism is due to small scale convection within the upper mantle that may behave much like edge-driven convection due to the Martian crustal dichotomy – the almost equatorial division between the heavily cratered, thick crust of the southern hemisphere and the thinner crust of the low-lying volcanic plains in the northern hemisphere.

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Tectonic Evolution of Shatsky Rise: A Plateau Formed by a Plume Head or Not?

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Shatsky Rise is an oceanic plateau located approximately 1600 km east of Japan, in the northwest Pacific Ocean. It is a basaltic mountain range with an area nearly equal to Japan or California, qualifying as one of the globe's larger LIPs. What makes Shatsky Rise unusual among Pacific plateaus is that it formed prior to the Cretaceous Normal Superchron, so that its relationship to spreading ridges can be observed by mapping adjacent magnetic lineations (Fig. 1). Those lineations indicate that Shatsky Rise formed at spreading ridge, similar to Cenozoic-age Iceland. Although much of the existing geologic evidence can be interpreted as supporting the plume-head hypothesis, some data leave room for doubt.

Shatsky Rise consists mainly of three bathymetric highs and a low ridge that sit at the confluence of the Japanese and Hawaiian magnetic lineations from M21 to M10 (Fig. 1; Sager et al., 1988; Nakanishi et al., 1989; Nakanishi et al., 1999). Magnetic lineations can be traced between the highs, implying the basaltic mountains are three large, individual volcanoes separated by lithosphere that has not been greatly altered by volcanism (Sager et al., 1999). Seismic profiles adjacent to the large edifices suggest that much of the low plateau is a sedimentary apron (Sager et al., 1999). The volume of volcanic material appears to decrease northwestward along with volcanic edifice ages. The southern high is largest, with a volume of $2.4 \times 10^6 \text{ km}^3$, whereas the central and northern volcanoes have volumes that are only 0.69 and $0.65 \times 10^6 \text{ km}^3$ (Sager et al., 1999). The ridge, extending from the north end of the rise, is smaller still. All of these larger volcanic edifices display complete isostatic compensation, suggesting they formed at about the same time as the lithosphere. This has been demonstrated for the southern volcanic high, as basement basalts cored on ODP Leg 198 have been dated at 144 Ma (R. A. Duncan, personal communication, 2002), in agreement with the accepted age of the magnetic lineations that surround the edifice (M21-M19, 148-145 Ma in the Gradstein et al., 1994 timescale). Because this date is older than the lithosphere beneath the other edifices farther north, the rise must become younger in that direction. Indirect evidence implies that the largest and oldest volcanic edifice, the southern high, formed very rapidly. The magnetic anomaly of this edifice is consistent with a predominantly reversed magnetic polarity that suggests formation mostly during a single magnetic polarity chron (Sager and Han, 1993). Assuming that the southern high erupted during the longest reversed chron near M19 (Chron M17) gives emplacement rates similar to estimates of flood-basalt eruptions ($1.7 \text{ km}^3/\text{yr}$; Sager and Han, 1993).

The tectonic evolution of Shatsky Rise, interpreted from the magnetic lineations and bathymetry, can be construed as the result of a plume that "captured" a triple junction (Nakanishi et al., 1999; Sager et al., 1999). Prior to M21 time, the Pacific-Farallon-Izanagi ridge moved NW relative to the Pacific plate, probably in a stable ridge-ridge-ridge configuration (Sager et al., 1988). At M21 time, the triple junction jumped 800 km eastward to the location of the southern volcanic high. Until about M10 time, approximately 16 Myr later, the triple junction moved NE with a speed and direction that cannot be easily reconciled with spreading rates and ridge geometry (Sager et al., 1988). It appears there were many small ridge jumps that affected the triple junction, with the largest corresponding to the emplacement of the large volcanic highs

(Nakanishi et al., 1999). Apparently successive eruptions from the Shatsky plume caused the triple junction to jump to the plume location, presumably because of the concentrated heat and upwelling. All of these observations seem consistent with a plume-head explanation: the massive initial eruption “captured” a nearby triple junction and kept it pinned at the plume location until the plume strength waned with the transition from plume head to tail.

Despite circumstantial evidence that supports the plume head model for plateau development, some pieces of evidence do not fit as well. Geochemical evidence for a magma source with a deep mantle origin is equivocal. Sr-Nd-Pb isotope ratios from Shatsky Rise basalts are varied, but tend to be more MORB-like than other Pacific plateaus (J. Mahoney, personal communication, 2002). Nb/Zr and Nb/Y ratios give the same result, although data from one dredge are similar to results from the South Pacific Superswell (Tatsumi et al., 1998). However, those dredge samples come not from one of the main volcanic edifices in Shatsky Rise, but a small, undated seamount that is located between the larger volcanoes and whose relationship to the plateau is unclear. Moreover, even if this seamount was formed as a part of Shatsky Rise and has a South Pacific Superswell source, the connections of the Superswell to the deep mantle is currently debated (e.g., Courtillot et al., 2003). Perhaps the most curious observation that does not fit the plume head model well is the remarkable spatial coincidence of Shatsky Rise and other plateaus with spreading ridges at triple junctions. Not only did Shatsky Rise form at a triple junction, but so did the Magellan and Manihiki plateaus. Moreover, other plateaus, whose relationships to ridges are less clear because they formed during the Cretaceous Normal Superchron (Hess Rise and Ontong Java Plateau), are in locations that suggest that they too were formed at triple junctions or spreading ridges. Unless there is some mechanism causes ridges and triple junctions to jump to or migrate rapidly toward plumes (e.g., Mahoney and Spenser, 1991), these coincidences imply that the ridges somehow cause the upwelling.

As our understanding of the mantle processes that form ocean plateaus continues to evolve, Shatsky Rise is a feature that can be used to test plume models. Because it was formed at a time when the Pacific plate was moving rapidly relative to the mantle, the volcanic signature of the plateau eruptions is spread out laterally, unlike some larger plateaus formed during periods of slow plate motion. Furthermore, the sedimentary cover on Shatsky Rise is thin except at the summits of the main volcanic highs and outcrops on the flanks are common. Thus, sampling Shatsky Rise is less challenging than many plateaus with thicker sediment mantles. Unfortunately, the outcrops tend to be weathered and coated with Mn-oxide crusts, so dredging has been ineffective in collecting the samples needed for accurate dating and geochemical analyses. Drilling on ODP Leg 198 showed that relatively shallow coring into basement can yield remarkably well-preserved Jurassic-age igneous samples.

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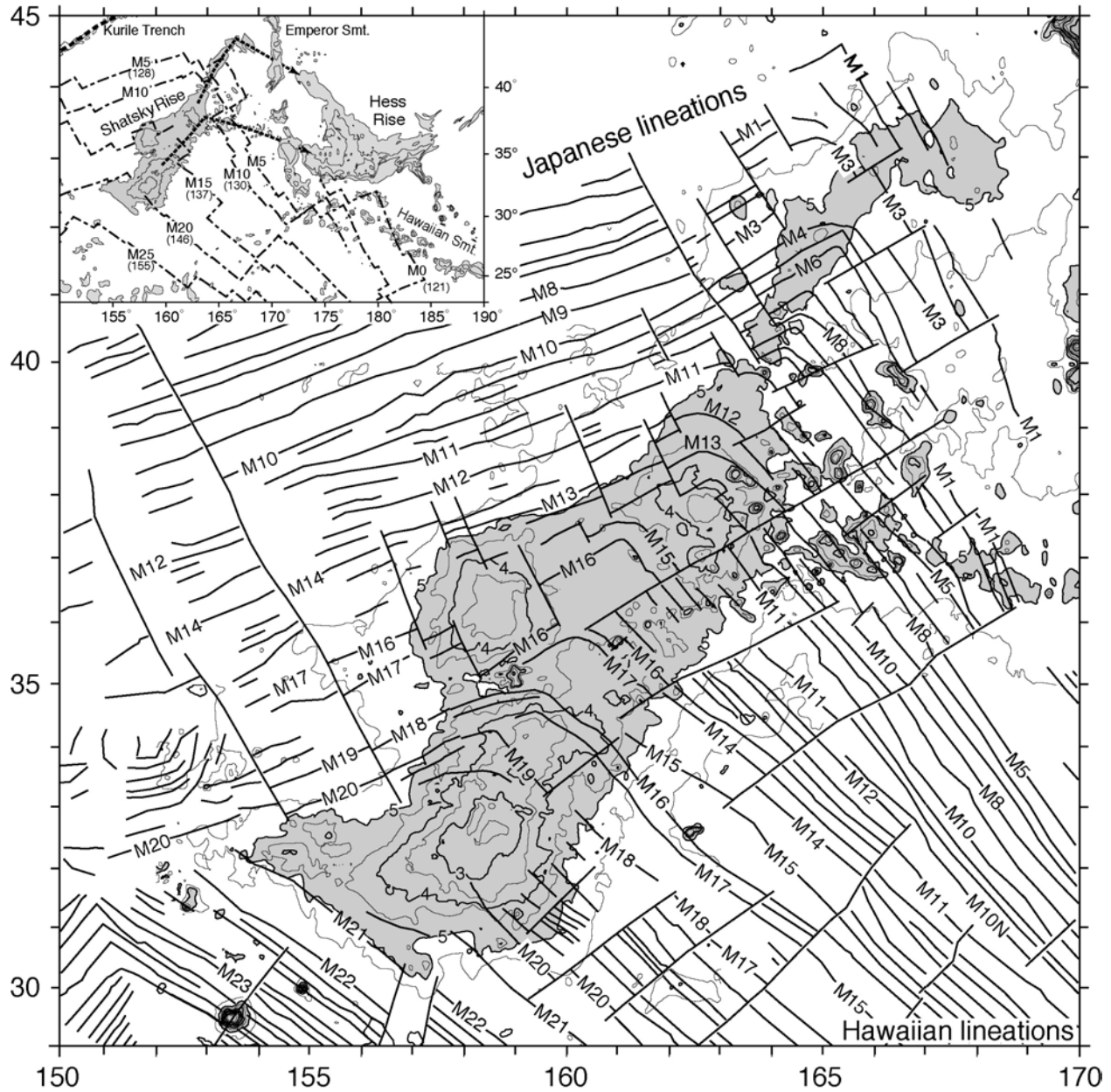


Figure 1. Bathymetry of Shatsky Rise and magnetic lineations. Bathymetry contours are shown at 500-m intervals with heavy contours at 1-km intervals (Sager et al., 1999). Gray area shows the plateau above 5 km depth. Heavy lines denote magnetic lineations and fracture zones (Nakanishi et al., 1999). Inset shows Shatsky Rise region including Hess Rise. Heavy dashed line marks the Kurile Trench, whereas heavy dotted lines represent two seamount trails that connect Shatsky Rise with Hess Rise.

Malani Magmatism of Northwestern Indian Shield: Implications of Mantle Plume?

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The Neoproterozoic magmatism in the Trans-Aravalli region (SW Rajasthan) of the northwestern Indian shield is characterised by anorogenic magmatism. This magmatic event dated around 750 my; post-dating Sirohi Group (Delhi Supergroup) and predating Marwar Supergroup. The rocks occur as hillocks, inselbergs and tors covering approximately 51,000 km² area. This comprises basalt, rhyolite, granite and various porphyry dykes.

Presently, most of the workers are of the view that the real cause of the Malani magmatism is Mantle Plume activity. The Plume resulted wide spread bimodal volcanism, plutonism and dyke intrusions. (Bhushan, 2000; Roy, 2001; Kochar, 2001; and Raval 2000)

After having detailed study of Malani outcrops it is concluded that Malani activity took place along almost parallel fractures in the cratonised thick northwestern shield. The salient characteristics of Malani event can be summarised as follows.

1. The initiation of the event is characterised by the bimodal volcanism with locally occurring conglomerate. At a number of places the basic flows are not exposed and underlie the felsic flows (Pandit and Amar Deep, 1997; Bhushan, 2000). Wherever present, the basic component of volcanism is restricted only to the lower part of the sequence and younger volcanic phases are free from any associated basic volcanics. The Malani magmatism is dominantly acidic in nature.
2. The Malani rocks are free from any type of metamorphism and penetrative deformation as indicated by well-preserved magmatic fabric and absence of any deformational related features.
3. The initiation of the magmatic cycle is marked by an extrusive phase (bimodal at places) followed by emplacement of plutonics (peraluminous and peralkaline granites) and the terminal phase is represented by dyke swarms emplaced through pre-existing volcanic as well as plutonic rocks. The important dyke swarms are present at Sankara, Redana, and Dhanta.

Taking clue from the linear disposition of conglomerates and other sedimentary features in close association with the Malani volcanics, it is obvious that the activity is lineament controlled. Presence of discontinuous conglomerate and associated hydroclastic units suggest development of shallow basins along the tectonic line. These linear fractures initially developed into shallow basins at places, which accorded conglomerate, grit, arkose followed by bimodal volcanics. The initial basic flows in aqueous conditions developed pillow lava structures at Sindreth, Khamal and Bambholai (Sharma, 1996). The bimodal volcanism took place through fissures and central conduits forming cones and calderas. Bhushan (2000) identified cones and fissures in the SW Rajasthan. It is quite likely that some of the magma got emplaced into the shoulders of these incipient basins, coeval with the volcanics.

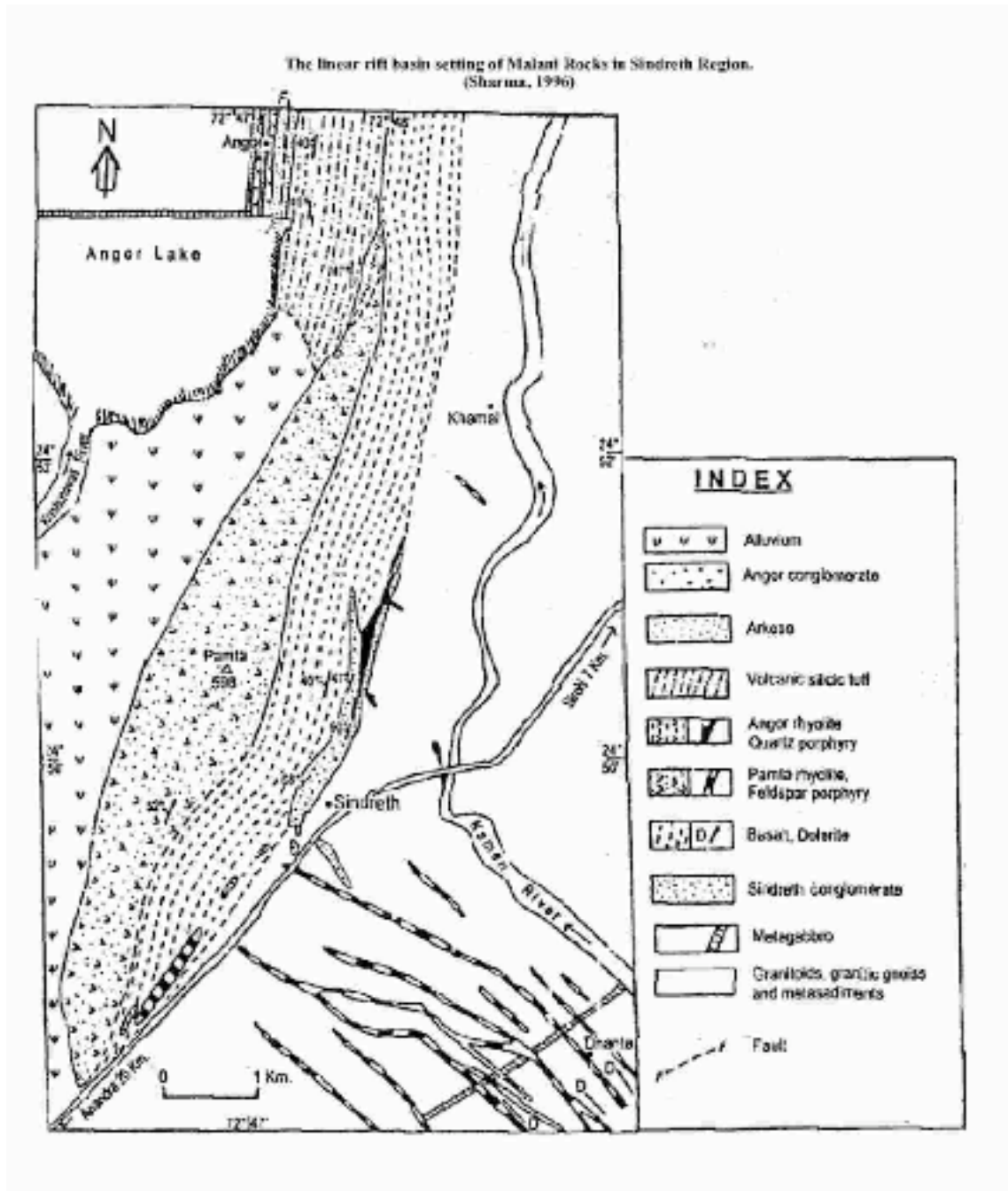
The crustal fractures manifest a rift setting in intra-cratonic and anorogenic tensional tectonic regime. These N-S trending subparallel rifts are separated by basement segment slices, visible at Undwaria, Sindreth, Bamholai, Miniari etc. The narrow-linear outcrop pattern of Undwaria-Sindreth-Miniari is the best example of this. The bimodal volcanism took place in the shallow and narrow basins, which show angular relationship with Sirohi Group and other basement rocks. Similar setting is also observed at Bambholai-Khamal region. These features may not be visible at other places due to sand cover. The plutonic activity is marked by emplacement of granites (Jalore and Siwana Mirpur, Isra and other outcrops) at the margins of the basins. It is quite likely that some of the granites were emplaced (along the margins) simultaneously with the outpouring of felsic lavas. Development of dykes transverse to these rift margins was related to secondary rift fractures (Sharma, 1996; Roy and Sharma, 1999).

The typical basin setting along north-south running tectonic grain and dominantly felsic character of the Malani rocks do not support mantle plume model for their origin. The presence of intermittent basement slices parallel to the Malani rocks indicates anorogenic rift setting. The crust prior to Malani activity was quite stable for long time. This is attributed by pre Malani fine-grained shale-carbonates depositories of the Sirohi cycle. The extensional tectonic regime initiated due to the fragmentation of the Rodinia Supercontinent caused the wide spread Malani felsic magmatism around 750 ma.

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Figure: The linear rift basin setting of Malani Rocks in the Sindreth Region. (Sharma, 1996)



THE DECCAN BEYOND THE PLUME HYPOTHESIS

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Since the rapid rise to dominance of the plume head-tail model for flood basalts (Richards *et al.*, 1989; Campbell and Griffiths, 1990), literally hundreds of papers have invoked, or supported, a plume head origin for the Deccan Traps of India. These papers are unanimous on two counts: (i) the Deccan originated from the Réunion hotspot which upwelled beneath India in the late Cretaceous, and (ii) the hotspot, now located on the African plate, represents a deep mantle plume. However, I consider the mantle plume model invalid for the Deccan.

I relate continental flood basalt (CFB) volcanism to continental rifting, which often (but not always) evolves into full-fledged sea-floor spreading (Sheth, 1999a). I relate the rifting itself to plate stresses and possible heat buildup under an insulating supercontinent, not to deep mantle plume heads. Long-term mantle insulation under a supercontinent, an entire shallow-level mechanism, may have surface effects similar to those predicted for “plume incubation” models. The three major, relatively young CFBs India hosts (Rajmahal, ~116 Ma; Indo-Madagascar, ~88-85 Ma; Deccan; 65-60 Ma) all formed during continental rifting followed by full breakup (between India-Australia, India-Madagascar, and India-Seychelles, respectively). Shallow-mantle geodynamic mechanisms, such as EDGE effects, continental mantle delamination, crack propagation through oceanic lithosphere, and crack-controlled melting and magma focusing, can well explain observations generally ascribed to plumes, such as “enriched” magma chemistry and systematic age progressions (Sheth, 1999b).

There is no petrological evidence that the sources of any Deccan lavas were “abnormally hot”. The short (1.0-0.5 MY) duration claimed by some for this grand episode is in conflict with recent data that suggest the total duration to have been of the order of 8-9 MY (Sheth *et al.*, 2001a,b). The eruption rates are not known to have been particularly high, and the large volumes erupted could be due to the great lengths of the eruptive fissures. No systematic age progression exists within the Deccan, and reliable ^{40}Ar - ^{39}Ar ages of 60-61 Ma for lava flows from Bombay (Sheth *et al.*, 2001a,b) are impossible to reconcile with a plume-model-based expected age of 60 Ma for the Laccadives Ridge, 1000 km south of Bombay. Also, if the ~69 Ma mafic dykes reported from Kerala, southernmost India (Radhakrishna *et al.*, 1994) do represent early Deccan-related magmatism, an entirely different, non-plume, passive, EDGE-model (Anderson, 1998a) for Deccan volcanism seems even more attractive. “Enriched” isotopic ratios such as higher-than-N-MORB values of $^{87}\text{Sr}/^{86}\text{Sr}$, encountered along the island/seamount chain southward from the Deccan and up to Réunion Island, are usually taken to be plume signatures, but such compositions may instead mark an involvement of shallow-level enriched continental mantle (Smith, 1993). High values of $^3\text{He}/^4\text{He}$ ratios also do not represent a deep mantle or plume component (e.g., Anderson, 1998b). The ~68.5 Ma alkaline complexes (Mundwara, Barmer) in the northern Deccan, previously ascribed to a plume (Basu *et al.*, 1993) on this basis, could derive from the continental mantle. The same may be true of 72-73 MY old mafic ophiolitic rocks outcropping in Pakistan, with Réunion-like chemical and isotopic compositions (Mahoney *et al.*, 2002). The systematic

age progression along the Chagos-Laccadive Ridge and up to Réunion Island, is due to southward crack propagation through the oceanic lithosphere, not to Indian plate motion over a plume (Sheth, 1999b). The narrow “hotspot track” may represent localized melting and magma focussing from a wider area (the “transform-fault effect”, Langmuir and Bender, 1984).

One interesting possibility needing exploration is whether eclogite could have been a source in part for the Deccan lavas. The rifted western continental margin of India follows the NNW-SSE Dharwar structural trend of the Precambrian southern Indian shield (e.g., Biswas, 1987), and the lines of breakup of India from Seychelles, and also Greater India (India + Seychelles) from Madagascar, may have been ancient sutures with large amounts of trapped eclogite and enriched mantle in them. If such eclogite constituted a major source for the Deccan, mantle fertility and not high mantle temperatures were important (see Yaxley, 2000). The ENE-WSW Narmada zone running along central India has been argued previously to represent an ancient suture between two protocontinents.

The interplay of the rift zones underlying the Deccan, which come together in west-central India, is apparently responsible for the roughly circular outcrop of the Deccan, resembling what is expected from a spherical plume head. Along the rift zones (the Cambay, Kachchh, Narmada-Tapi and Godavari rifts), EDGE effects (Anderson, 1998a), and crustal extension (Sheth, 2000) were important. The “hotspot track” on the oceanic crust, as argued, is similarly related to melting and magma focussing along a southward-propagating fracture. To conclude, non-plume, plate tectonic models are fully capable of explaining the Deccan in all its greatness.

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Geographical variations of mantle source fertility beneath Iceland

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The nature of mantle sources is partially recorded in the composition of oceanic island basalts. The record is, however, not always straightforward to decipher. The only direct evidences of mantle composition and lithology, which upon melting can produce mafic magmas, are the orogenic peridotitic massifs and mantle xenoliths. These samples suggest that the non-refractory mantle is principally composed of lherzolite with small but significant proportions of fertile pyroxenites. During upwelling of such a lithologically heterogeneous mantle, the pyroxenites should melt first and probably to a larger extent, because they would reach their solidus at a greater depth than the enveloping lherzolites. Preferential melting of garnet pyroxenite would produce melts characterised by high La/Yb and ($^{230}\text{Th}/^{238}\text{U}$) since garnet retains Yb and U relative to La and Th, respectively. Mixing of such melts with those generated from spinel lherzolite would result in lower La/Yb and ($^{230}\text{Th}/^{238}\text{U}$) in the mixture. If mantle plumes have hotter centre than their periphery then lower proportions of garnet pyroxenite melts may be expected in mantle derived basalts from the centre of the hot spots.

Recent basalts from the off-rift volcanic zone on the Snæfellsnes peninsula in western Iceland offer the possibility of investigating over 100 km, lateral compositional variability in a presumed mantle plume head if the magmas ascend vertically from their source region. The alkalinity of these basalts, as well as La/Yb, ($^{230}\text{Th}/^{238}\text{U}$) and $^{87}\text{Sr}/^{86}\text{Sr}$, decreases along the volcanic zone, from west to east, towards the centre of Iceland. These geochemical variations, therefore, are also correlated with the SiO₂ contents of the basalts, suggesting a link between major and trace elements and isotope ratios in their mantle source. These compositional variations are taken to indicate that melting of garnet pyroxenites generates a significant component of the basalts, in the Snæfellsnes volcanic zone. This component appears to decrease towards the centre of Iceland. There, the composition of primitive basalts in the rift zones may be dominated by melts of overwhelming lherzolithic source. This could be the result from a large extent of melting due to a higher temperature in the core of the presumed mantle plume.

The alkali basalts furthest to the west in the Snæfellsnes volcanic zone, having the highest La/Yb and ($^{230}\text{Th}/^{238}\text{U}$), or garnet signature, also have the highest $^{87}\text{Sr}/^{86}\text{Sr}$ and $^3\text{He}/^4\text{He}$ lower than 8 (R/Ra). These isotope ratios are compatible with the hypothetical garnet pyroxenites in the periphery of the Icelandic plume being recycled oceanic crust. The inference from this study is that garnet pyroxenite melts are likely to be involved in the genesis of most oceanic island basalts, but are most readily identified

in basalts from regions such as fracture zones or off-rift volcanic zones where geothermal gradient is likely to vary significantly. Finally, elevated ratios of Sr isotopes and La/Yb are better tracers for mapping mantle fertility than possible centres of plumes.

Chemical Variations and Melting Systematics along the Western Galápagos Spreading Center, 90.5° - 98°W

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From west to east along the Galápagos Spreading Center (GSC), there are remarkable co-variations in axial depth, axial morphology, crustal thickness and the composition of erupted lavas. West of 95.5°W the GSC has an axial valley, crustal thickness is about 5.5 km and nearly all erupted lavas are N- (normal) MORB. Between 95.5°W and ~93°W the axis shows transitional morphology, crustal thickness increases to ~6.5 km and erupted lavas are transitional (T-) MORB showing modest increases in K/Ti, Nb/Zr and radiogenic isotope ratios. Between ~93°W and the Galápagos transform zone at 90.5°W the axial region is dominated by axial high morphology, crustal thickness up to 8 km and the eruption of enriched (E-) MORB extending to maxima in K/Ti and Nb/Zr. In addition to enrichments in highly incompatible elements and radiogenic isotopes, E-MORB are characterized by high Na₂O and H₂O, and low CaO/Al₂O₃ and SiO₂ at a given MgO, features generally consistent with relatively low extents of partial melting. In an attempt to determine the relative contributions from enriched mantle, active upwelling, and potential temperature in the development of the crustal thickness and chemical gradients along the GSC, we have developed a melting equation that explicitly incorporates a deeper zone of hydrous melting into the decompression melting regime beneath the ridge axis. The model solves for a range of variables that are constrained to match the crustal thickness and concentrations of K, Na₂O, H₂O and Ti in lavas at various locations along axis. Modeled variables include the depths to the anhydrous and hydrous solidi, (sensitive to temperature and source water content, respectively), productivity in the hydrous region, source composition, and the flow rate of material passing through the hydrous zone. The results of this model indicate that incorporation of hydrous melting reduces the required mean extent of melting, even for N-MORB. The production of GSC E-MORB requires an enriched source, but the magnitude of this enrichment correlates inversely with the extent of active upwelling. Chemical heterogeneity including H₂O can account for most of the variation in crustal thickness and composition along the western GSC; the maximum potential temperature anomaly associated with this section of ridge is 34°C. Our preferred explanation for the eastern, shallowest, most chemically enriched portion of the axis is that it is produced by only a slight ($11 \pm 11^\circ\text{C}$) temperature anomaly coupled to an enriched source with moderately active upwelling.

Dynamics of the Iceland Plume: Recycling the Iapetus Ocean?

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In order to better understand plume-mantle dynamics, the origin and history of mantle material entrained in plumes needs to be well understood. This can be achieved by studying the geochemical signature of products of mantle plumes, i.e. basaltic lavas, placed in a context that is supported by geophysical evidence when available. Here we present a model for recycling of oceanic crust and lithosphere after subduction and mixing with a primitive mantle component. The model is based on the Nd-Os isotopic and major and trace element characteristics of the studied lavas in combination with constraints from previously published isotopic data and can explain the geochemical composition of both the current and past Iceland Plume.

The main sample area used for this study is the Snæfellsnes Transect, which runs from the Western tip of Snæfellsnes Peninsula to the Langjökull Volcanic System in the Western Riftzone, Iceland. This transect was not only chosen because of the systematic compositional changes, already indicated during previous studies of the post-glacial basaltic lavas along this transect (Jakobsson, 1972; Sigmarsson, 1992; Hardarson, 1993) and which might be related to their relative distance from the plume axis, but also because the lavas along this transect represent a snapshot of the underlying mantle in recent times and therefore represent a radial cross-section of the Iceland Plume. Additional samples used in this study are several primitive basaltic lavas from Þeistareykir, part of the Northern Riftzone.

The lavas that have been analysed can be divided into two main groups and a transitional group based on location and chemical composition. The composition of the axial group, represented by lavas from the main rift-zone, is dominated by melts from a mantle component with a strongly depleted trace element signature, a high ϵNd value ($\sim +10$), a superchondritic Os isotope ratio ($^{187}\text{Os}/^{188}\text{Os}$ is ~ 0.138) and a relatively unradiogenic Sr isotope signature ($^{87}\text{Sr}/^{86}\text{Sr}$ is ~ 0.7030). In contrast, the composition of the off-axis group, represented by most of the lavas from Snæfellsnes Peninsula, is dominated by melts from a mantle component with an enriched trace element signature relative to the axial group and a low ϵNd value ($\sim +5$), a sub-chondritic Os isotope ratio ($^{187}\text{Os}/^{188}\text{Os}$ is ~ 0.126) and a relatively radiogenic Sr isotope signature ($^{87}\text{Sr}/^{86}\text{Sr}$ is ~ 0.7034). The lavas from the transitional group have intermediate major and trace element signatures and isotope characteristics.

The data from this study form near perfect trends for trace element ratios and isotope ratios (see figure 1). These trends are considered to represent a mixing relationship between melts derived from the two main components with a superimposed melting effect.

All the studied lavas are thought to be derived from two main components within the Iceland Plume itself based on previously published diagrams (Thirlwall, 1995; Fitton et al., 1997; Kempton et al., 2000). Assuming that these two major plume components are co-genetic, a two stage LREE depletion has to take place to explain the observed $\epsilon\text{Nd} +5$ and $\epsilon\text{Nd} +10$ end member compositions. The first stage of depletion is by achieved by melting of PM, forming DMM. This is the mantle source for new oceanic crust and lithosphere. Both crust and lithosphere are then subducted and mix with a primitive mantle component resulting in the reservoir tapped by the Iceland Plume. The recycling of the oceanic crust is supported by the HIMU-like Ce/Pb ratios and K-depletion as described by Thirlwall (1997). Adding a

primitive mantle component is necessary in order to explain that the most depleted Nd isotope signature corresponds with a radiogenic Os isotope signature, while the less depleted Nd isotope signature corresponds with a subchondritic Os isotope signature. This is clearly in contrast with what would be expected as a result from simple recycling.

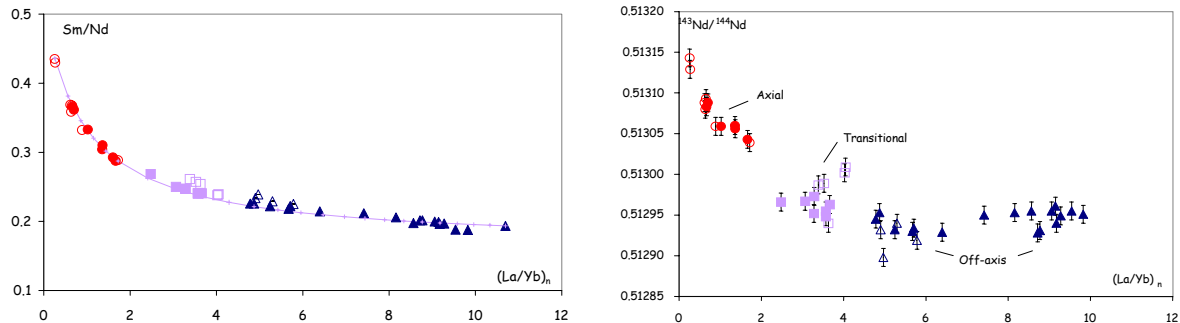


Figure 1. Two diagrams in which the near perfect trends described above are illustrated. Sm/Nd ratios (left) and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (right) are plotted against $(\text{La}/\text{Yb})_n$ ratios. The trends are thought to represent mixing with a superimposed melting effect.

The time constraints for this model are bracketed by a maximum model age of 1500 Ma, based on the negative $\Delta 7/4$ values found in Icelandic lavas (Thirlwall, 1997), and a minimum model age based on the calculated Re-depletion age of ~ 380 Ma. In order to explain current day REE and Nd and Os isotope characteristics of the Iceland Plume as well as the range found in older volcanic material attributed to the Iceland Plume (e.g. West Greenland) the actual time constraints for the model prove to be the most important factor. Modelling of the REE elements in combination with Nd and Os isotopes on the basis of these constraints suggests that formation of the recycled oceanic crust and lithosphere took place ~ 600 Ma ago, while subsequent subduction and mixing with a primitive mantle component (15% of the total volume) took place ~ 520 Ma ago (figure 2). This coincides particularly well with the formation and subduction of the Iapetus ocean.

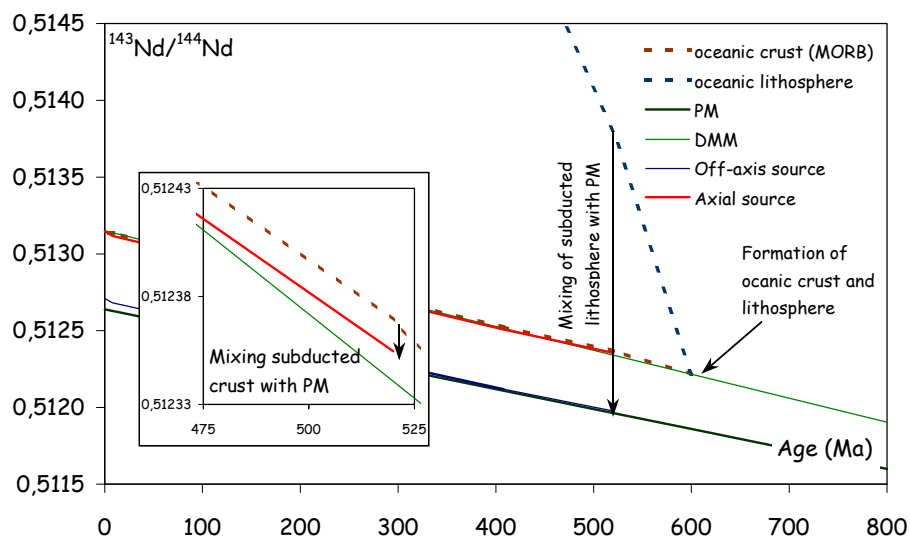


Figure 2. Evolution diagram illustrating the effect of mixing the subducted slab with a primitive mantle component on the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios and $^{187}\text{Os}/^{188}\text{Os}$ ratios through time for the two main mantle components that are sampled by the Iceland Plume.

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The Fate of Subducted Oceanic Crust and the Sources of Intraplate Volcanism

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Relative to other planetary bodies (Moon, Mars) the Earth's mantle shows a suppressed rate of isotopic evolution (Smith and Ludden 1989). The key difference between Earth and the other bodies is the operation of subduction. Such crustal recycling, as in marble-cake or streaky mantle models (Allègre and Turcotte 1986, Fitton and James 1986) would provide a satisfactory buffer to moderate mantle isotopic evolution. However, for two decades such processes have been largely ignored because of a requirement to accommodate the mantle plume model. The isolation of subducted crust in the latter (Fig. 1a) sets up a series of circular arguments, such as a perceived necessity to buffer the depleted mantle with plume residues (e.g. Morgan and Morgan 1999), which appear to make the plume model indispensable. However, with objective evaluation it becomes apparent that the logic used to mandate the plume model can not be justified and the model itself is unnecessary. One of the principal arguments used to substantiate the plume model and to reject alternatives, is based on the trace element ratios Nb/U and Ce/Pb. Similar ratios between MORB and OIB compared to crustal compositions have been taken as evidence that crust can not be recycled into the convecting mantle (e.g. Hofmann 1997) except for the possible exception of in the Indian Ocean mantle. Such logic is flawed because it fails to consider the possibility of formation of the depleted mantle MORB-source by remixing of recycled material with a more depleted component than the depleted mantle MORB-source. This hypothetical "very depleted" component can be equated with mantle from which crust has been extracted. Simple mass balance calculations indicate that Nb/U ratios comparable to those observed in the depleted mantle are only found when the entire volume of mantle contributes to crustal generation. However, extraction of continental crust from 30-50% of the mantle leads to the generation of very depleted residues with Nb/U ratios of 70 to 150. The Nb/U- Ce/Pb ratios in MORB can then be explained by a ternary mixing array between subducted sediment, recycled basaltic crust, and very depleted mantle.

The fate of subducted oceanic crust in any layered mantle regime is therefore remixing into the convecting mantle, not isolation into plume sources. The questions that then arise on adopting a streaky/marble-cake mantle structure, are to the nature and distribution of source components for intraplate volcanism and the causes of melting. Relative to MORB, the sources of such rocks are enriched in K. This compositional difference would appear to support generation of intraplate melts from sediment-rich streaks (Fig. 1b). However, if variations in recycled sediment type provide a suitable explanation for the large range of $^{176}\text{Hf}/^{177}\text{Hf}$ relative to $^{143}\text{Nd}/^{144}\text{Nd}$ seen in MORB, a large range of Hf isotopic variation should also be seen in intraplate basalts. The general lack of such a signature suggests K-enrichment is related to a process which does not fractionate Lu from Hf, such as the formation of volatile-bearing phases such as amphibole and phlogopite in the mantle wedge

at convergent margins (Fig. 1c). Concentrations of such minerals along collisional sutures may be particularly important in the generation of intraplate volcanism in opening ocean basins, where intraplate tracks invariably correlate with lineaments in the continental architecture (Smith 1993). In long-lived ocean basins, high pressure amphiboles may provide a source of potassic metasomatism, either on thermal equilibration with the mantle or if destabilised by fluids from deeply subducted carbonates. Volatiles-bearing components would also play a key role as fluxes to facilitate melting of streaky/marble-cake mantle such that intraplate volcanism can be generated without deep-seated thermal anomalies.

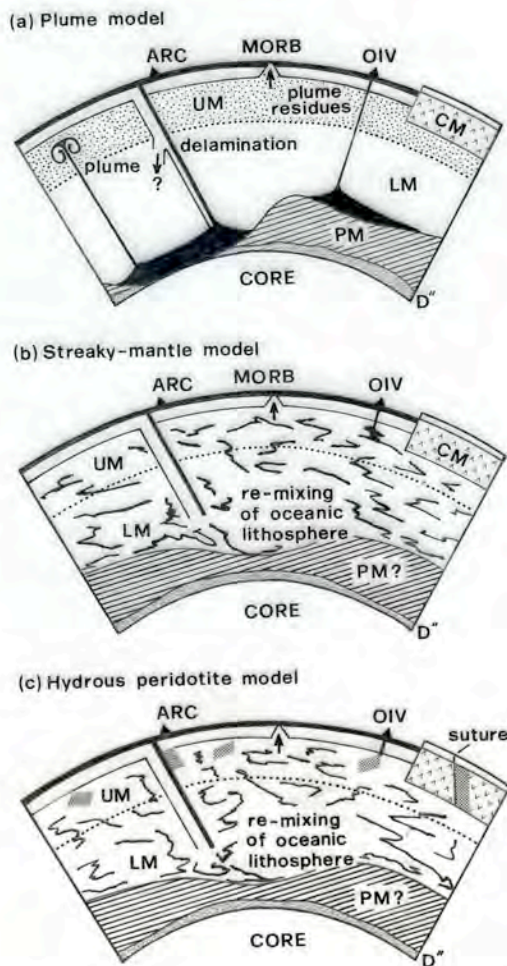


Figure 1. Fate of subducted oceanic crust and origin of ocean island volcanism (OIV) according to (a) plume (b,c) streaky/marble-cake mantle models. Abbreviations: UM upper mantle; LM lower mantle; CM continental mantle; PM primitive mantle.

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The Regular Distribution of Intraplate Volcanism in the Pacific Basin

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Before the concept of hotspots/plumes was widely invoked, Jackson and Shaw (1975) proposed the distribution of intraplate volcanism in the Pacific basin was controlled by plate tectonic processes producing changes in the stress field of the plate. At the time such concepts could not be proven as the bathymetry of the ocean floor and the tectonic history of the basin margins were poorly known. With the increasing popularity of the mantle plume concept, interpretations switched to emphasis of a random distribution of volcanism generated from deep-seated thermal anomalies. However, few examples of intraplate volcanism in the basin have been shown to conform to the predictions of the plume model. Potentially only three examples (Easter-eastern Mid Pacific Mountains, Louisville-Ontong Java, Marquesas-Hess/Shatsky) possess the oceanic plateau-island chain sequence as expected from the plume head-tail model (Clouard and Bonneville 2001). Non-linear age progressions are found in many island chains, including the Cook-Austral-Marquesas, Marshall-Gilberts, and Line Islands. Even the Hawaiian-Emperor chain, generally considered the archetypical example of plume volcanism, does not the features expected in the plume model. The volcanism lacks an associated plateau, eruption rates have increased rather than decreased over time, and paleomagnetic evidence (Sager 2002) indicates any hotspot could only have been stationary for half its history.

Correlations between volcanic output along the ocean island chains can now be correlated with basin-wide plate re-organisations such as at 25 and 5 Ma (Kamp 1991) suggesting the stress field model was essentially correct more than a quarter century ago. Reconstruction of the evolution of the Pacific basin (Fig. 1) demonstrates that oceanic plateaus were generated in zones of tension in the wake of retreating triple junctions, and that ocean island chains may be divided on the basis of propagating- or leaky- fracture origin. The latter, including the Louisville, Marshall-Gilbert, Line Island, and Cook-Austral-Marquesas chains are those characterised by non-linear age progressions. Such volcanism followed pre-existing NNW–SSE trending fracture zones such as the Kashima-Eltanin and Emperor-Easter megatrends (Smoot 1999). The fracture zones form part of a pattern of orthogonally intersecting lineaments initiated by transform faulting along ridge systems during the early history of the Pacific plate. Volcanism that can be attributed to propagating fractures includes the Sala y Gomez, Juan Fernandez, and Caroline chains which extrapolate to breaks in nearby subducting slabs suggesting stressing of the plate by convergent margin geometry. The Emperor chain is unique in the stress field model in representing volcanism along a propagating fracture induced at a divergent margin. The location and orientation of this chain is attributed to the geometry of the Kula-Pacific ridge following plate re-organisations at 82 Ma which prematurely halted triple-junction volcanism on Meiji seamount. Subsequent volcanism along the Hawaiian chain can be explained by re-orientation of the stress field to control by convergent margin geometry following abandonment of the Pacific-Kula ridge, and does not require a change in Pacific plate motion at the time of the Hawaiian-Emperor bend (43 Ma).

The distribution of Pacific intraplate volcanism is therefore more regular than would be expected than in the plume model, and can be explained as a result of shallow

volatile-bearing sources tapped under developing hotcell conditions, with the change from oceanic plateau to island chain volcanism reflecting changes in the stress field as the Pacific plate changed from having an intra-oceanic setting bordered by ocean ridge systems, to subducting beneath the basin margins (Smith 2003).

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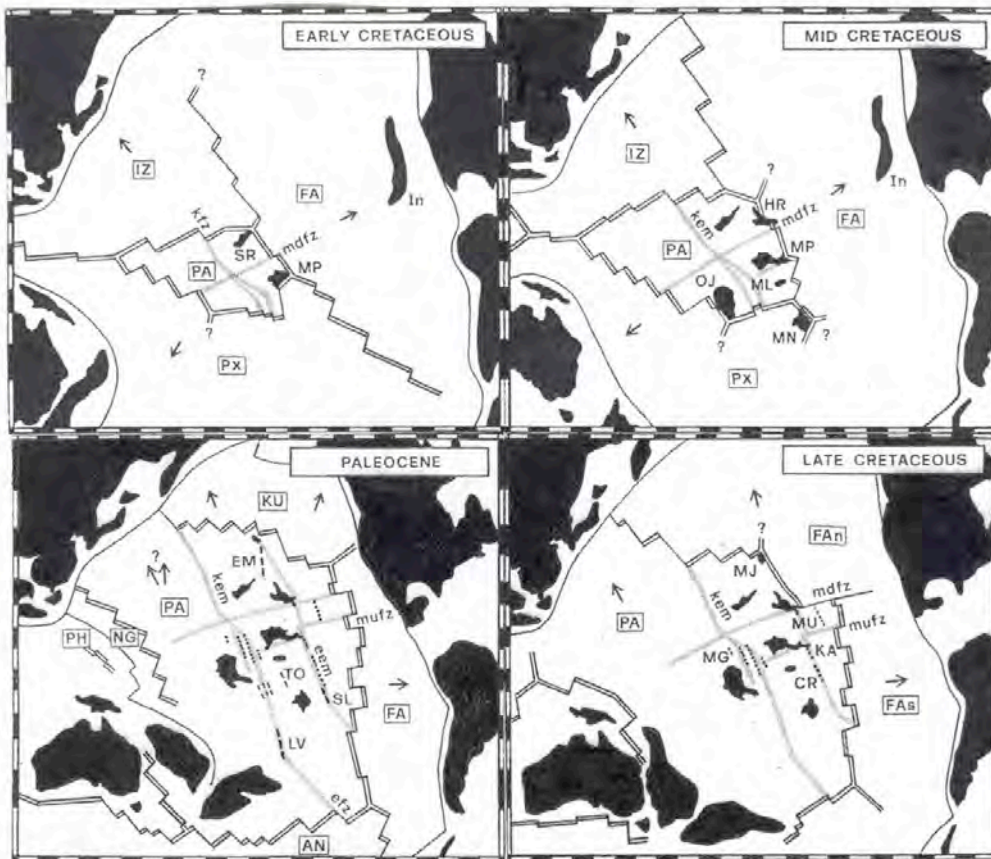


Figure 1. Evolution of the Pacific basin from the Early Cretaceous through Paleocene illustrating the relationship of oceanic plateaus to triple junctions and ocean island chains to pre-existing lineaments (after Smith 2003). The lineaments (grey) form an orthogonal pattern reflecting their origin as predominantly as transforms along spreading centres bounding an initially rhombic Pacific plate. Plates FA Farallon (FAn north, FAs south), IZ Izanagi, KU Kula, NG North New Guinea, PA Pacific, PH Philippine, PX Phoenix. Lineaments: kem Kashima-Eltanin megatrend, eem Emperor-Easter megatrend, mdfz Mendocino fracture zone, mufz Murray Fracture zone. Oceanic plateaus: SR Shatsky Rise, MP Mid Pacific Mountains, ML Magellan Rise, OJ Ontong Java, MN Manahiki, HR Hess Rise, MJ Meiji. Ocean island chains (dashed and dotted lines): MG Marshall-Gilbert, CR Christmas Ridge, KA Karin, MU Musicians, SL South Line, TO Tokelau, LV Louisville, EM Emperor.

Venus as a Mantle Plume Laboratory, S. E. Smrekar¹ and E. R. Stofan², 1) Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109; 2) Proxemy Research, 20528 Farcroft Lane, Laytonsville, MD 20882.

The similarity in the size and bulk density between Venus and Earth give them a similar capacity for heat production. Yet Venus shows no evidence of plate tectonic. Instead, it experienced wide spread resurfacing approximately 750 m.a. (1), possibly driven by global crustal overturn (2,3). The rate of geologic activity declined following this resurfacing, leading to the hypothesis that Venus is currently in a stagnant lid convective regime (4-6).

Despite the differences in tectonic style, Venus has many Earth-like hotspot rises (e.g. 7-9). Seven volcanic rises are very similar to terrestrial hotspot rises, in that they have extensional rifts, large shield volcanoes, broad topographic swells, and gravity anomalies suggesting deep compensation (10-11). These are the primary hotspots, with a possible origin as plumes rising from the core mantle boundary. A recent analysis of the data available for terrestrial hotspots suggests that a limited number of primary plumes arise from the core-mantle boundary on the Earth (12). Other secondary plumes are generated as when a super plume impinges on the upper mantle-lower mantle boundary spawning smaller thermal instabilities, or are due to local rifting and melting (12,13). Within this definition, Venus and Earth appear to have a comparable number of primary plumes.

Secondary plumes on Venus have a very different character on Venus than on Earth. Coronae are believed to form over small-scale upwellings. There are over 500 coronae, with 95% having a diameter between 100 and 400 km (14). Although most coronae have associated volcanism, they are defined on the basis of their annulus of fractures and their topographic morphology. Although the volcanism, radial extensional fractures, and the dome or plateau morphology found for many coronae are consistent with typical models of mantle upwelling, coronae differ in key ways. They are typically smaller and have a range of topographic forms, with nearly half of all coronae having interior depressions. Fracture annuli are not observed at larger hotspots either on Venus or Earth.

A variety of models have been proposed to explain the unusual topographic morphology of coronae. Koch and Magna (15) proposed a spreading drop model to form some of the interior depression topographic forms. A model in which an upwelling plume lead to delamination of the lower lithosphere at the edges of the plume explains most topographic forms (16). A plume impinging on a depleted mantle layer can generate surface depressions (17).

One possible explanation for the lack of coronae on Earth is that the presence of a low viscosity zone under the oceanic lithosphere causes the plume head to spread laterally and dampen any surface topography (16). This would be particularly pronounced for small plumes. Under continents, small plumes might be unable to deform the lithosphere, particularly in the presence of a depleted mantle layer. Alternatively Jellenik et al. (17) propose that coronae form on Venus due to the absence of subducting slabs. Without slabs to cause enhanced thermal gradients at the core mantle boundary, smaller scale plumes form. The effect of buoyancy can also create different scale plumes (18).

Recent work examining the admittance signature for coronae suggests the possibility that the density structure of the lithosphere plays a key role. Smrekar et al. (21) examined the admittance signature for those coronae that have a fracture annuli 50% or less

complete, defined as Type 2 coronae (22). No relationship is seen between either crustal or elastic thickness and diameter, as suggested by prior models (15,20). Instead, the elastic and crustal thickness correlate with some topographic morphologies. Rim only coronae, which are predicted to form through isostatic rebound (16), have a more limited range of estimated crustal thickness (50-100 km) than other topographic forms (19). Rim only coronae are predicted to form rimmed depressions, once isostatic and thermal equilibrium are reached. These coronae typically have a bottom loading signature and relatively large elastic thickness values, consistent with formation via delamination of the lower lithosphere.

Additionally, nearly half of the coronae in the gravity survey appear to be isostatically compensated (21). All of the topographic forms considered in the gravity survey are represented in the isostatically compensated group. These coronae are interpreted to be inactive, implying that all morphologies can represent the final state of a corona (19). This includes topographic forms such as domes and plateaus that are typically assumed to indicate the presence of a plume at depth. In fact, none of the plateaus and domes studied had a relatively thin elastic lithosphere, large depth of compensation, and bottom loading signature that is consistent with a plume at depth (21).

The results of the gravity survey suggest that processes such as delamination and isostasy play a significant role in the formation and compensation of coronae. Some coronae may even form via delamination without a plume to initiate the process. Type 2 coronae, which were examined in the gravity survey (21), are more commonly found in the plains than along fracture belts. The plains regions may be tectonically inactive areas, which could favor the transition of basalt to eclogite (14,21). The presence of a high-density layer at depth would tend to favor delamination.

A stagnant lid regime results in a lithosphere that does not cool monotonically. Instead the mantle heats up over time, causing the lithosphere to remain at a constant thickness or even thin with time. The tectonic stability of the lithosphere in such a regime would allow slow phase transitions to occur, favoring delamination and allowing isostasy to be achieved over time. These processes may be key to understanding why coronae form only on Venus, the effects of a stagnant lid on tectonic history, and the relative contribution of upwelling plumes to the formation of coronae and heat loss on Venus.

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Spots yes, hot barely or not

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Sea floor heat flow data play a role in discussions of hotspots and plumes similar to that of the dog whose failure to bark helped Sherlock Holmes locate the missing racehorse Silver Blaze. The small size or absence of a heat flow anomaly at midplate hotspots such as Hawaii has been crucial for assessing possible mechanisms causing the volcanism and uplifted topography. The uplift was originally thought to reflect a hot mantle plume, rising from deep in the mantle, penetrating the approximately 100-km-thick oceanic lithosphere and causing heating to about 50-75 km of the surface (e.g., Crough 1983, McNutt and Judge 1990). The thermal perturbation from such thinning predicts heat flow significantly higher than that associated with unperturbed oceanic lithosphere, which cools with age as it spreads away from the midocean ridges where it formed. Although anomalously high heat flow was initially reported, subsequent analysis showed that most if not all of the apparent anomalies resulted from comparing the data to reference thermal models that underestimated heat flow elsewhere (Von Herzen et al. 1989, Stein and Abbott 1991, Stein and Stein 1993, 1994). Hence subsequent models generally assume that the uplift results from the dynamic effects of rising plumes (Liu and Chase 1989, Sleep 1994, and the associated compositional buoyancy, whose thermal effects are concentrated at the base of the lithosphere and hence would raise surface heat flow at most slightly, because tens of millions of years are required for heat conduction to the surface.

Given Foulger's (2003) challenge to the common model in which Iceland's formation is due to a deep mantle plume, we examined sea floor heat flow data from the region. We find (Stein and Stein, 2003) that heat flow values near Iceland on the North American (west) side of the mid-Atlantic ridge are comparable to those for oceanic lithosphere of this age elsewhere, and thus show no evidence for significantly higher temperatures associated with a mantle plume. Heat flow is significantly higher on the Eurasian plate, east of the mid-Atlantic Ridge, than to the west on the North American plate. This puzzling asymmetry, which appears to be real even given the limited data, is opposite that expected from models in which Iceland was formed by a fixed mantle plume.

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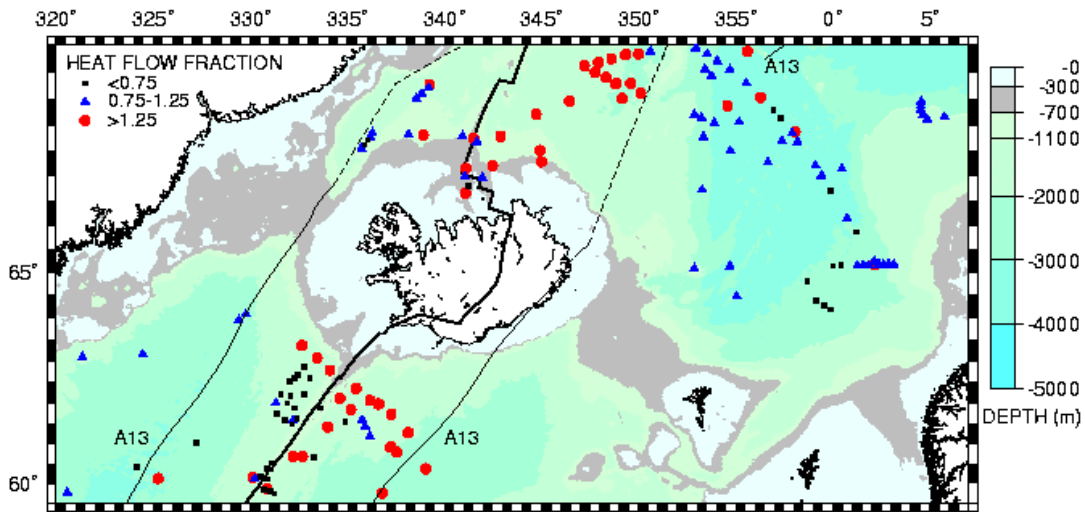


Figure 1: Bathymetry and heat flow for the Iceland region. Heat flow shown as heat flow fraction, observed values normalized by global average values for that lithospheric age (Figure 2). Lithosphere younger than about 35 Myr indicated by the positions of magnetic anomaly 13 (solid line) or approximated by dashed line.

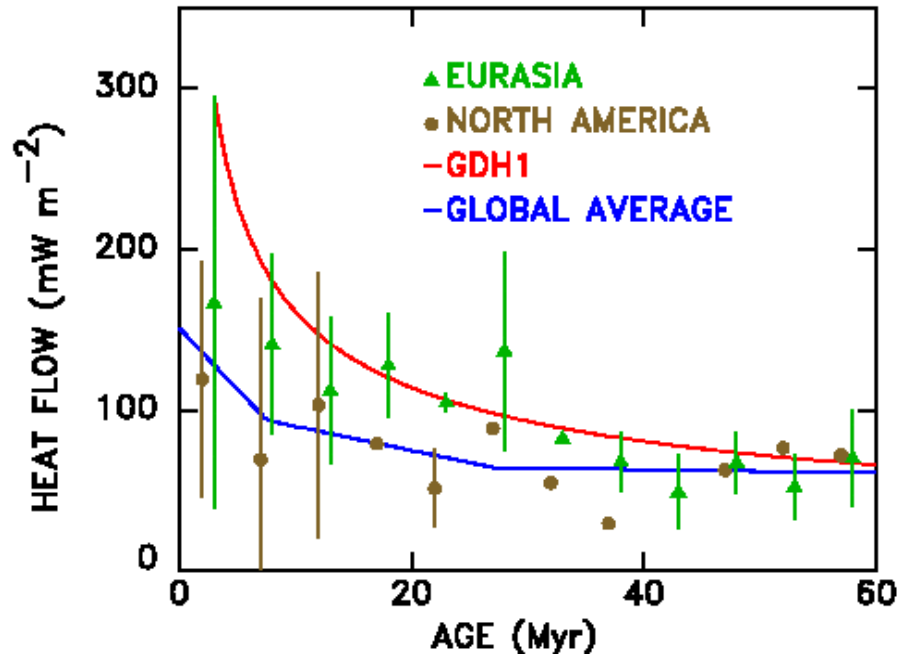


Figure 2: Heat flow data near Iceland (Figure 1) grouped in 5-Myr bins for the Eurasian and North American plates, compared to the predictions of the GDH1 thermal model which does not include the effect of hydrothermal circulation, and a linear fit to the global average values.

The size and fate of the Pan-African Plume Mantle

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Overview

The upwelling of deep mantle material to the shallow mantle in the form of large plume-heads has been considered as a major mechanism of heat and material transfer to the Earth surface. It was suggested that the rising plume-heads were responsible for production of large igneous provinces such as the oceanic plateaus, and growth of continental crust during major orogenic events [1]. It is not clear, however, to which extent the uppermost mantle is affected by the rising plumes. Are they regional (in the sense of hot spots) or have a “global” impact? What size of the mantle is occupied by the plume material and for how long the plume survives there before being mixed and consumed by the “depleted asthenospheric mantle”?

Here, I focus on the evolution of the late Proterozoic Pan African continent (comprising the basement of several segments of the Phanerozoic Gondwana), and its parental “plume mantle”. The lithospheric mantle of the Pan African continent provided the sources of alkali basalts that erupted ubiquitously over Gondwana during the Phanerozoic. I use the distribution of the alkali basalts to estimate the mass of the juvenile lithospheric mantle and the size of its parental “plume- mantle”.

Juvenile Pan-African crust of Gondwana

In the early Phanerozoic time, the continental crust of the Earth was clustered in two large supercontinental masses: Gondwana and Laurentia. The early Paleozoic Gondwana (the continental masses of Africa, South America, Arabia, India, Australia, Antarctica and New Zealand) comprised Archean to late-Proterozoic crustal terrains that had been affected by the late Proterozoic orogenic processes that are loosely termed as the “Pan African orogeny” (lasted roughly between ~900-500Ma). This included mobilization of the pre-existing crust, metamorphism and generation of juvenile crustal terranes. The juvenile Pan-African crustal rocks are characterized by initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios of 0.7027 ± 3 , and ϵNd values of +4 to +6. The relatively low $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios indicate mantle-type sources for the crustal magmas, while the initial ϵNd values are significantly lower than those expected from late Proterozoic depleted (MORB type) mantle source (the ~800 Ma Gabal Gerf ophiolite in Sudan shows $\epsilon\text{Nd} = +8$). Stein and Goldstein[2] suggested that the juvenile ANS magmas were derived from enriched plume-related mantle that was transformed by the subduction mechanism to new continental lithospheric. Late Proterozoic crustal magmas with Nd-Sr isotopic compositions that are similar to the ANS magmas were documented in several other late Proterozoic juvenile crustal

terrains of Gondwana suggesting that the formation of enriched “plume mantle” during the Pan African orogeny was global.

Alkali basalts from the Pan African lithospheric mantle

Phanerozoic alkali basalts overlie many of the Pan African crustal terrains (Fig. 1). An important characteristic of the Phanerozoic basaltic magmatism over the Gondwana is their large spatial distribution and short contemporaneous activity. Alkali basalts with similar geochemical and isotopic characteristics erupted simultaneously over thousands of kilometers (e.g., over more than 5000 km along the Antarctica-New Zealand-Tasmania continent and along more than 3000 km in the ANS and over North Africa). This configuration cannot be easily explained by conventional plume models, but can be accommodated by models of “fossil plume” reactivation (e.g. [3-5]). Stein and Goldstein [2] linked the production of the Phanerozoic basalts in the ANS to the lithospheric mantle that was produced during the Pan African events. Extending this model, I suggest that the “fossil plumes” or the lithospheric mantle roots beneath the Gondwana fields are parts of the “Pan African” lithospheric mantle that was fossilized when subduction processes on the Pan-African continental margins stopped. The important consequence is that the basalts provide information on the distribution of the Pan African lithospheric mantle, and in turn on the properties of its parental plume mantle.

The composition of the uppermost mantle during the plume event

The remarkable uniformity in the Nd and Sr isotope compositions of the Phanerozoic alkali basalts from various parts of the Gondwana (grand averages of available data: $\epsilon_{\text{Nd}}=4.7\pm 0.7$ and $^{87}\text{Sr}/^{86}\text{Sr}=0.7028\pm 3$) is analogous to that observed in Mid-Ocean Ridge Basalts (MORB). The uniformity in MORB compositions has been used as the major argument in the widely accepted supposition that MORB represents the asthenospheric upper mantle. **It can similarly be argued that, at the time of formation of the Pan-African juvenile crustal terranes and the lithospheric sources of the Gondwana basalts, the enriched, plume type mantle material largely dominated the uppermost mantle.** This requires a substantial supply of enriched and fertile material to the uppermost mantle on a global scale. Stein and Hofmann [1] proposed that continental crust growth is associated with episodic large upwelling events in the mantle (MOMO overturn events). During these events, lower mantle material rises as large plume-heads, producing oceanic plateaus that may be later accreted to the existing continents. The rising plumes replenish the upper mantle in incompatible trace elements, and make it fertile for basalt formation. Thus, the “mid-ocean ridges” of late Proterozoic time produced oceanic crust with enriched chemical composition.

The uniformity in the Nd and Sr isotopic compositions of the Phanerozoic Gondwana basalts stands in contrast to the much larger variation in the isotopic values of the oceanic island basalts (OIB). The heterogeneity in OIB compositions has been attributed to the large variety of deep mantle sources and their different histories or to shallower mantle sources (e.g., delaminated segments of heterogeneous lithospheric mantle). However, the role of enriched-plume related oceanic lithosphere (of the Nauru basin type) as an important component in the production of OIB has been overlooked.

The size and fate of the Pan African “plume-mantle”

Considering a static model, the mass of the "plume mantle" during juvenile lithosphere formation is equivalent to the sum of the masses of juvenile lithospheric mantle + juvenile continental crust + residual depleted mantle. This simplified configuration can be described by the following mass-balance equations:

$$(1) M_m = M_{RDM} + M_L$$

For the total masses of the “plume mantle”, lithosphere and residual depleted mantle:

$$(2) M_m * C_m = M_{RDM} * C_{RDM} + M_L * C_L$$

For the mass of specific elements:

By rearranging eqs. 1 and 2, one gets:

$$(3) M_m = M_L (C_L - C_{RDM}) / (C_m - C_{RDM})$$

Where:

M_m is the mass of "plume mantle "

M_L is the mass of lithospheric mantle + crust

C_L is the concentration of a trace element (e.g., Nd) in the lithospheric mantle and crust

C_{RDM} is the concentration of a trace element in the residual depleted mantle

C_m is the concentration of a trace element in the "plume mantle"

The major unknown here is M_m -the mass of the “plume- mantle”. The mass of the juvenile lithospheric mantle is estimated from the distribution of the Phanerozoic basalt fields over Gondwana (Fig. 1), assuming a lithosphere thickness of ca. 100 km.

The volume of the Pan-African juvenile lithospheric mantle is estimated to be in the order of $3-4 \times 10^9 \text{ km}^3$ and its parental "plume mantle" has the approximate volume of $1-2 \times 10^{10} \text{ km}^3$, which is less than 10% of the volume of the upper mantle. If distributed uniformly over the globe the Pan African "plume- mantle" would occupy only the uppermost < 50 km of the mantle. If the plume material is not uniformly distributed, but rather spreads beneath “the Gondwana hemisphere”, the “plume mantle” may extend to deeper depth. In any event this estimate indicates that the upwelling of the Pan-African plume-head(s) represents a limited event that could not affect the composition of the entire uppermost mantle, and therefore cannot be described as a complete overturn of the mantle. Rather the process can be defined as a "transient intrusion event" involving large upwelling of lower mantle material to the shallower part of the mantle. This view is consistent with some geochemical constraints on the input of lower-mantle material into the upper mantle. For example, strontium isotope composition of MORB limits the total time integrated exchange between the lower and upper mantle to ca. 30% of the lower

mantle mass and the ^{40}Ar budget suggesting that ~50% of the total radiogenic argon produced during Earth history is still stored in the lower mantle [6,7].

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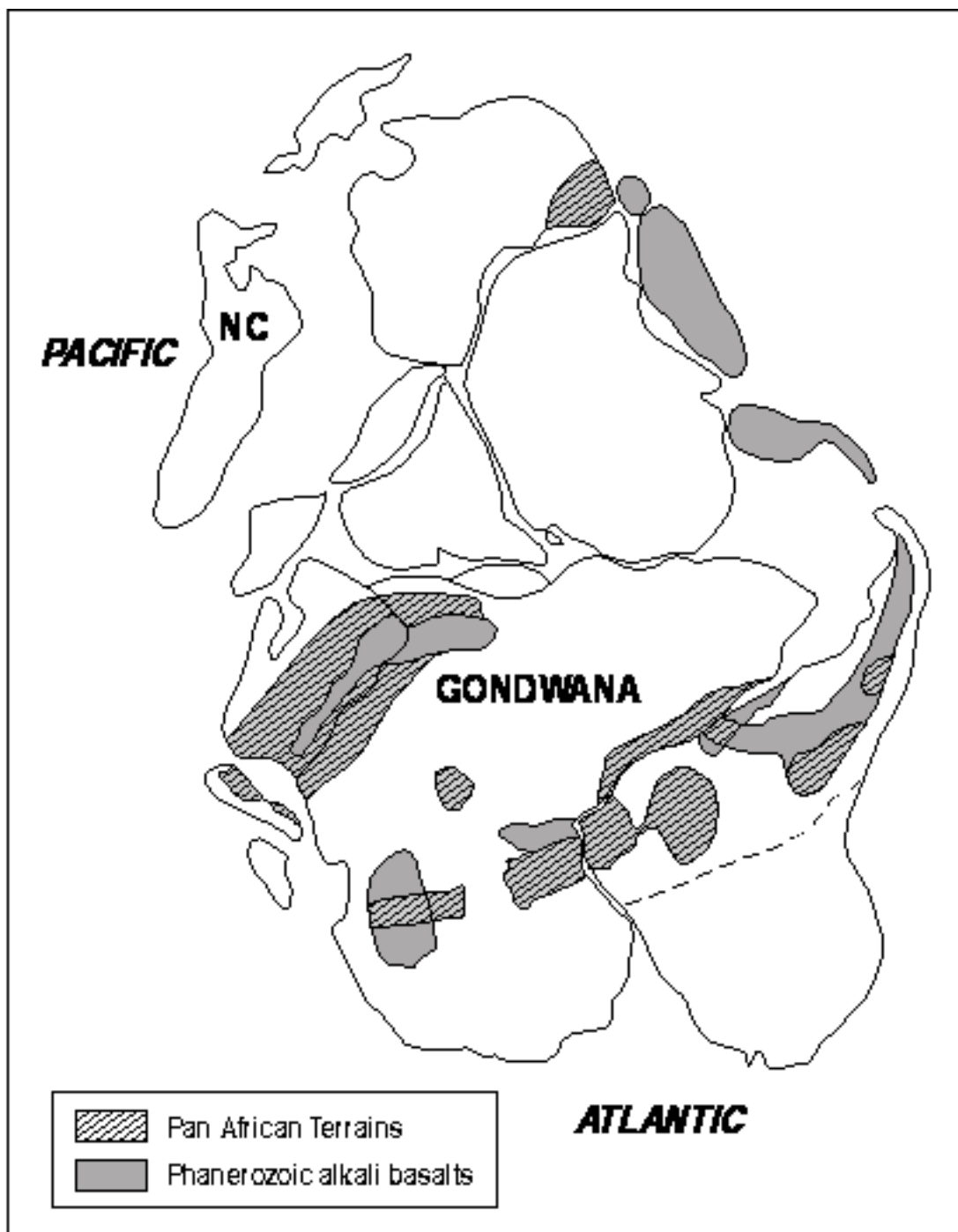


Fig. 1. Distribution of Pan African juvenile lithosphere terrains

Morphology and Distribution of Hotspots on Venus. Ellen R. Stofan¹ and Suzanne E Smrekar². ¹Proxemy Research, 20528 Farcroft Lane, Laytonsville, MD 20882, ²Jet Propulsion Laboratory, California Institute of Technology, Pasadena CA 91109.

Magellan spacecraft data have revealed that the surface of Venus is dominated by features of volcanic origin, many of them likely to be related to upwelling (e.g., Solomon et al., 1992). Patterns of tectonic and volcanic features failed to support pre-mission models of plate tectonics (e.g., Head and Crumpler, 1987), and the average surface age of 300-750 my [Schaber et al., 1992; McKinnon et al., 1997] hinted at a possible history of global-scale resurfacing (e.g., Strom et al., 1992). Thermal evolution models of the planet range from periodic plate tectonics [Turcotte, 1993] to stagnant lid convection [Solomatov and Moresi, 1996]. We have been investigating the characteristics of possible hotspot features on Venus, which include topographic rises (~1500 km across, 1.5-5 km high), coronae (100-2400 km across, 0.8-2 km high), and large volcanic provinces (up to 6000 km across). The morphology, evolutionary history and distribution of these features can provide insights into the behaviour of mantle upwelling on the terrestrial planets. We interpret the variations in styles of surface volcanism, surface deformation, topography and gravity signatures to indicate differences in the nature of the underlying thermal upwellings.

Topographic rises on Venus have diameters of at least 1000 km and have large associated volcanic edifices (e.g., Senske et al. 1992; McGill, 1994; Stofan et al. 1995). Stofan et al. [1995] classified rises into three categories: volcano-dominated, rift-dominated and corona-dominated. Of the ten rises identified to date, five are volcano-dominated, two are rift-dominated and three are corona-dominated. Topographic rises appear to be randomly located on Venus. Minimum melt volumes, calculated from edifice volumes are estimated to be at least 10^4 - 10^6 km³. Some of the volume of the topographic rise may include early-stage flood basalts, but we have seen no clear evidence of this. The different categories of topographic rises do not appear to indicate an age progression, but instead reflect the variations in lithospheric structure, plume characteristics and regional tectonic environment [Stofan et al., 1995; Smrekar et al., 1997]. In particular, analysis of the geologic histories and gravity signatures of corona-dominated rises indicates that they might form from thermal anomalies originating from a relatively shallow interface, possibly the upper-lower mantle boundary [Smrekar and Stofan, 1999]. The characteristics of volcano and rift-dominated rises are most consistent with formation by 'primary plumes' (e.g., Courtillot et al., 2003).

Coronae are volcano-tectonic features believed to form over small-scale mantle upwellings [Basilevsky et al., 1986; Stofan et al., 1991; Janes et al., 1992]. Our updated corona database contains 513 features [Stofan et al., 2001; Glaze et al., 2002], an increase from the 326 coronae of the previous survey [Stofan et al., 1992]. Coronae occur in three distinct geologic settings: along chasmata (rifts), in the plains, and at volcanic rises [Stofan et al., 1997]. Coronae overlap in size and association with large amounts of volcanism with large volcanic edifices (e.g., Herrick and McGovern, 2000). While some large volcanoes are located at topographic rises, other features appear to be randomly located in the plains or along chasmata, similar to coronae. We are in the process of assessing the relationship between large volcanoes and coronae, in particular comparing their histories and gravity signatures.

We have identified ten characteristic topographic signatures for coronae, and compared them to model predictions [Smrekar and Stofan, 1997]. About half of coronae are depressions, some are plateaus up to 3 km high, and others consist of rimmed plateaus or rims surrounding flat interiors. The wide range in topography at coronae has led us to conclude that models involving thermal upwelling accompanied by delamination best fit corona characteristics [Smrekar and Stofan, 1997; Smrekar and Stofan, 2003]. Contrary to other models (e.g., McGovern and Solomon, 1998), we do not find that elastic lithospheric thickness appears to be a controlling factor in corona morphology [Smrekar and Stofan, 1999; 2003]. The relationship between corona topographic shape and size is consistent with the model predictions of Smrekar and Stofan [1997], in which later-stage coronae (depressions and rimmed depressions) are smaller due to the inward migration of the delaminating ring [Glaze et al., 2003]. Coronae are likely to be products of secondary plumes (e.g., Courtillot et al., 2003).

A number of large igneous provinces, or major concentrations of volcanic and volcano-tectonic features have been identified on Venus, most notably the Beta-Atla-Themis (B-A-T) region ($\sim 10^6$ km²) [Crumpler et al., 1997]. The provinces range from the B-A-T region with its distinct concentrations of large flow fields, coronae, and small to intermediate-size volcanoes, to groups of arachnoids (volcanic depressions surrounded by ‘webs’ of fractures), to large flow fields. Over 200 large flow fields have been identified and classified on Venus [Lancaster et al., 1995; Magee and Head, 2001]. Mylitta Fluctus is one of the larger fields, with an area of 300,000 km² and estimated thickness of 250-400 m [Roberts et al., 1992]. Most of the venusian flow fields are areally more extensive and less thick than terrestrial large flow provinces [Roberts et al., 1992; Lancaster et al., 1995]. The regions appear to be more common in the equatorial region of the planet. Although all of these regions have been suggested to be associated with thermal upwellings, little detailed analyses of these regions or their gravity signatures have been done. Magee and Head [2001] did note a strong association between rift zones and flow fields, suggesting that lithospheric extension may be critical to their formation. On Venus, the lack of plate motion might cause flow fields to be superposed by ongoing volcanic activity associated with the plume tail. The venusian large volcanic provinces must have plumes that have sufficient buoyancy fluxes to result in large amounts of melting, but either lack sufficient tails to create more numerous topographic rises or are in an early stage of development.

Our analyses of the geologic and geophysical characteristics of hotspots on Venus suggest that the features vary in stage of evolution, and likely originated at different times over the visible history (~ 750 my) of the planet (e.g., Guest and Stofan, 1999). Our work to date on topographic rises and coronae suggests that the variations in size, morphology and distribution of features related to thermal upwellings primarily reflect fundamental differences in the characteristics of the plume/thermal upwelling. Most topographic rises are likely to be formed by primary plumes, while coronae and corona-dominated rises result from shallower upwellings. Extension clearly plays a critical role in the formation of large flow fields [Magee and Head, 2001], coronae [Stofan et al., 1992; Stofan et al., 1997] and possibly the rift-dominated volcanic rises.

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Mantle source composition, melting regime and mantle flow in the NE Atlantic

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Covariations of radiogenic isotope ratios and trace elements ratios in Icelandic and NE Atlantic basalts indicate their formation by progressive melting of heterogeneous mantle sources. Alkaline basalts from the volcanic flank zones in Iceland, Jan Mayen and Vesteris Seamount are formed by low degrees of partial melting of the heterogeneous mantle and extracted mostly from the fertile source components. The enriched, alkaline melt fractions are gradually diluted by melts from increasingly depleted and refractory source components as melting progresses. Olivine tholeiites of the Icelandic rift zones and nearby ridges are formed by advanced melting of depleted sources.

The NE Atlantic tholeiites and alkaline basalt share important geochemical features that are distinct in a global perspective, including high $^{87}\text{Sr}/^{86}\text{Sr}$ for a given $^{143}\text{Nd}/^{144}\text{Nd}$, low $^{207}\text{Pb}/^{206}\text{Pb}$ and high U/Pb and Nd/Pb. The low-degree alkaline melts differ from the tholeiites by their higher concentrations of large ion lithophile elements and steep chondrite-normalized REE-patterns. The predominant fertile and refractory components have HIMU-tendencies in the form of high U/Pb and U/Th ratios and high concentrations of Nb, Ta and other HFS-elements. Both components have $^{207}\text{Pb}/^{204}\text{Pb}$ ratios below the Northern Hemisphere Reference Line and low to moderate $^{206}\text{Pb}/^{204}\text{Pb}$ ratios, indicating an origin as recycled oceanic lithosphere of Paleozoic age, probably from subducted Iapetus ocean floor.

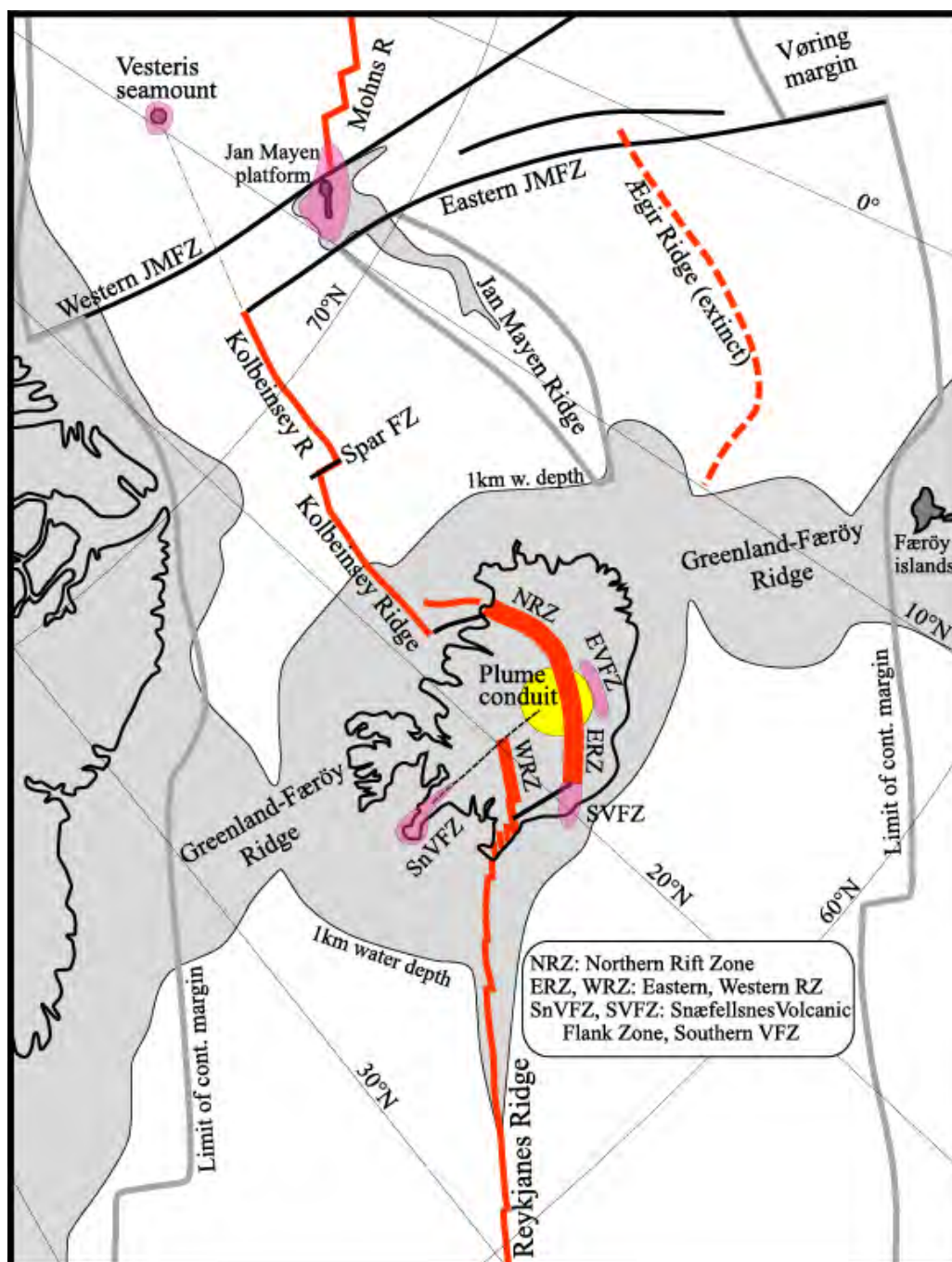
The fertile and refractory components appear to represent upper and lower portions of recycled oceanic lithosphere, respectively. This inference is based largely on O-isotopic data consistent with hydrothermal seafloor alteration at low and high temperatures, producing low $\delta^{18}\text{O}$ values in the upper basaltic crust and low $\delta^{18}\text{O}$ in the lower oceanic lithosphere with cumulate sequences and residual mantle harzburgite. The primitive rift zone tholeiites are further characterized by positive Sr- and Eu-anomalies in chondrite-normalized REE-diagrams, indicating that their mantle sources contain significant components of plagioclase-bearing cumulates. Another feature of alkaline and tholeiitic basalts from Iceland, Jan Mayen, Vesteris Seamount and most of the NE Atlantic ridge segments is positive ΔNb -values. The common geochemical features including the Sr-, Nd-, Pb-isotopes and ΔNb -values apply to the northern Reykjanes (north of 61°N), Mohns and Knipovich ridges. The basalts along the southern part of the Mohns ridge grade compositionally into the enriched and alkaline Jan Mayen lavas. The Kolbeinsey ridge basalts, however, tend to be very refractory with low $^{87}\text{Sr}/^{86}\text{Sr}$, high $^{143}\text{Nd}/^{144}\text{Nd}$, unradiogenic Pb and negative ΔNb . Extensive depletion of material that has first flowed and melted along the Northern Rift Zone (NRZ) may explain the loss of some of the Iceland plume signatures along the Kolbeinsey ridge.

Decompressional melting occurs as the mantle source ascends vertically beneath the spreading axes. Geophysical and geochemical data also indicate the importance of lateral (subhorizontal) mantle flow along the ridges away from a deep vertical flow channel under the northwestern margin of Vatnajökull. Although seismic tomography models are barely providing sufficient resolution to identify narrow plume conduits, the Iceland plume appears to extend only to the base of the mantle transition zone. The combination of vertical, oblique and subhorizontal flow of progressively melting material produces a thermal and compositional mantle structure that may be further constrained by geophysical and geochemical data. Based on the propagation of V-shaped ridges along the spreading axes from Iceland, the lateral plume flow is inferred to reach at least 1000 km towards NE and SW. The shallow flow of suprasolidus mantle channeled along the ridges, is most likely accompanied by a flow of mostly subsolidus and more fertile mantle deflected more evenly in all directions from a deep level (200-300 km depth ?) of the Iceland plume stem.

The alkaline and transitional basalt volcanism of the Eastern Volcanic Flank Zone (EVFZ) can be explained by its location near the eastern periphery of the Iceland plume stem. The rift zone configuration leads to a shallow deflection of the plume flow mainly towards NE, W and SW, resulting in a minor supply of undepleted material towards E and SE. The Southern VFZ forms the propagating tip of the Eastern RZ. The southwesterly supply of deflected plume material is considerable, but shallow and extensive melting of refractory mantle is confined to the ERZ. Further south the melting zone is terminated at a deeper level, resulting in a higher proportion of enriched melt fractions.

The Snæfellsjökull volcanic system at the western tip of the Snæfellsnes VFZ and the Jan Mayen island are located 300 and 830 km from the Iceland plume stem, respectively. The alkaline volcanism and large volcanic productivity anomalies at these locations are inconsistent with passive ascent of a relatively shallow and depleted mantle source. Supply of material with substantial amounts of enriched components through a deeper melting zone is required. The different tectonic settings of Snæfellsjökull and Jan Mayen, however, indicate different forms of vertical mantle flow component.

The initial wet solidus intersection, probably at more than 200 km depth in the central part of the Iceland plume stem, results in efficient partitioning of H into low-degree, but mobile melt fractions. The resulting viscosity increase in the dehydrating residue may inhibit the flow of the plume stem, leading to a deep-seated partial deflection of a nearly undepleted mantle source. This deeply deflected source may rise obliquely with a major westward flow component along the transform structure of the Mid-Iceland Belt and experience its main solidus intersection under Snæfellsjökull. Deeply deflected flow spreading out in all directions may become subhorizontal at distances of 300-500 km from the plume conduit. Beneath Jan Mayen the subhorizontal flow will be deflected upwards because of plate separation along the southern part of the Mohns ridge, inducing decompressional melting of the enriched source.



Sea-floor basement morphology: Distinguishing hotspot effects from plate tectonic effects—Examples from Iceland and the Azores

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Whatever processes create “ridge-centered” hotspots (e.g., Azores, Iceland, and Galapagos) also modulate the oceanic crust formed by seafloor spreading. (A thicker crust and off-axis volcanism are two features commonly attributed to hotspot modulation). Much has been learned about the thickness, structure and composition of the oceanic crust where the Mid-Oceanic Ridge axis passes over or near a hotspot-generating anomaly in the mantle below the crust. However, it is the UPPER surface of this crust—the seafloor topography at the spreading axis, which becomes “basement topography” once buried under sediments—that is most readily characterized at high spatial resolution by multibeam bathymetry, sidescan sonar, or reflection seismology. (Where oceanic crust emerges above sea level, optic and radar imaging reveal this upper crustal interface at even greater resolution, but erosion by water and glacier ice (Iceland) scrape off the corresponding morphology except for recent volcanic eruptions and fault scarps).

We investigate two MOR segments (one extinct) for clues as to what morphology is due to ordinary plate tectonics and what reflects hotspot influence or modulation of this process. Basically, features which are commonly also observed on “normal” MOR segments can scarcely be “blamed” on hotspot influence. In addition, we offer an example of a rare, if not unique morphological feature near Iceland that seems still to have been created by plate tectonic processes unrelated to hotspots. We sidestep the question of whether these morphologic clues have any bearing on the question of origin depth of a possible Iceland or Azores mantle plume. We leave open the possibility that the Azores and Iceland are different “species” or even “genera” in the hotspot family.

We focus on the extinct Aegir Ridge, active ca. 55-25 Ma just north of the Iceland hotspot (1), and on the active primary accreting plate boundary on the Azores Plateau (Terceira Rift; 2,3). These plate boundaries are comparable in terms of very slow opening rates and rift valley width and depth (see below). However, the Aegir Ridge did not actually cross the Iceland-Faeroe Ridge, and should therefore be considered the equivalent of the modern Reykjanes or Kolbeinsey Ridge in relation to the paleo-Iceland hotspot. Furthermore, no dramatic central volcanic complexes (such as those forming the volcanic islands and seamounts along the Terceira Rift) were developed along the Aegir Ridge during its 30 Ma of existence (1).

Opening rates along the Aegir Ridge ranged from 8 mm/a near Iceland to 13 mm/a in the northern Norway Basin during the period 55-36 Ma, but must have decreased to lower and then zero rates by ca. 25 Ma, the probable time of extinction (1). We nominate the Terceira Rift (2,3) as the slowest opening organized spreading boundary along the modern MOR system: Although the rates are too slow to have been “recorded” by magnetic lineations, they can be calculated from plate motion closure about the Africa-North America-Eurasia triple junction. The rates

calculated from NUVEL-1A (4) range from 4.4 mm/a near the triple junction to 3.7mm/a at the intersection of the Terceira Rift with the GLORIA transform.

The Aegir Ridge rift valley ranges from 40 to 50 km in width, somewhat wider than most active slow-spreading ridges. Partially sediment-filled, the basement valley is up to 3000 m deeper than the adjoining rift mountains, i.e., a greater relief than normal for slow-spreading active ridges. We attribute the wide, deep valley to slow spreading and possible slow extension after spreading ceased. The southern Aegir Ridge is oblique to the opening direction, but individual rift valley wall escarpments are normal to the calculated opening direction. None of the above features are atypical of slow and/or extinct rifts far from hotspot influence. Furthermore, despite 100% coverage by multibeam bathymetry, no diachronous basement features similar to those first reported by Vogt from the Reykjanes Ridge (5,6) and Kolbeinsey Ridge (7) are apparent in the morphology.

A remarkable, possibly unique basement ridge (Treitel Ridge) was discovered as a result of detailed mapping of Aegir Ridge. Treitel Ridge (manuscript in prep.) is a narrow (ca.100 km long; up to 1000 m basement relief), asymmetrical basement ridge on the western flank of Aegir Ridge closest to Iceland. Within the resolution of magnetic lineations, Treitel Ridge formed at the Aegir axis about Chron 18n time, and appears to be a morphological “topochron”. Its narrow form and steeper inward (towards the Aegir Rift axis) slopes further distinguish it from the V-shaped (diachronous) ridges (5,6; also called “chevrons”(8)) prominent on the present Reykjanes Ridge and its flanks. The location of Treitel Ridge on the Aegir Ridge flank closest to Iceland invites attribution of the feature to some ca. 40 Ma influence of the Iceland hotspot, but we favor a plate tectonic explanation: At about anomaly 18 time, spreading between Greenland and North America ceased (9). This “forced” a change in plate motion between Greenland and Eurasia (see Foulger, these Proceedings). A new rift began to propagate along the Greenland margin, creating a separate Jan Mayen microplate (10), whose eastern margin was the Aegir rift axis. The major transform that existed at the southern end of Aegir Ridge was no longer parallel to the opening direction. We propose that Treitel Ridge registered this adjustment to the new plate motion required by the extinction of Ran rift in the Labrador Sea. We speculate that formation of this narrow volcano-tectonic ridge represents evidence that the extinction of Labrador Sea spreading was relatively abrupt.

The only feature of Aegir Ridge that seems to require hotspot influence is the relatively shallow basement depth, particularly along the southwestern rift mountains, which rise to basement depths of 2000 m (shallow for 25-30 Ma crust!) near the intersection of Aegir rift with the Iceland-Faeroe Ridge. These shallow depths are indeed on the “Iceland” side of the rift. However, the exact location of the Iceland hotspot “center” at this time (ca. 25-30 Ma) is not that certain. Furthermore, the existence and sense of rift mountain asymmetry (higher on the western flanks) is similar to that on ridges farther and even remote from Iceland hotspot influence (11).

In the case of Terceira Rift, the rift valley depth is comparable to the central MAR. The valley widths are somewhat wider than for the MAR, but this may reflect the ultra-slow spreading and therefore thicker axial mechanical lithosphere. The spacing of magmatic centers (ca. 100 km) is comparable to that found along the ultra-slow spreading Southwest Indian and Gakkel ridges, most of whose length is far from hotspot influence.

Only the AMPLITUDE of this topographic segmentation (2000-4000 m) greatly exceeds that of “normal” slow MAR segments and registers the more abundant magma supply that has created the Azores Plateau. The location of the Terceira rift near the northeast margin of the plateau suggests the plateau was formed by successive NE jumps of the rift axis. This is consistent with the existence of a relative stationary Azores mantle “hotspot” across which the Gripps-Gordon (12) model predicts a SW motion (20mm/a) of the Terceira Rift (2).

In summary, we find that most morphologic features of the Aegir Ridge and the Terceira Rift reflect plate tectonic processes similar to those at other slow-spreading segments of the MOR, far from hotspot influence. For the Aegir Ridge, only the high southern rift mountain topography appears to be of hotspot origin. For the Terceira Rift, the great along-axis volcanic center relief, and the off-axis volcanism and elevated plateau topography, is “hotspot” in morphology. However, the present “absolute” motion of the Terceira rift in a relatively fixed hotspot frame (12) is consistent with progressive northeast jumps of the rift – “attempting” to remain over the hotspot—to form the present Azores Plateau (2).

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ISOTOPIC DETECTION OF POSSIBLE CORE-MANTLE INTERACTIONS IN PLUME SOURCES: RULES OF ENGAGEMENT. Richard .J Walker, Department of Geology, University of Maryland, College Park MD 20742. rjwalker@geol.umd.edu

If some plumes arise from the core-mantle boundary *and* there is limited chemical interaction between the core and mantle within D", it is possible that these plumes could contain a unique chemical or isotopic fingerprint that is characteristic of the core. Unequivocal identification of a core component in a plume could, thus, potentially settle a major question in geodynamics. The ideal tracers of a core component in a plume would be elements that occur in very high abundance in the core relative to the mantle and crust, such that a very modest transfer of matter from the core to the mantle would result in a resolvable geochemical signature. The most sensitive suite of elements for geochemically identifying possible core contributions to plumes that may arise from the core-mantle interface are the highly siderophile elements (HSE = Pt, Re, Os, Ir, Pd, Ru, Rh, Au), and certain moderately siderophile elements (MSE) such as Ag, W and Mo. Because of the extreme preference of the HSE for metal relative to silicates, the formation of the core nearly quantitatively sequestered the Earth's HSE, and also likely dominates the budget of moderately siderophile elements. Mass balance is, therefore, potentially optimal for detection of core additions to the mantle.

Formation of the inner core may have fractionated the HSE, as occurred in early and late stage differentiates of asteroidal cores (e.g. Cook et al., in review) and as is predicted from experimental studies (e.g.. Jones and Malvin, 1990). We have previously suggested that the coupled ^{187}Re - ^{187}Os and ^{190}Pt - ^{186}Os isotope systems may be useful in identifying the presence of an evolved outer core component in mantle-derived rocks (Walker et al., 1995; Brandon et al., 1999). This hypothesis is based on conclusions from the study of asteroidal core crystallization (e.g. Cook et al., in review) and experimental studies (e.g. Walker 2000) indicating that Pt, Re and Os have become fractionated from one another in the outer core as a consequence of inner core crystallization. If a significant inner core formed sufficiently early (e.g. within 2 Ga of planetary formation), the outer core may exhibit coupled ^{187}Os and ^{186}Os enrichments of >7% and >0.01%, respectively, relative to chondrites.

Coupled enrichments in ^{186}Os - ^{187}Os have been detected in intrusive rocks related to the *ca.* 250 Ma Siberian flood basalt event, the *ca.* 89 Ma Gorgona Island (Colombia) komatiites, and in some rocks generated recently by the Hawaiian plume (Walker et al., 1997; Brandon et al., 1999; Brandon et al., 2003)(**Fig. 1**). All three igneous systems are generally assumed to be associated with large mantle plumes. We have argued that the coupled

enrichments detected are not consistent with the presence, via recycling, of any presently characterized crustal materials in the source regions of these magmatic systems, or via contamination of the magmas within the crust.

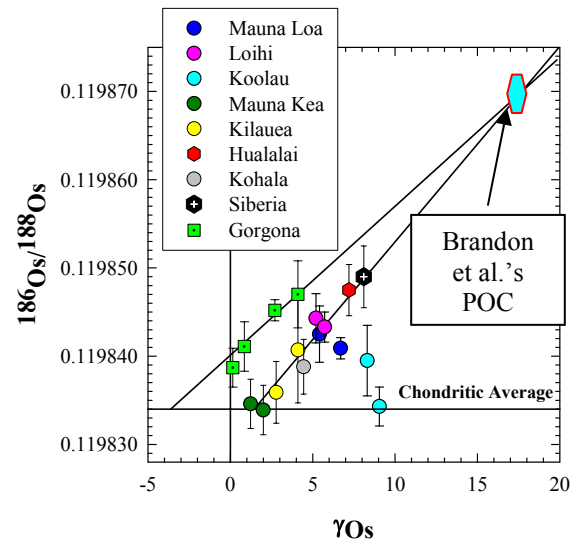


Figure 1. Correlated variations in $^{186}\text{Os}/^{188}\text{Os}$ relative to initial γ_{Os} (initial $^{187}\text{Os}/^{188}\text{Os}$ relative to the bulk solar system average). Brandon et al. (2003) concluded that the two trends may converge (POC: point of convergence) on a common endmember with $^{186}\text{Os}/^{188}\text{Os} \approx 0.119870$ and $\gamma_{\text{Os}} \approx +17$.

The ^{186}Os - ^{187}Os “test” for core material, however, is not without its weaknesses. It requires relatively early crystallization of the inner core, and processes other than core-mantle interaction could be envisioned to explain our observations. Additional data and new methods will be necessary to more convincingly demonstrate or refute core-mantle interactions.

Decay of so called “short-lived” nuclides during the first few tens of Ma of solar system history almost certainly led to the generation of isotopic compositions in the core that are resolvable from the silicate earth. Consequently, short-lived isotope systems are the best hope to confirm a core signature implicated by Os isotopes (Os isotopes alone can never “prove” the existence of a core component in mantle-derived rocks). It is even possible that a short-lived system can uniquely identify a core component. It is also important to note that a potential short-lived tracer for core-mantle interaction has an advantage relative to ^{186}Os - ^{187}Os in that the magnitude of the anomaly would not be time dependent (all growth of the daughter isotope is complete

within approximately the first 50 Ma of the birth of the solar system). Hence, if the short-lived systems can be demonstrated to be viable indicators of core contributions to modern plumes, then the contributions to ancient magmatic systems could also potentially be examined.

There are a variety of nuclides with approximately 10^6 to 10^7 year half-lives that were seeded into (or created within) the solar nebula just prior to the initiation of condensation. The positive identification of the decay products of short-lived systems with HSE or MSE, including ^{107}Pd - ^{107}Ag and ^{182}Hf - ^{182}W , have been carefully documented for a variety of meteorites and components within meteorites (*cf.* Chen and Wasserburg, 1996; Lee and Halliday, 1996). Anomalies for the ^{98}Tc - ^{98}Ru , ^{99}Tc - ^{99}Ru and ^{97}Tc - ^{97}Mo systems could also be useful, however, our recent work suggests that Tc did not exist in sufficiently high abundance to be useful for this application (Becker and Walker, 2003; Becker and Walker, in review).

Of greatest importance here is that recent work on the ^{182}Hf - ^{182}W isotopic systematics of chondrites (Yin et al., 2002b; Kleine et al., 2002; Schoenberg et al., 2002a) has convincingly refuted earlier claims that the Earth has a chondritic W isotopic composition (Lee and Halliday, 1996). Instead, chondritic meteorites have ϵ_{W} values (part per 10,000 deviation of $^{182}\text{W}/^{184}\text{W}$ relative to terrestrial standards) of approximately -2. This is potentially very important for identification of core components because the earlier results required rather late core formation and a minimal difference between the isotopic composition of W in the core and mantle. The new results require that the mantle and core have different W isotopic compositions (presumably combined they must have a chondritic composition). Mass balance requires that the core must have an ϵ_{W} value that is slightly less than -2, compared to 0 for mantle rocks. Thus, sufficient transfer of core W to a plume could generate a resolvable negative ϵ_{W} value in the mixture. The magnitude of the deviation would depend on the proportion of W transferred, and this in turn is dependent on the ratio of W in the core to W in the mantle. Unfortunately, W is apparently an MSE and is likely not as highly concentrated in the core as the HSE (though this bears additional testing). Assuming the core has W concentrations 10 times that of the mantle, transfers of 0.2 to 1% core material to a plume, as has been proposed to explain ^{186}Os - ^{187}Os enrichments, would result in ϵ_{W} deviations from the terrestrial standard of 0.2 units or less. The possible level of heterogeneity is just within the resolution capabilities of modern mass spectrometers.

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Upper Mantle Physical State and Lower Crustal Igneous Input: A Test of Current Models with Data from the U.S. Great Basin

Extended Abstract for
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Introduction

Hotspot activity and continental rifting have motivated numerous models for physical state of the upper mantle prior to extension and the importance of igneous material input to crustal growth. Upper mantle conditions and dynamics are believed to determine the amount and characteristics of igneous crustal input; alternately, igneous underplate properties provide clues as to the mantle physical state and fluxes which generated the underplate (e.g., Holbrook et al., 2001; Korenaga et al., 2002; Trumbull et al., 2002). The tectonically active southwestern United States interior (Humphreys and Dueker, 1994; Parsons, 1995) exemplifies one of three types of mantle hotspot in the classification of Courtillot et al. (2003). Within there, the eastern Great Basin spanning central Nevada to central Utah resembles the early-middle stages of margin formation, and thus allows formulation of tests for globally applied models of upper mantle physical state and its controls on subcrustal magmatism at volcanic continental margins, such as the Greenland-Iceland province or the central Atlantic continental margins (Lizzaralde and Holbrook, 1997; Holbrook et al., 2001; Ernst and Buchan, 2002).

Igneous Underplate and Mantle Physical State

Research on margins over the past decade has yielded a new appreciation of the large volumes of magmatic material produced at most rifted continental edges (e.g., White and McKenzie, 1989; Menzies et al., 2002). The amount and composition (and thus seismic properties) of magmatic underplating that occur during rifting depend on three factors: mantle potential temperature (T_p), active upwelling ratio (χ), and thickness of the lithospheric lid (defined by degree of lithospheric thinning (β)) (Kelemen and Holbrook, 1995; Holbrook et al., 2001; Korenaga et al., 2002) (Figure 1). Magma volume increases with T_p and β and decreases with increasing lid thickness. V_p increases with T_p , decreases slightly with increasing lid thickness, and is relatively insensitive to β .

The simplest magmatic state under the eastern Great Basin (EGB) would be of negligible underplate (Figure 1a). Surface divergence rate equals mantle upwelling rate ($\chi = 1$) which is passive stretching as defined by White and McKenzie (1989). Mantle thermal state shows limited change from that just following the general exhumation of the Great Basin by Middle Miocene time, presumed to be the average current mantle adiabat (ACMA; Thompson, 1992), as discussed

later, with $T_p \approx 1280^\circ\text{C}$. The melting window is narrow and shallow, and average velocity is medium (6.9-7.0 km/s; Korenaga et al., 2002). This state is compatible with modeling of White and McKenzie where little underplate (<1 km) is produced for $\chi = 1$ until the late rift stages near final oceanic breakthrough, which the EGB has not reached. There are no significant high-T plumelike upwellings.

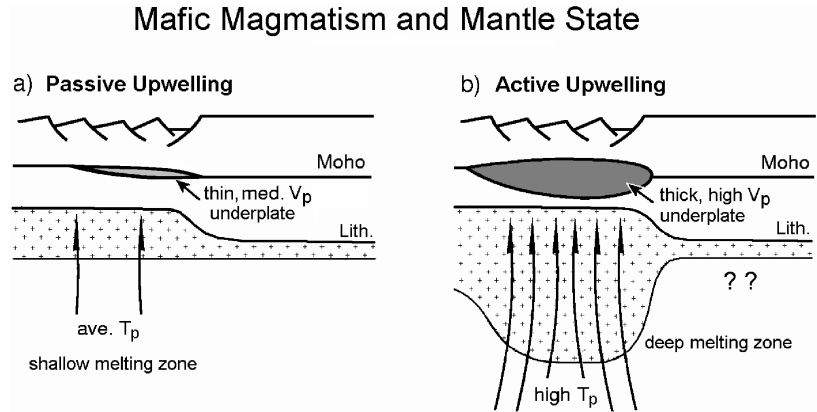


Figure 1. Modes of rift magmatism, applied to the eastern Great Basin: a), thin, moderate V_p (6.9-7 km/s) rift pillow forms in response to normal T_p and passive upwelling; b), rift pillow of substantial thickness and high V_p (>7.3 km/s) forms in response to high T_p and active upwelling in upper mantle.

At the other extreme, suppose material of higher T_p than ACMA was introduced below the EGB in the Late Cenozoic, and a high active upwelling ratio χ has existed pumping a greater amount of source material through the solidus (Figure 1b). Underplate in this case could be thick (10 km or more) and of high V_p (7.3 km/s or more for $T_p > \text{ACMA} + 200^\circ\text{C}$) because the solidus is intersected at greater depth, allowing a broader, deeper melting interval for a given degree of mantle upwelling and melt compositions with increased MgO and reduced SiO_2 (White and McKenzie, 1989; Kelemen and Holbrook, 1995; Korenaga et al., 2002). Concentrated upwellings of high-temperature (plume?) mantle material currently are advanced as the means of producing significant quantities of pre-rift or syn-rift magma occurring before oceanic breakthrough at continental margins (Menzies et al., 2002).

Intermediate scenarios are possible, of course. Passive upwelling of high T_p material could generate a limited amount of high- V_p underplate. High active upwelling ratio may form a moderate degree of underplate but of only medium V_p . Thick, high V_p underplate under the easternmost GB is suggested by four lines of evidence, each of which is quite tentative, as discussed next.

Mantle Physical State, Evidence on Crustal Underplating, and Mode of Extension in the Eastern Great Basin

The Late Cenozoic-present extension in the Great Basin shows a concentration at its eastern margin (Smith et al., 1989; Wannamaker et al., 2001). Correspondingly, Quaternary basaltic magmatism shows a clear N-S trend in extended western Utah evolving from alkalic compositions characteristic of ancient metasomatized mantle lithosphere to tholeiitic magmas

more akin to fertile, asthenospheric input (Hawkesworth et al., 1995; Nelson and Tingey, 1997; Wannamaker et al., 2001). The crust in the eastern GB appears to have thinned by ~ 2 since the mid-Miocene (Wernicke, 1992; Wannamaker et al., 2001), meaning rather limited mantle material would have been pumped through the melting interval in a passive mode. Lithospheric-scale simple shear rifting and other modes of NUE can produce high-degree mantle thinning (χ) locally promoting melt production and underplating (Buck and Su, 1989; Harry and Bowling, 1998; Boutelier and Keen, 1999). Degree of mantle thinning also may be cryptic to crustal indicators.

Great Basin extension initiated in north-central Nevada but quickly spread throughout the province, with broadscale exhumation and differentiation from the Colorado Plateau in the Early-Middle Miocene (Zoback et al., 1994; Stockli, 1999; Dumitru et al., 2000). The broadscale upper mantle thermal regime in the region likely is close to the current global average adiabat ACMA (Thompson, 1992) (Figure 2). From V_p and V_s estimates, Goes and van der Lee (2002) advance a geotherm for the central GB which is close to ACMA. Likewise, a deep electrical resistivity profile for the central GB compared to conductivity of dry lherzolite indicates that the geotherm there should be no greater than ACMA (Wannamaker et al., 1997, 2002).

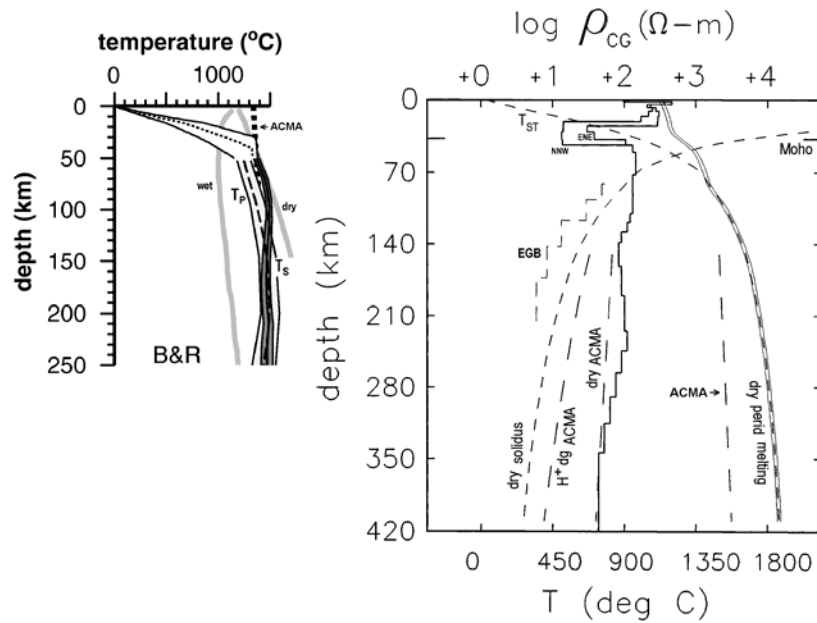


Figure 2. Left: Mantle thermal profile for central Great Basin from seismic tomography by Goes and van der Lee (2002), interpreted to be close to ACMA. Right: Mantle resistivity and thermal profiles in same region from deep MT data (Wannamaker et al., 2002). Curves toward right are temperature, curves toward left are resistivity. “Dry ACMA” is resistivity predicted for dry lherzolite and ACMA geotherm. “ $H^+ dg$ ACMA” is with hydrogen defects in random olivene at reduced water activity (0.15). “Dry solidus” is unmelted lherzolite resistivity at dry solidus temperatures. ENE and WNW denote the principal directions and magnitudes of anisotropic resistivity in the lower crust. EGB is resistivity profile for eastern GB near latitude $39^\circ S$ (Wannamaker et al., 2001).

Furthermore, given that central GB mantle resistivity is consistent with ACMA temperatures and dry lherzolite, the upper mantle there appears largely absent of elements which

would alter geophysical properties from their dry peridotite state (cf. Karato, 1990). Original volume of low-melting components there apparently was insufficient to create its own resolvable underplate in the central GB (Holbrook, 1990; Satarugsa et al., 2000); the same is presumed throughout the GB prior to the modern concentration of rifting in the EGB.

Since the Late Miocene (~10 Ma), extension and volcanism have concentrated near the GB margins, in particular the relatively straight tectonic margin of western Utah (Christiansen and Yeats, 1992; Wernicke, 1992; Wannamaker et al., 2001). Though debated, evidence is perhaps stronger that highly non-uniform extension (NUE) occurs under the easternmost GB and its transition zone (TZ) with the Colorado Plateau based on heat flow, Curie depth, some crustal thickness models, earthquake travel times, and intracrustal conversions (Bodell and Chapman, 1982; Wernicke, 1985; Smith et al., 1989; Sheehan et al., 1997; Wannamaker et al., 2001; Gilbert et al., 2003; but see Pakiser, 1989) (Figure 3). High-Vp “rift pillow” material of 7.4-7.5 km/s in the 20-35 km depth interval there is suggested from earthquake travel times and tomography, and receiver functions. Garnet should be present to increase velocity only for depths >30 km.

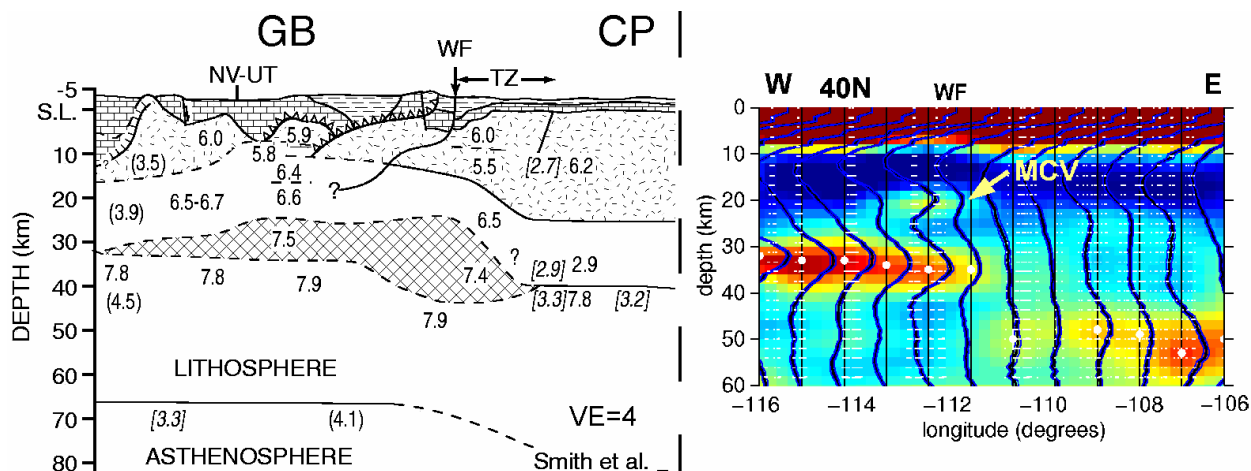


Figure 3. Left: crustal cross section through west-central Utah redrawn from Smith et al. (1989). P-wave velocities (km/s) are in bold, S-wave in parentheses, and densities (gm/cc) are in italics. Right: West-east cross-sections of common conversion point (CCP) stacked receiver functions in the western U.S. at approximately the latitudes of our proposed transects (Gilbert et al., 2003). Radial P waves are scaled to unity. Red color indicates positive polarity conversion, blue color represents negative. Stacked receiver functions shown in black, thinner blue lines are 1 s.d. derived from bootstrap resampling. Seismic rays shown as dashed white dotted lines converging to projections of instrument locations. Depths calculated assuming $V_p = 6.4$ km/s and $V_p/V_s = 1.73$. Mid-crustal converter beneath Wasatch Front (WF) is MCV.

Although various mantle tomographic images show the lowest upper mantle velocities and P and S delays under the easternmost GB and TZ (Humphreys and Dueker, 1994; Savage and Sheehan, 2000; Dueker et al., 2001; Lastowka et al., 2001), upper mantle temperature estimates have not been advanced to test the relation between mantle state and lower crustal underplating. However, a high conductivity concentration under the TZ implies mantle involvement in large-scale NUE to depths of 300 km or more (Figure 4). If dry peridotite compositions apply, as some evidence suggests, conductivity implies that temperatures reach the solidus with generation of some melt. Since this solidus is significantly hotter than ACMA, the EGB temperature profile represents a high temperature upwelling.

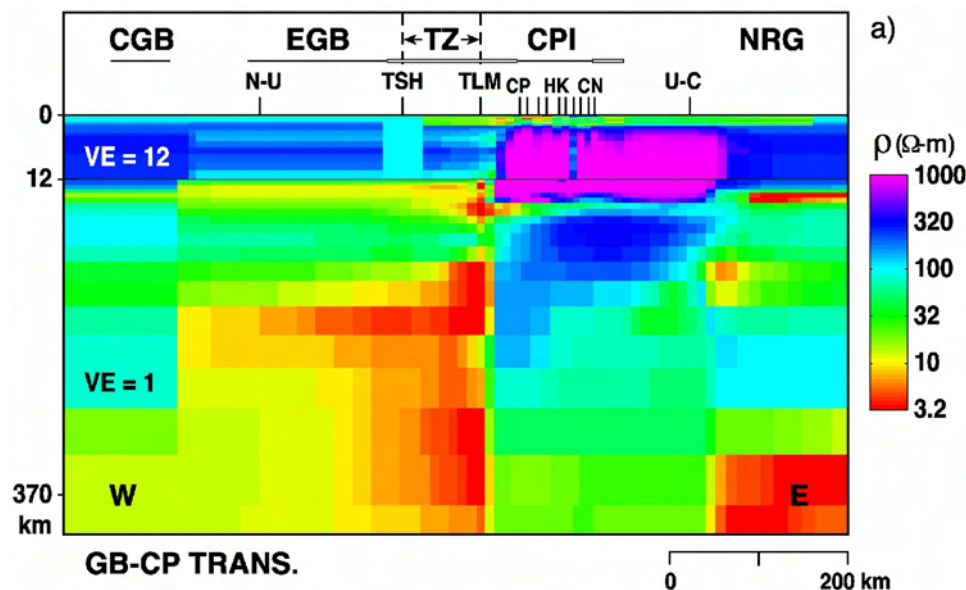


Figure 4. Inversion model of MT data to 10,000 s period in the EGB and the CP (Wannamaker et al., 2001). Projection of data profiles drawn overhead. Geographic features include Tushar Range front (TSH, S. extension of Wasatch Front), Thousand Lake Mountain (E. limit of TZ normal faulting), Capital Reef (CP), Hanksville (HK), and Canyonlands (CN). NRG is northern Rio Grande Rift. Open thin bars over TZ and eastern CPI are subject to upcoming MT fill-in. The starting average for EGB is shown in Figure 2.

Conclusions

Evidence is sparse, but taken at face value the properties of the eastern Great Basin confirm globally applied models relating mantle physical state to properties of mafic igneous underplate. A high-T upwelling is interpreted under the easternmost GB and TZ which may have generated a high- V_p rift pillow. We note that these melting models are based on laboratory experiments and have not been tested rigorously, largely because they have been applied to ancient rifted margins where any mantle anomalies responsible for producing margin magmatism are long gone. Further efforts should be made to obtain images of low velocity, attenuation, phase transition topography, and high conductivity in the upper mantle which will constrain mantle temperature and domains of melting. Studying a region like the GB-CP transition is advantageous, in that it is active allowing lower crustal structure and mantle state to be directly compared.

Experience in this setting begs the question of whether definition of a globally averaged temperature profile such as ACMA has fundamental tectonic significance. Does unequivocal identification of temperatures significantly higher than ACMA through geophysical or geochemical means require the conclusion that material has risen from mid-mantle or greater depths? However, identification of mantle temperatures will rarely be straightforward. Deriving mantle temperatures from either seismic velocity or electrical conductivity requires assumptions about mantle composition, in particular that dry peridotite or some other simple composition can be approximated. Presence of volatiles or alkali components in significant quantities promote melting and complicate velocity- or conductivity-temperature relations. The eastern Great Basin may be one of the few intracontinental regions where adequate constraints are at hand.

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An Alternative Model for the Origin of Non-Hotspot Intraplate Volcanism in the Pacific

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The formation of intraplate volcanic ridges on the south Pacific seafloor do not appear to be governed by conventional models of tectonic plate motion over a stationary hotspot. Although these volcanic seamount chains are oriented perpendicular to the plate spreading axis and parallel to the Pacific plate motion direction, geochemical evidence suggests that volcanism along the Pukapuka ridge propagates from west to east at a rate of 300 mm/yr, approximately 3 times faster than the plate velocity (Sandwell et al., 1995). Pronounced isotopic anomalies are found along the East Pacific Rise (EPR) (Mahoney et al., 1994) where the Pukapuka ridge is projected to have intersected the spreading center; anomalies with the same mixing trend that is found in Pukapuka (Janney et al., 2000), suggesting that there is material transport of anomalous mantle associated with volcanic ridge formation. We present seismic and petrologic results from the MELT and GLIMPSE marine geophysical expeditions indicating that enriched, and/or hotter mantle may underlie Pukapuka-type seamount chains. We are developing a new, alternative model in which intraplate volcanism is the surface expression of return flow to the EPR concentrated in low-viscosity channels or fingers in the asthenosphere, controlled by anomalous composition or temperature. Laboratory fluid experiments demonstrate that viscous fingering develops when a lower viscosity fluid is introduced into a higher viscosity fluid, consistent with classical Saffman-Taylor instabilities but for fluids where inertial forces are unimportant and viscous forces dominate.

Bathymetry, gravity, seismic, and magnetotelluric measurements from the MELT experiment demonstrate a pronounced asymmetry between the eastern and western sides of the EPR. Subsidence of the Nazca plate with increasing age is much more rapid than for the Pacific plate and seamounts are more abundant on the Pacific plate (Cochran, 1986, Scheirer et al., 1998). Body wave and Rayleigh wave tomography indicate lower seismic velocities beneath the Pacific plate (Toomey et al., 1998; Forsyth et al., 1998;). Surface wave and shear wave splitting studies indicate greater anisotropy on the Pacific side (Forsyth et al., 1998; Wolfe and Solomon, 1998). While ridge migration has been suggested as a mechanism for the observed asymmetry, geodynamic numerical models of mantle flow beneath a spreading ridge find that ridge migration alone has little effect on melt production or mantle temperature structure (Toomey et al., 2002; Conder et al., 2002). Anomalous temperatures or composition must be invoked in addition to the ridge migration.

P and S wave delays from the GLIMPSE experiment (Figure 1) show anomalous mantle directly beneath the volcanic ridges, and preclude the suggestion that volcanism is formed by lithospheric cracks that passively tap widely distributed, preexisting melt (Winterer, 2003). Further, the young Hotu Matua volcanic complex shows no indication that cracks or structural lineations precede volcanic activity. Normal faulting, which might support the lithospheric boudinage hypothesis (Winterer and Sandwell, 1987), is also not observed. Lastly, much attention has been given to convective rolls that are thought to originate as instabilities from cooling of the lower lithosphere in the presence of plate motion (e.g. Richter, 1973; Richter and Parsons, 1975). This hypothesis, however, must be reconciled with the observation of linear, low, free-air gravity

anomalies (Haxby and Weissel, 1986) that flank the volcanic ridges and the rapid propagation of the volcanism from west to east towards the EPR opposite plate motion direction. Evidence of age progressions within the south Pacific volcanic ridges, continuity of isotopic signatures between intraplate volcanism and the spreading axis, enriched, water-rich magmas, and asymmetric seafloor subsidence, volcanism, and seismic wave propagation surrounding the EPR point to transport of anomalous asthenospheric material from the west to east beneath the Pacific plate.

In order to model mantle flow within the asthenosphere, we perform laboratory fluid experiments that consider viscous fingering between two fluids in a thin horizontal layer. Displacement of a high viscosity fluid by a low viscosity fluid is found to proceed in the form of fingers or instabilities that develop due to a differential pressure gradient between the two fluids (Hill, 1951; Saffman and Taylor, 1958). The Reynolds number (Re), which describes the ratio of inertial forces to viscous forces, is much less than unity both in the oceanic upper mantle and in our laboratory experiments, assuring that inertial forces are negligible and viscous forces dominate. Preliminary results suggest that for viscosity ratio greater than about 10, the low viscosity fluid will propagate within a fluid-filled horizontal layer as fingering instabilities that grow radially from the point of introduction (Figure 2). We suggest viscous fingering of mantle material as a possible transport mechanism to explain the geophysical and geochemical observations described above and provide agreement with satellite gravity lineations presented in previous studies (Haxby and Weissel, 1986). Volatile rich or hot mantle material introduced into the colder, more depleted ambient asthenosphere in the superswell region of the southern Pacific plate (Phipps Morgan, et al., 1995; Gaboret, et al., 2003) may flow eastward towards the EPR more rapidly in low viscosity channels that develop due to Saffman-Taylor instabilities.

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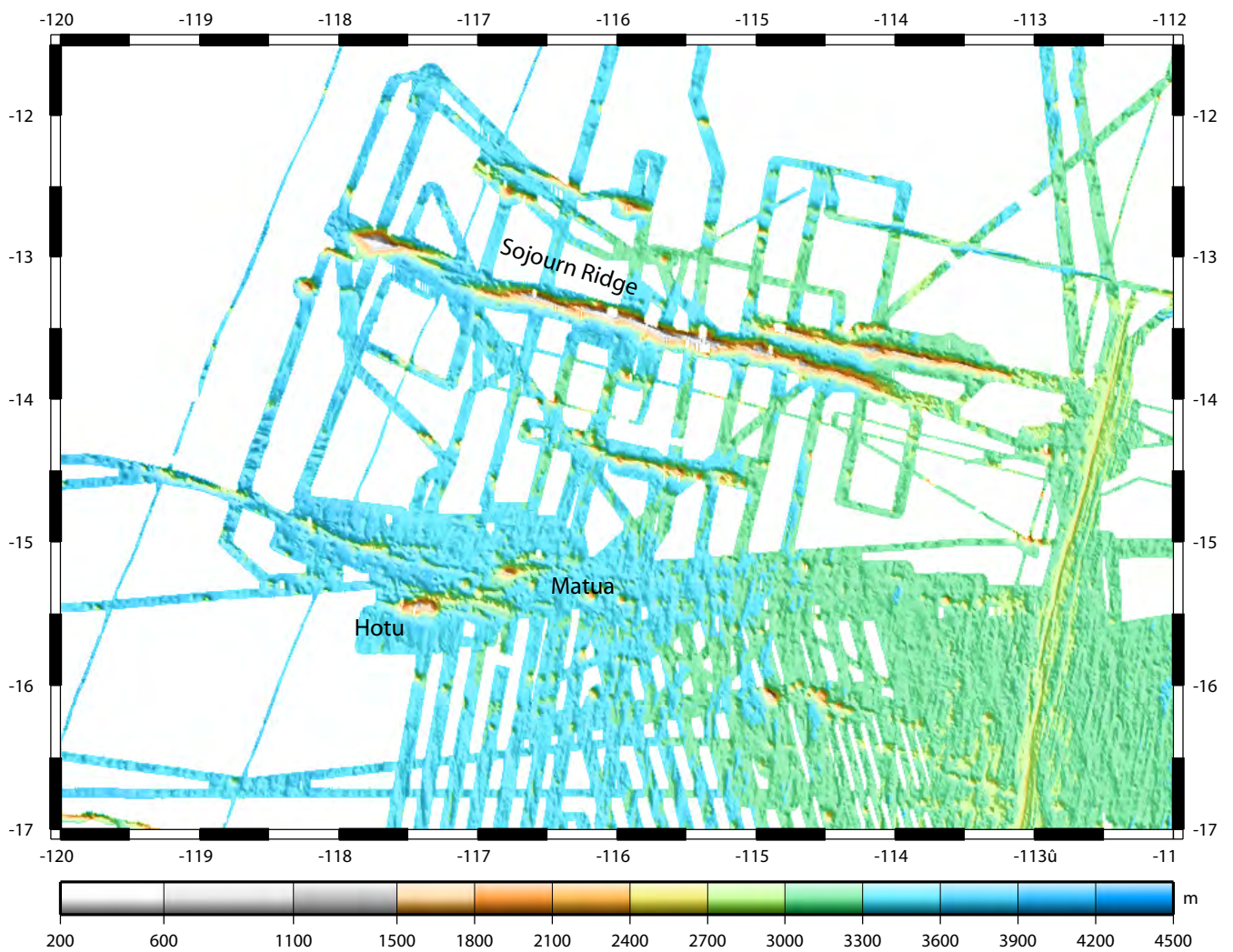
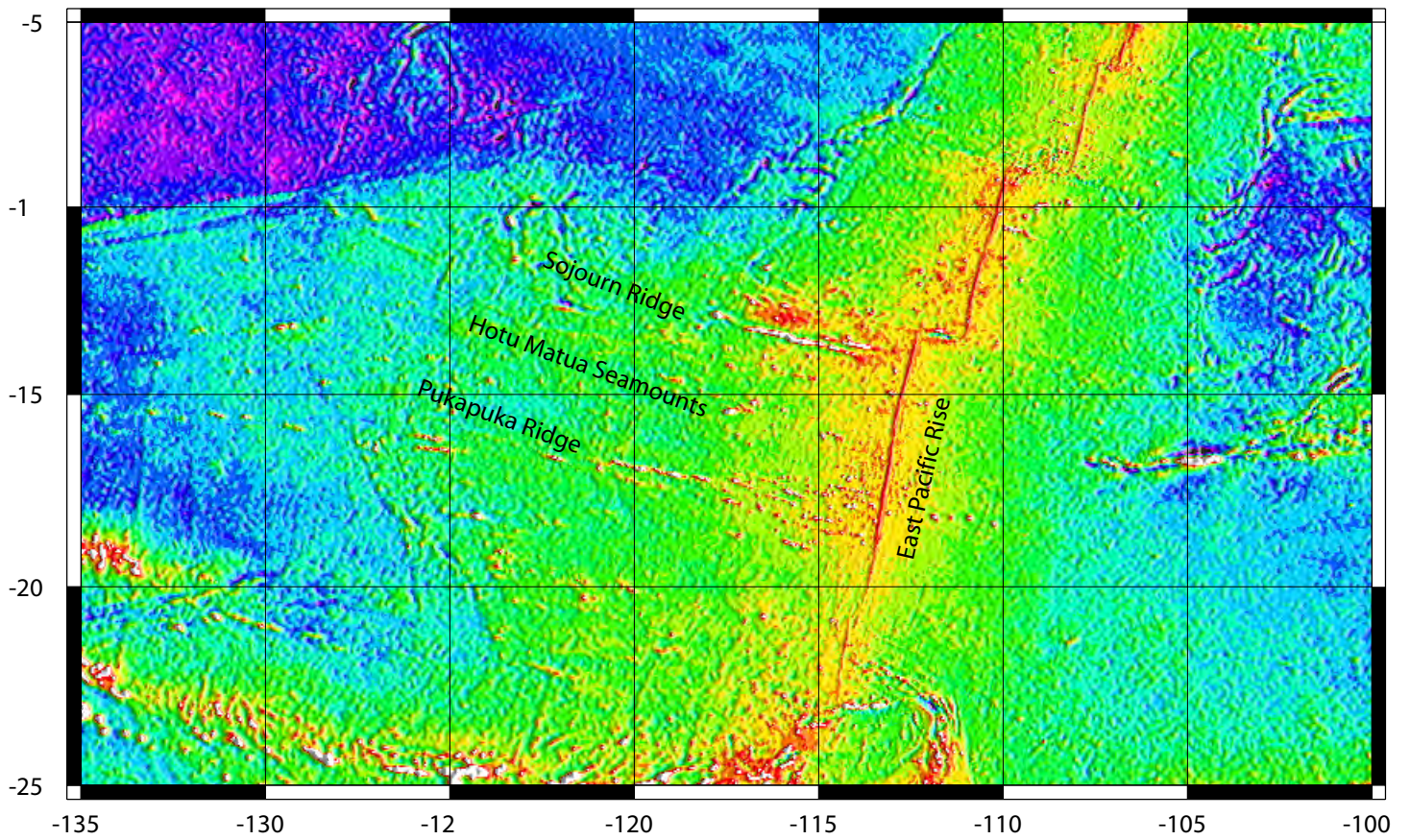


Figure 1a.) Compiled bathymetry for south Pacific and b.) local Hotu Matua area (after GLIMPSE experiment)

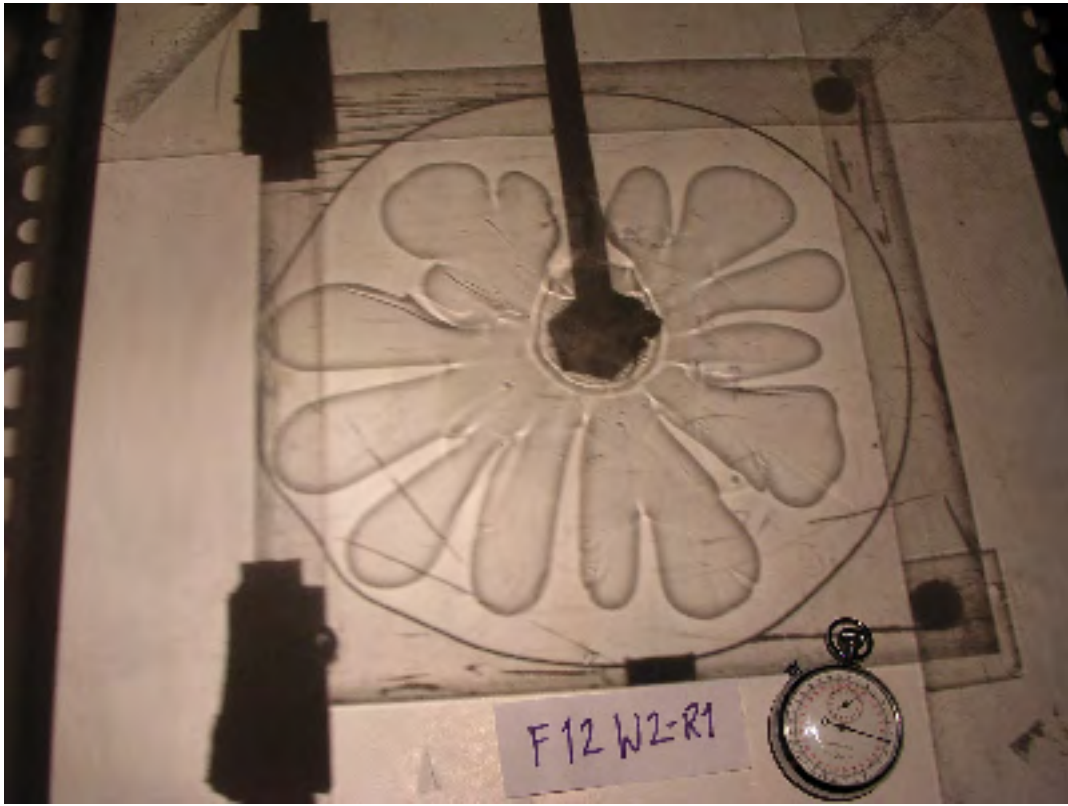


Figure 2. Viscous fingering in laboratory fluid experiments.

Sea-floor spreading and deformation processes in the South Atlantic Ocean: An evaluation of the role of mantle hotspots

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Recent improvements in our ability to enhance the resolution of satellite-derived marine gravity data (Geosat and ERS1) from 30-40 km down to ~10 km wavelength, has provided a unique opportunity to look in detail at the complex tectono-magmatic processes involved in the opening of the South Atlantic Ocean. This improvement in resolution has been achieved by innovative research and development funded by an oil industry consortium study conducted by the Leeds University spin-off company GETECH [1-3]. Application of new methodologies, developed as part of the above research, can significantly improve our understanding of oceanic crust-forming processes, particularly the role of near-ridge mantle hotspots and intra-plate deformation, hitherto impossible to resolve fully with current datasets.

Our current knowledge, based on existing research, indicates that slow-spreading mid-ocean ridges such as the Mid-Atlantic Ridge are strongly segmented along their axes by transform faults (Fig. 1). In the South Atlantic these transform faults are typically spaced some 50-100 km apart, reflect the relative plate motion directions of the newly formed crust and occur at offsets of the normal faulted median rift valley that marks the axis of the ridge. The sites of these active transforms are regions of decreased magma generation, resulting in the transform zone being starved of volcanism and expressed as a deep trough in the oceanic crust. At greater distance from the ridge crest, the transform motion ceases. This change occurs at the adjacent ridge offset; between the ridge offsets the transform separates different plates while beyond the ridge offsets the transform separates older (or younger) crust belonging to the same plate. Beyond this transition the transform fault is referred to as a fracture zone or flow line. The starved nature of the flow line and the differing age of the oceanic crust across the flow line generates a distinct bathymetric and gravity feature (Fig. 2) that can be traced for large distances away from the ridge axis and preserves evidence of former plate tectonic processes and movement vectors.

Studies by Silveira & Stutzmann [4] indicate that there are significant differences in S wave velocity structure beneath the northern and southern Mid-Atlantic Ridge (MAR). In the northern MAR negative S-wave anomalies occur down to 150 km depth, whereas in the South Atlantic these extend to 300 km depth. This difference can be related to the presence of hotspots along or close to the ridge axis in the South Atlantic, which locally generate trails of volcanic islands and seamounts as the plate migrates over the hotspot. These hotspot trails strike at an oblique angles to the flow-lines suggesting deeper processes are in operation than those that formed the flow-lines. These hotspots have traditionally been linked to upwelling convective instabilities or "mantle plumes", originating from thermal boundary layers at the base of the upper mantle (670 km discontinuity) or even the core-mantle boundary (2,900 km). The seamount trails may provide an important record of the local linear absolute velocity of the African plate in the hotspot reference frame. Additionally, establishing the processes which form these trails of South Atlantic hotspot volcanism can provide important insights into the role of mantle plumes in continental break-up and the subsequent evolution of the ocean basins.

Within the South Atlantic there are a number of distinct hotspot traces (see Figure 1) in the form of linear aseismic ridges (e.g. Rio Grande Rise, Walvis Ridge) and chains of seamounts and oceanic islands (e.g. St. Helena Seamounts). The two best developed hotspot trails have been explained by the upwelling of deep mantle plumes since the onset of continental break-up in the Early Cretaceous – the St Helena Seamount chain and the Walvis Ridge, linked to the St Helena and Tristan mantle plumes respectively [5]. The geometry of the St Helena seamount chain is, however, complicated at the continental end by the Tertiary-Quaternary volcanic activity of the Cameroon Volcanic Line (Fig. 1) which straddles the continental margin and appears to represent reactivation of this lineament. The Walvis Ridge extends from the continental margin of Africa to the active volcanic islands of Tristan da Cunha and Gough. The

SW end of the hotspot trail consists of a ~ 400 km wide region of scattered seamounts, small ridges and islands, similar in morphology to that of the St Helena Chain. The Rio Grande Rise is considered to be the conjugate of the Walvis Ridge on the South American plate, although its geometry is far less regular.

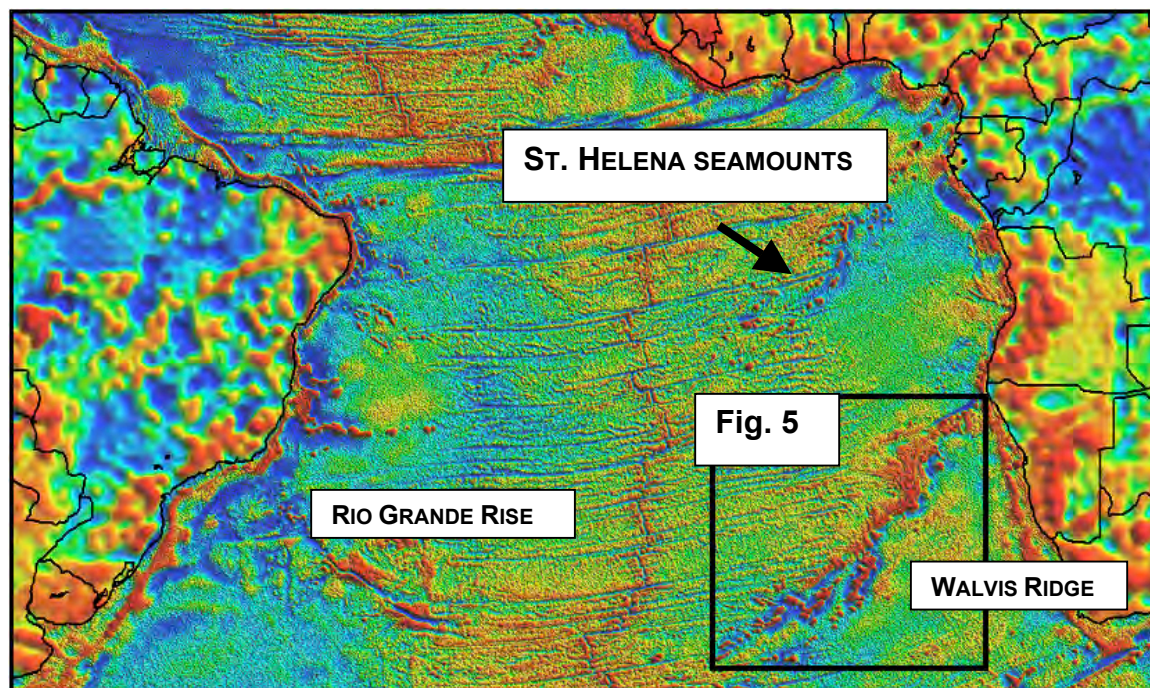


Figure 1: The satellite free air gravity of the Central, Equatorial and South Atlantic Oceans. Inset boxes indicate the location of Figures 6 and 7.

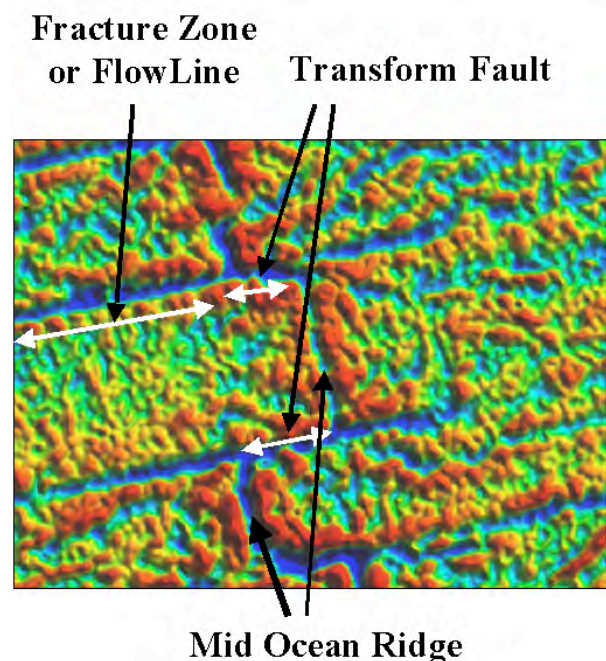


Figure 2: Gravity Image of the mid South Atlantic mid-ocean ridge

O'Connor et al. [6] have demonstrated (by Ar-Ar dating) that individual seamounts in these hotspot chains probably form very rapidly, in less than 1 Myr. These authors suggest that the migration rate of volcanism along the St Helena seamount chain has slowed significantly in the past 19 Myr to about 2 cm yr^{-1} which they relate to a 33% reduction in the absolute motion of the African plate. Such a slow-down could be

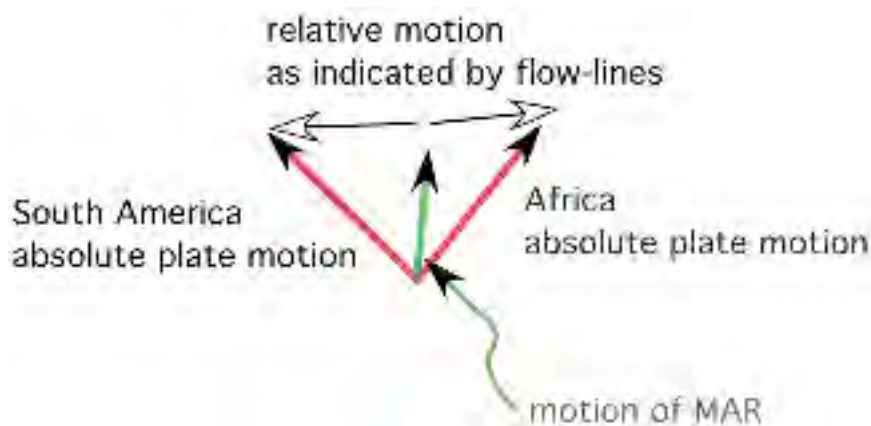


Figure 4: Vector diagram showing relations between absolute plate motions, relative motions and motion of the mid-ocean ridge (MAR)

The absolute plate motion results indicated on Figure 4 are based on the simple assumption that there is no net-rotation of the lithosphere (i.e. using a reference frame that yields zero for the integral of $\mathbf{v} \times \mathbf{r}$ over the Earth's surface, where \mathbf{v} is the plate velocity at position \mathbf{r}). The GFZ-Potsdam GPS data (red arrows in Fig.3) are considered to be the most reliable. For a given plate the motion should conform to rotation about a Euler pole. For South America (east of the Andes), as noted above, the plate motion is consistently to the NW with an Euler pole located at approximately 25.4 °S and 126.4 °W. For South America west of the Andes the motions to the NE are consistent with the subduction of the Nazca plate beneath the Andes.

KEY OBSERVATIONS

- That the absolute plate motion directions of Africa and South America have been consistent since the opening of the South Atlantic Ocean commencing in the Early Cretaceous (ca 126 Ma).
- That plate motion direction strongly influences fault tectonics (within both continents and oceans) on both micro- and macro-scales. Faults parallel to the direction of plate motion should be more susceptible to reactivation, whilst faults orthogonal to the direction of plate motion should remain locked. This may have important implications for the localisation of kimberlites and other centres of alkaline magmatic activity within Brazil and central-southern Africa.
- That the concept of “hotspot traces” is over simplistic and that a model involving the periodic release of plate stresses in the form of shear/wrench faulting is more appropriate.
- That the fabric of the ocean crust records relative (shallow phenomena related to flow lines) and absolute (deeper phenomena related to mantle hotspots) motion vectors that can help trace absolute plate motions back in time.
- That intra-plate deformation has an important role during sea-floor spreading. Deformation appears to be quite common within the South Atlantic oceanic crust and may be the controlling factor in the development of the Walvis Ridge and Rio Grande Rise aseismic ridges. These ridges appear to have developed as a response to periodic changes and release of intra-oceanic plate stress. This has resulted in wrench and shear movement and deformation along these features (Fig.5).
- The aseismic ridges appear to have had a controlling influence on periodic stress release within the oceanic parts of the plate as well as on the adjacent continents.
- Stress release within Africa and South America may have resulted in localised decompression melting of the mantle, evidenced by the presence of alkaline volcanic complexes and kimberlite diatremes that are younger than the main phase of ca 135 Ma flood basalt volcanism (Parana-Etendeka provinces) that pre-dates continental break-up.
- If the internal structure of the Walvis and Rio Grande ridges can be better resolved, we may gain fundamental new insights into the origin of these intra-oceanic plate structures and adjacent continental weak zones, leading to the development of generic models for other regions of the Earth.

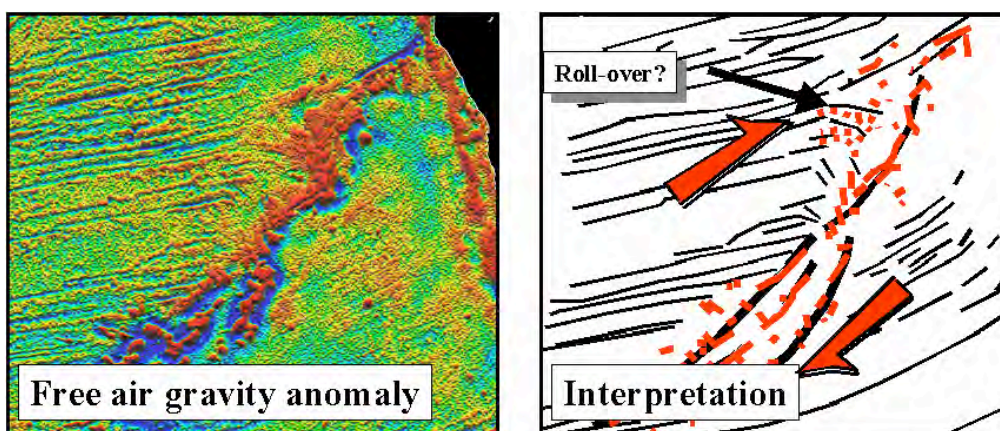


Figure 5. Gravity field of the Walvis Ridge and possible interpretation in terms of stress release at discrete times resulting in shear movement and deformation

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The geodynamic setting of Tertiary-Quaternary intra-plate magmatism in Europe: The role of asthenospheric diapirs or mantle “hot fingers”

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Despite significant advances in our understanding of the nature of mantle convection, we still have few constraints on the geometry of the thermal (and chemical anomalies) widely referred to as mantle plumes. Numerical and analogue modelling has indicated that several different scale lengths of convective instability are possible, with upwellings originating from boundary layers in the Earth's mantle such as the 650 km discontinuity or the core-mantle boundary. A variety of “evidence” has been used to argue for the existence of lower mantle plumes including teleseismic tomography, Sr-Nd-Pb-He isotope studies of oceanic island basalts and the existence of flood basalt provinces in the geological record, much of which is indicative but by no means conclusive. Far stronger evidence exists for the existence of plume-like convective instabilities in the upper mantle in geological settings in which magmatism is long-lived and plate motions relatively slow. The fundamental issue is whether these short-wavelength instabilities are a distinct mode of upper mantle convection, driven from below (perhaps by base heating of the 650 km discontinuity), or whether their locations are constrained by pre-existing lithospheric heterogeneities and the regional stress field.

Palaeocene-Recent volcanism within western and central Europe (Fig.1) is spatially and temporally linked to the development of a major intra-continental rift system and to domal uplift of Variscan basement massifs on a scale of 100s of km [1,2]. Mantle melting has been attributed to the diapiric upwelling of small-scale, finger-like, convective instabilities from the Transition Zone at the base of the upper mantle (Fig. 2), identified on the basis of local seismic tomography experiments in the Massif Central (France) and the Eifel (Germany) [3,4,5]. The local seismic tomographic studies indicate the existence of localised zones of mantle upwelling from depths of 400-650 km, 100-300 km across and up to 100-200 degrees Centigrade hotter than ambient mantle (if the velocity anomaly is attributed only to temperature). Global seismic tomographic studies also reveal the existence of broad zones of low velocity material in the upper mantle and, additionally, suggest the existence of a zone of low seismic velocities at depths of 900 to 1400 km in the lower mantle, extending from Iceland to the Eifel volcanic province of northern Germany, the Massif Central of France, the Hoggar massif in northern Africa and the Canary Islands [6,7]. We must therefore consider the possibility that the diapiric upper mantle upwellings inferred to have triggered the Tertiary-Quaternary volcanic activity within Europe could be linked dynamically to the upwelling and lateral spreading of the Iceland mantle plume. However, although Icelandic basalts are characterised by He isotope signatures ($R/R_A > 20$) consistent with a lower mantle source for the plume, detailed seismic tomography studies indicate that the Iceland plume itself may be an upper mantle phenomenon [8].

There is an interesting anti-correlation between the locations of the major volcanic fields, and the mantle diapirs which underlie them, and the location of a zone of seismically fast material in the base of the upper mantle (Fig.3) which has been interpreted as a zone of subducted oceanic lithosphere slabs [9]. The dynamics of mantle convection beneath Europe thus appears to be intimately linked to the closure of the Tethys ocean and the formation of the Alpine orogenic belt.

The spectrum of primitive mafic magma compositions within the European volcanic province ranges from melilitite nephelinites and melilitites, through basanites and alkali basalts to subalkaline tholeiites; these are considered to be the products of variable degrees of partial melting of a relatively homogeneous HIMU-like reservoir within the upper mantle, the European Asthenospheric Reservoir or EAR [10]. Variations in the trace element and Sr-Nd-Pb isotopic characteristics of the magmas are consistent with mixing of partial melts from both lithospheric and asthenospheric mantle sources. A component geochemically

similar to the EAR also exists within the Icelandic plume system; this is preferentially sampled by relatively rare, small degree, partial melts (nephelinites and alkali basalts). Thus both geophysical and geochemical data could be used to support a geodynamic link between the Palaeocene-Recent activity of the Icelandic mantle plume system and the magmatism much further to the south in western and central Europe.

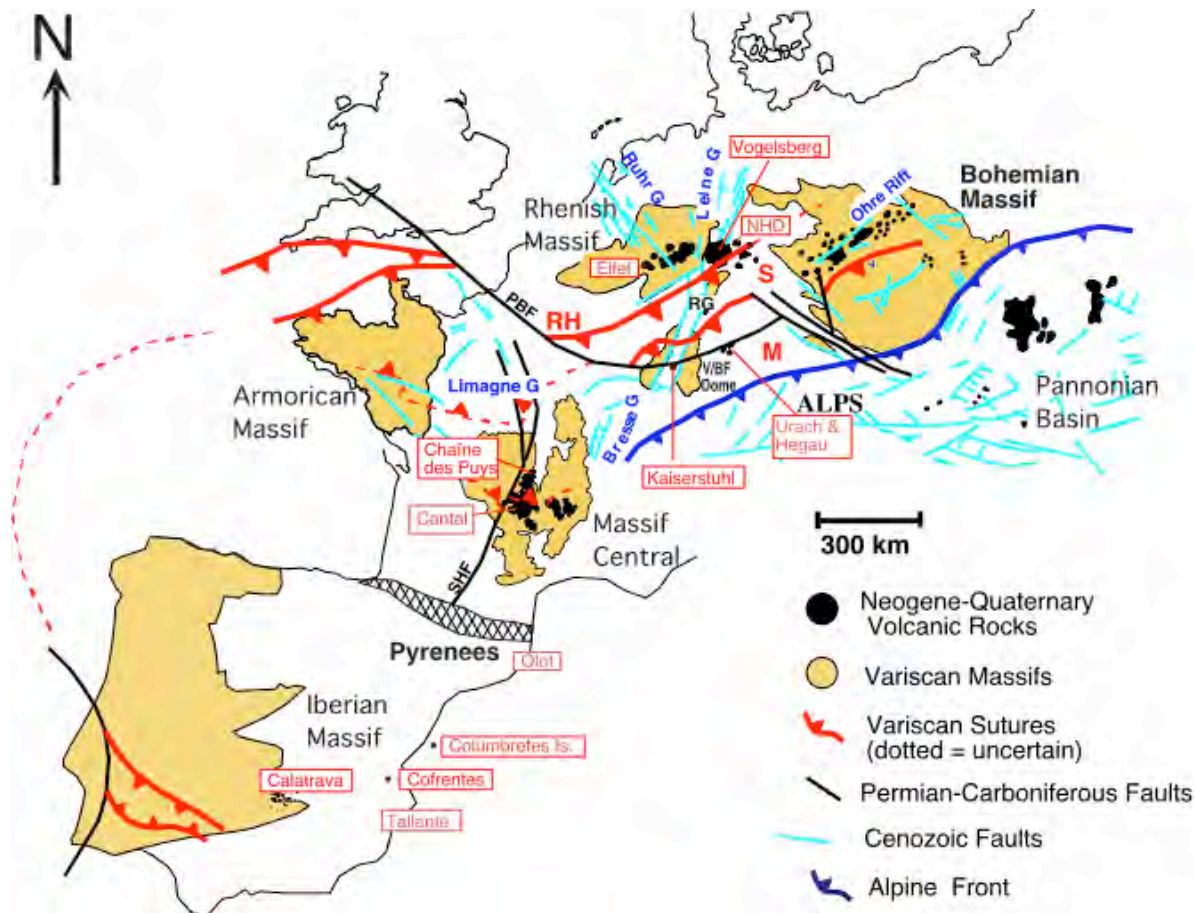


Fig.1 Location of Tertiary-Quaternary volcanic fields in Europe After: Wilson & Patterson [2]
Variscan terranes: RH – Rhenohercynian; S- Saxothuringian; M- Moldanubian
Abbreviations: RG- Rhinegraben; SHF – Sillon Houiller fault; PBF – Pas de Brays fault

The major Tertiary-Quaternary magmatic provinces within western and central Europe are located along tectonic sutures between micro-plates which collided during the Variscan orogeny some 300 Myr ago [1]. There is, however, little correlation between the zones of maximum lithospheric extension (e.g. the Rhinegraben rift system), which transect the Variscan lithospheric terrane boundaries at a high angle, and the location of the main volcanic fields (Fig.1).

Within the Massif Central of France the largest volcanic edifice, Cantal, is located at the intersection of three distinct lithospheric terrane blocks, Fig.4) [11]. This raises the question of whether the mantle diapirs beneath the Eifel and Massif Central volcanic fields really are upwellings from the Transition Zone or if they are tectonically induced zones of decompression melting which have propagated downwards from the base of the lithosphere.

Tertiary-Quaternary volcanism in Europe HOT FINGERS or Mini-PLUMES ?

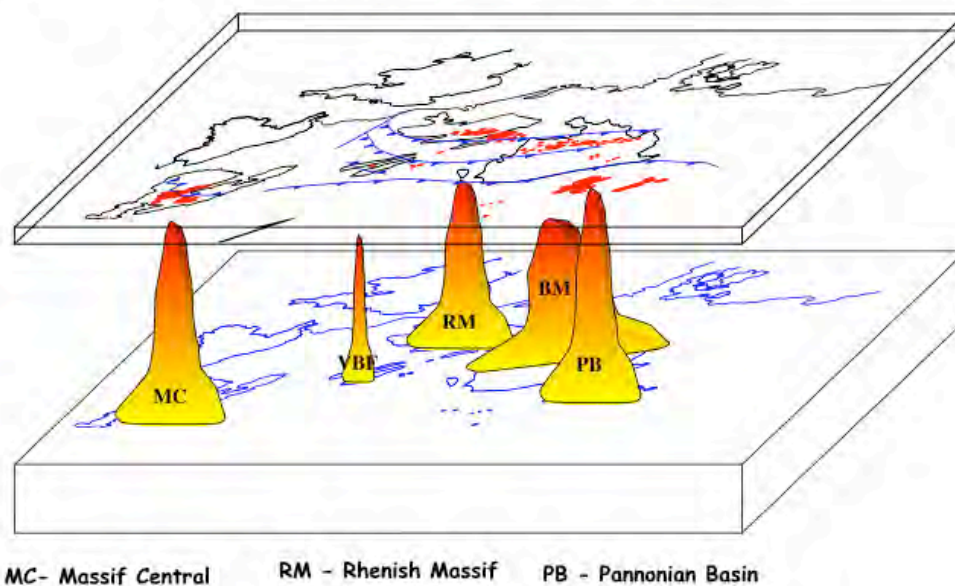
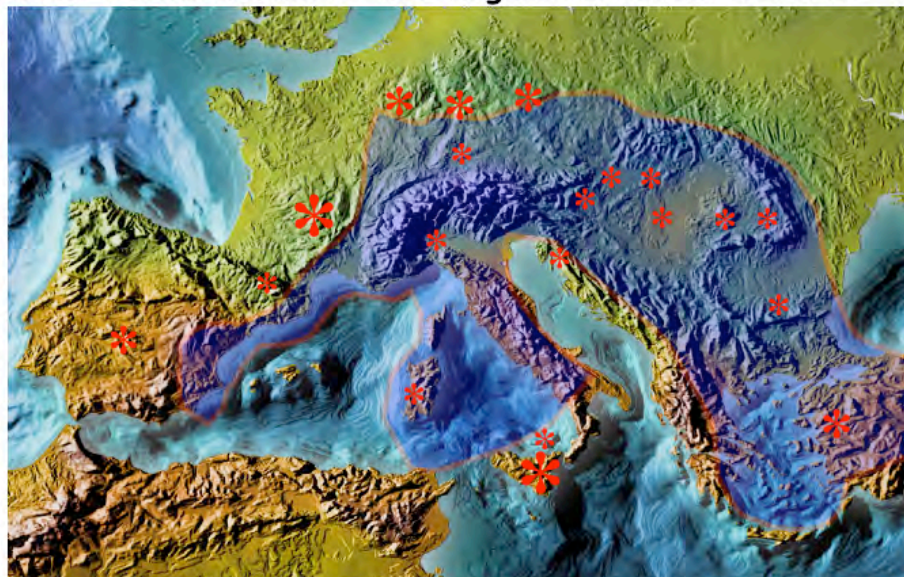


Fig. 2: The location of Tertiary-Quaternary volcanic fields in Europe and diapiric instabilities in the upper mantle. After Granet et al. [3]. (BM –Bohemian Massif; VBF- Vosges-Black Forest Dome)

Distribution of volcanic fields

Links between mantle "hot fingers" & subducted slabs



* Volcanic fields

Subducted slabs at 500-600 km
Piromallo et al. (2001)

Fig.3 Location of Tertiary-Quaternary volcanic fields in relation to a zone of seismically fast material at the base of the upper mantle interpreted by Piromallo et al. [11] to be subducted oceanic lithosphere.

Schematic model for magma generation

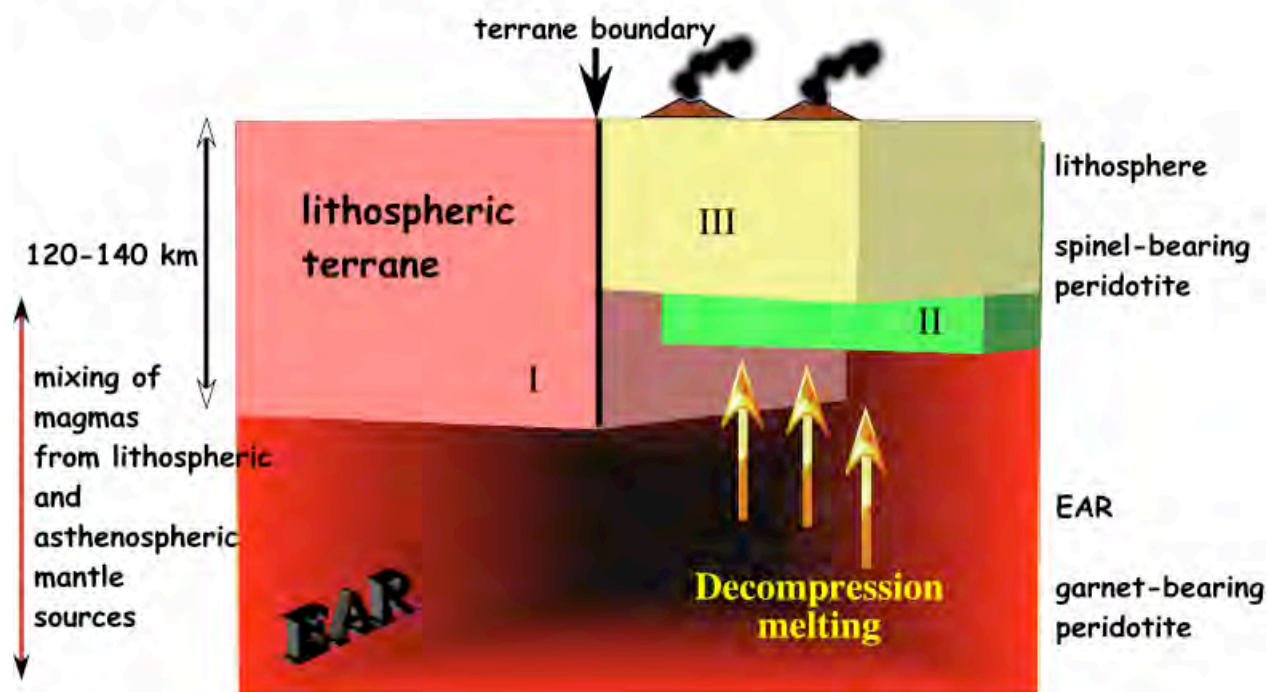


Fig. 4 Nucleation of diapiric upwelling at the intersection of lithospheric terrane boundaries in the French Massif Central. After Babuska et al. [11]

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**SEAMOUNT CHAINS RESULT FROM EPISODIC CHANGES IN IN-PLATE
STRESS THAT OPEN CRACKS THROUGH THE LITHOSPHERE AND
PERMIT MAGMA ASCENT : THEY DO NOT REQUIRE PLUMES**

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Newer data support, as an alternative to the plume/hotspot hypothesis for seamount chains, formation via in-plate, stress-generated cracks through the lithosphere, first discussed by Jackson et al. (1972, 1975). Many Pacific mid- to late Cenozoic seamount chains comprise intermittently spaced volcanic ridges, in some places aligned en echelon to the overall trend of the chain, a pattern that reflects tensional and shear stresses in the lithospheric plate. The overall trend is a line of incipient cracking, generally close to the average direction of plate motion in the fixed-Antarctica reference frame. These chains are not copolar nor do they have the same angular rate of progression. By contrast, Early Cenozoic and Cretaceous chains are not demonstrably aligned in the directions of plate motion nor were the volcanoes in a chain formed at the same paleolatitude. The ridges mark episodic deviations in the stress field, permitting cracks to open through to the asthenosphere. The upper parts of the asthenosphere are at the solidus temperature, as manifested in seismic velocities indicating the presence of small fractions of melt. Cracks allow magmas to ascend to the surface where they erupt to form the volcanoes and ridges. Fertility variations in the asthenosphere favor large volumes of magma along cracks suitably situated. Massive volcanoes commonly conceal their narrow-ridge origins. Variable melt volumes from the heterogenous "plum-pudding" upper mantle provide the full range of compositions represented in seamount rocks. Heat-flow and lava-composition data do not support the notion of hotspots: temperatures are not greater than normal. Cracks typically break through in younger parts of the lithosphere, which are thinner and weaker than older lithosphere, but cracking is possible anywhere the lithosphere is thin or weak or the tensional stresses adequate. Cracking is common, for example, along the thinned lithosphere of the boudinage-like structures imaged on regional gravity maps, as in the Pukapuka chain, which follows one of the regional gravity lows. Here, as in several other chains, the time sequence of volcanic ridges is not progressive. Bouyant magma can actually magma-fracture the overlying plate even if it, in the absence of melt, could retain its integrity. The crack mechanism requires no excessive "hotspot" temperatures and no plumes.

Geology, structure, and source of the Kikkertavak Anorthosite, northern Labrador, Canada

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The origin and petrogenesis of anorthosites are poorly known. Several issues dominate the “Anorthosite Problem”. These include issues related to the source and fractionation of magmas, the emplacement and crystallization of these magmas, and the overall tectonic regime in which these processes take place. Current research of the Labrador Research Group of the Department of Earth Sciences, Memorial University of Newfoundland is addressing these issues through a study of the regional geology of the Nain Plutonic Suite of Labrador.

The Nain Plutonic Suite is considered the type example of a Proterozoic Anorthosite-Mangerite-Charnokite-Granite (AMCG) suite. Proterozoic AMCG suites occur worldwide; they have generally been interpreted as anorogenic magmatic products, and attributed to a variety of processes, especially plumes. Current work in the NPS has revealed that pluton emplacement was associated with shear zones and dyke swarms. These features indicate that the plutons were intruded into an extensional/transpressional environment and do not reflect diapirism above a mantle plume.

As a contribution to this research project I have mapped most of a large (20 x 60 km.) pluton of anorthosite called the Kikkertavak Anorthosite. This body is one of the youngest components of the c. 1360-1300 Ma Nain Plutonic Suite. Most of the older anorthosite plutons are partly deformed and recrystallized, whereas the Kikkertavak Anorthosite contains largely pristine mineralogy and structures. The pluton is a rectangular, tabular body and was intruded into Archean quartzo-feldspathic gneisses and deformed and recrystallized Proterozoic anorthosites and related rocks. The Kikkertavak Anorthosite has in turn been intruded by monzonitic and granitic sheets, and has been dissected by north-south and east-west trending mafic dykes and shear zones.

The feldspar crystals that comprise most of the Kikkertavak Anorthosite come in a range of sizes and textures that can be directly correlated with the crystallization history of this pluton. The pluton also contains relatively small amounts of olivine, orthopyroxene, and magnetite/ilmenite. Contact products with host rocks include heterogeneously layered troctolites, norites and anorthosites along the walls of the pluton and more massive troctolites and norites in the roof.

Several textural and mineral facies are observed within the Kikkertavak Anorthosite, and the relationships within and between these facies preserve a physical record of the history of this pluton. The cumulate texture, mineralogy, and differences in texture between components of the Kikkertavak Anorthosite suggest that it is the product of more than one melt and has had a protracted history of crystallization and fractionation. A detailed study of individual feldspar crystals is underway that will place

further constraints on the source and crystallization history of these rocks. The feldspar crystals from the different phases of the Kikkertavak Anorthosite have the potential to preserve trace element and isotopic signatures spanning the crystallization history and prehistory of the pluton. The evolution of these values within individual feldspar crystals will provide key information on the source, fractionation, and possible contamination involved in the production of anorthositic magmas.

Lithospheric control on silicic magma generation associated with the Ethiopian flood basalt province

Gezahegn Yirgu, Dereje Ayalew and Cindy Ebinger

Abstract

Most of the Ethiopian flood basalts erupted 30 Ma, during a 1 Ma period, to form a vast volcanic plateau. Inter-layered with the flood basalts, particularly at upper stratigraphic levels, are sequences of felsic lavas and pyroclastic rocks of rhyolitic or less commonly trachytic compositions. Immediately after this peak of activity, shield volcanoes formed on the flood volcanics, after which the volcanism was largely confined to regions of rifting. Geophysical and geochemical data in the region support the presence of one (or more) plume(s) during the formation of the volcanic province.

The Ethiopia-Yemen traps overlie Mesozoic-Paleocene shallow marine sequences, indicating that the traps erupted onto lithosphere stretched to a minor degree during the breakup of Gondwana, which had only partially thermally reequilibrated. Thus, the location of the flood basalt province may have been influenced by pre-existing lithospheric structure, but the degree of pre-plume thinning would have been too small to explain the silicic volcanic rocks as products of decompression/adiabatic melting of underplated basic igneous rocks. Instead, the silicic volcanics are consistent with their derivation by processes of fractional crystallization of mantle derived basaltic melts combined with assimilation.

On the other hand, there is abundant evidence that the Mesozoic Karoo-Ferrar, Parana-Etendeka and Deccan flood volcanics were generated beneath thinned lithosphere in response to regional extension. In these regions, sea floor spreading occurred during or shortly after trap emplacement, except in the older Karoo-Ferrar province where full-fledged ocean floor spreading initiated some 13 M.y. following peak flood volcanism. The origin of silicic volcanic rocks in Etendeka-Parana has been attributed to the process of partial melting of underplated basic igneous rocks or continental crust.

Overall, data and observations suggest a broad relationship between lithospheric thickness and the process of felsic melt generation associated with LIPs. In those areas where lithospheric thinning has taken place, extension would have triggered decompression melting of still-hot underplated basalts at the base of previously thinned/stretched continental crust, causing widespread silicic volcanism.

In contrast, flood-basalt magmatism under normal or un-stretched lithosphere creates favorable conditions for the development of crustal magma chambers where basaltic magmas pond and evolve toward silicic derivatives by a combination of fractionation processes and assimilation of surrounding basement and/or roof rocks.

Petrological and geochemical investigations can broadly discriminate mantle-derived fractionated silicic volcanics from those originating by partial melting of underplated basalts and crust. This provides a possible tool to infer lithospheric structure at the time of felsic magmatism associated with LIPs.