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ICELAND PLUME TECTONICS, SOME SPECULATIONS AND FACTS

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Abstract

Geoscientific observations in and around Iceland can not be explained by a simple model of ridge spreading and associated transform motion. A comprehensive model must include a mantle plume beneath the center of Iceland, i.e. a hot spot. A plume tectonic model is supported by observations from seismology as well as from many other fields of geology and geophysics. Among significant features of the Iceland plume tectonics is the magnificent gravity anomaly observed by satellite sea surface measurements (Figure 1). Other significant features of the Iceland plume tectonics is the existence of strain waves and the dominance of stable motion and other subcrustal aseismic processes on the crustal deformation and failures. This has consequence for earthquake prediction. The fast processes above the Iceland hotspot favour the discovery and monitoring of such processes.

Interplay between pull and push

The Iceland hotspot postulated by (Vogt, 1974) and (Morgan, 1981) was seismologically confirmed as an uppwelling plume by 3–D inversion of teleseismic P wave residuals by (Tryggvason et al., 1983). At 275–375 km depth, i.e. in the deepest layer modelled, a 100 km wide chimney–like structure of low velocity (4%) was resolved (Figure 1). At 100–200 km depth low velocity material stretches from the plume center especially in directions between S and W. At 0–75 km, low velocity anomalies coincide with the active rift zone in northern Iceland, but stretch also towards SW. Assuming that the velocity variations result from temperature variations only and assuming a viscosity of 10^{19} Pas, it was estimated that the upwelling rate is 1.7 m/year in the narrow plume chimney,14 km^3/year, i.e. much more production than the 0.1 km^3/year expected sum of erupted material and dyke intrusions in a 10 km thick crust in Iceland, assuming the Nuvel–1 spreading rate of 1.94 cm/year (DeMets et al., 1990). So excess upwelling material must flow to the side. There is further morphological and geochemical evidence (Vogt, 1974; Magde and Smith, 1994; Appelgate and Shor, 1994) for excess material flow outward from the center of Iceland, especially to SW. Satellite gravity data show effects of intrusive material from the plume to a distance of 700 km to the SW from the Reykjanes Peninsula as a V shaped anomaly narrowing in that direction (Figure 1), where the bottom changes from an even axial–high structure to a median valley farther south. Studies of helium isotopes (Poreda et al., 1992) indicate a center of upwelling mantle material 30–50 km SW of the seismic plume center. The reason for this difference may be that the plume at depth is "moving" under cold east Iceland crust just starting to heat it up. Additional evidence for upwelling plume in the mantle beneath Iceland comes from interpretation of gravity measurements (Kaban et al., 1994; Þórarinsson, 1994) (Figure 1), which locate it above the deep–seated seismic anomaly. The excess elevation that they observe above the plume has to cause a flow to the side. Thus the tectonics of Iceland is governed by the interaction between two processes,

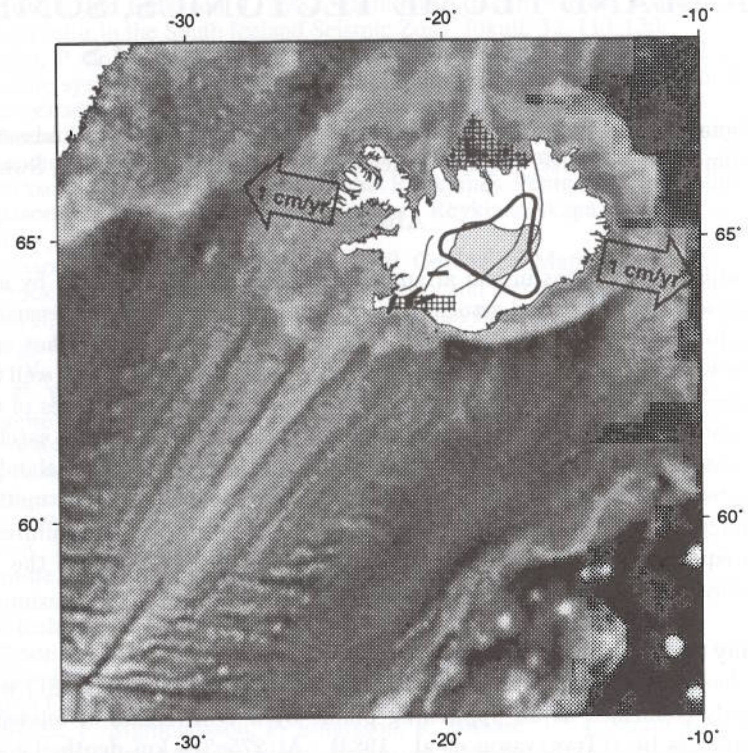

Figure 1: *Iceland and its relation to the mid-Atlantic Ridge. The eastward offset of the ridge is seen and outlines of a hot mantle plume at depth (thick contour in C-Iceland). The outlines of 60 mgal Bouguer gravity low compared to Iceland as a whole as inferred by Þórarinsson is shown in grey and the outlines of V shaped gravity anomaly on the Reykjanes Ridge as inferred from satellite measurements (SIO and NOAA,world_grav.img.7.2). The general large scale divergency of the plates (1.9 cm/year) is indicated as well as the rift zones (thin lines) and the interconnecting transform zones (hatched areas). Compressional stress directions from stress inversion of local earthquakes are shown in thick line segments (R. Slunga, pers. comm.).*

i.e. 1) Drift of the two large-scale blocks from the mid-Atlantic Ridge; 2) Push from the center of the hotspot (Stefánsson and Halldórsson, 1988). Apparent changes with time in the direction of regional stresses within Iceland may be attributed to an interaction between these two processes. Evidence for this also comes from study of paleostresses (Bergerat et al., 1990) indicating that local forces in the mantle play a significant role in the plate divergency. In situ stress measurements (Schäfer and Keil, 1979; Haimson and Rummel, 1982) point towards compression outwards from the plume center and rift zones in Iceland, and that the horizontal compressions are stronger than the "tensions". Both these results are supportive of push as being more prevailing than WNW–ESE pull. Volumetric strainmeters in SW Iceland indicate compression of 400 Nstrains/year (Ágústsson, 1994). Comparison of the strain energy of 2 cm/year divergency and the release of historical earthquakes (Stefánsson and Halldórsson, 1988) shows total agreeement. However

the authors assumed 5 km thicker crust than later evidence indicates and as they pointed out a considerable part of the motion must be taken up aseismically, i.e. the total earthquake moment is larger than expected from 2 cm/year plate motion.

Strain waves and the dominance of stable flow at depth

Many examples of temporal association of historical seismicity over large areas and also volcanic eruptions, have led to the concept of strain waves (Stefánsson et al., 1993). The time lag between such events is commonly of the order of weeks to years. The distance between two associated events is generally so large compared to the size of the events that it is not probable that one causes or triggers the other. In such cases it has been suggested that both or all the events are triggered by a common large source, slowly propagating (Stefánsson et al., 1993). Besides the historical indications for strain waves, repeated GPS measurements in Iceland (Foulger et al., 1992; Foulger et al., 1994; Sigmundsson et al., 1995) made with time intervals of 2–3 years show great variance from one period of time to another. This also suggests an episodic character of the plate motions in Iceland, possibly caused by frequent intermittent intrusive activity at depth.

Contraction on a few borehole strainmeters in southern Iceland during the first five months of 1990 (Ágústsson, 1994) may be a direct observation of a strain wave, which was triggering a row of events during 1990–1991 in SW Iceland and on the Reykjanes Ridge.

The short distance between parallel faults of historical earthquakes and the fact that recently active faults frequently cross each other (Einarsson et al., 1980; Einarsson, 1991) has also been explained by the dominance of stable flow at depth over the crustal processes (Stefánsson and Halldórsson, 1988). Extended swarm activity where movement occurs repeatedly on the same fault many times during a short time interval, has been observed (Rögnvaldsson et al., 1994), points in the same direction.

The deep structure of Iceland

Seismic refraction/reflection results (Pálmason, 1971; Flóvenz and Gunnarsson, 1991; Bjarnason et al., 1993) indicate seismic velocites in SW Iceland as shown schematically in (Figure 2). In the following the layer between about 10 km depth i.e. down from the lower crust to 20–30 km depth will be called the intermediate crust–mantle layer. It has too low body-wave velocity for what is expected in the mantle. However the chemical composition is typical for the uppermost mantle (Guðmundsson, 1994).

Magnetotelluric observations (Beblo and Björnsson, 1980; Hersir et al., 1984; Eysteinsson and Hermance, 1985) indicate low resistivity (10–20 Ohms) at depths between 10–20 km and even shallower in the volcanic zones, i.e. in the intermediate layer.

Earthquakes stop to large extent at 5–12 km depth limit in the South Iceland seismic zone (Stefánsson et al., 1993), implying ductile response to slow strain loading below these depths. Based on heat gradient measurements and the thermal model of (Pálmason, 1973) the depth of the earthquake ductility is at the 750°C isotherm (Figure 2). This is in agreement with brittle–ductile transition in oceanic areas and conistent with dry-olivine rheology as measured in laboratory experiments (Wiens and Stein, 1983). On the other hand there is a good transmission

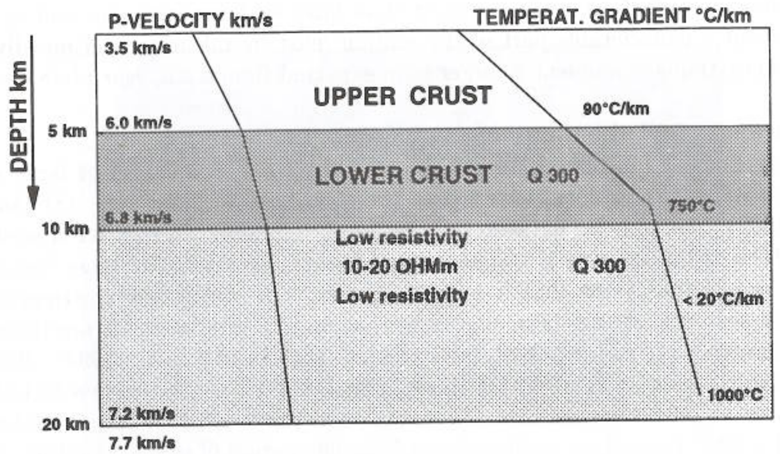

Figure 2: *The structure of Iceland is shown schematically down to 20 km depth. The depth to the brittle/ductile boundary, i.e. to the 750° C isotherm which is at 8 km at this particular place is shown. The low resistivity inferred from MT measurements is shown.*

for short period S waves down to 20 km depth, Q_s=300 (Menke and Levin, 1994). Thus the rocks in the intermediate layer respond elastically rather than ductilely to high strainrates as are in high frequency elastic waves. It is assumed that the material in the intermediate layer is olivine rich peridotite. To comply with the elastic/ductile response to different strain rates and laboratory experiments with olivines (Kirby, 1983) we conclude that the temperature increase from the brittle/ductile transition down to 20–30 km is at most 200°C. Following structure of the intermediate layer can explain the above observations: A crystal matrix of olivine rich peridotite close to dunite. Water components, carbondioxide are expected to exist as intergranular fluids at temperatures exceeding 750°C in these depth ranges. The bottom of the intermediate, i.e. the seismic refractor at 20–24 km depth in SW Iceland, is probably due to pressure/temperature controlled change in rheological properties. The fluids in the intermediate layer are extracted from the mantle below the refractor. They lower the resistivity and the heat gradient. They carry heat energy and pressure from depth, to be trapped by the brittle crust.

The intermediate layer is the locus of large aseismic motion and flow avalances much faster than the average velocity of the plate motion. This aseismic motion is linked to episodic deformation in the brittle crust above and with intrusions and other results of gravitational differentiation below. Here may be the source of strain waves, which may also have their origin in pulsations deeper in the mantle plume. A very interesting aspect is viscosity lowering below the seismic zones and the rift zones related to the fast relative plate motions there. Velocity of migration of seismic activity in the South Iceland seismic zone (SISZ) seems to be of the order of 5–10 km/day. This would indicate effective viscosity below the brittle crust of the order of 10^{17} Pas, much lower than inferred for Iceland as a whole from glacial rebound, which is 10^{19} Pas. From tilt measurements in a lake in the eastern rift zone effective viscosity has been found an order of a magnitude lower than the usual value (Sigmundsson et al., 1995).

Conclusions

Most of the upwelling flow of the Iceland plume, 14 km^3, takes place in a 100 km wide chimney at depth. It flows to the sides at around 100–200 km depth, most pronounced towards the Reykjanes Ridge. This outward flow is compensated by extraction of light weight vertically rising basaltic fluids. This fits well to the petrological composition of most of the Icelandic basalts which are extracted from the mantle at a depth of about 120 km above the plume and 80 km in the Reykjanes Peninsula (McKenzie and O'Nions, 1991). The basalt fluids rise to the surface from the flattened plumehead, above the center of the plume, and also in the active rift zones, even out to 700 km distance on the Reykjanes Ridge. These fluids compose the intermediate layer at 20–30 km depth which together with a push from the elevated central Iceland create the special features of the Iceland plume tectonics. A part of the outward flow takes place in the intermediate layer (20–30 km depth) and the brittle crust to compensate for the excess elevation of central Iceland.

The brittle crust can store shear strain energy for time period of the order of hundred years causing stick–slip motion and large earthquakes. The intermediate layer can store shear strain energy only for a short period of time. The different process velocities for the flow below and for the stick slip motion in the crust lead to local and temporal changes in stress intensity and stress directions in the brittle crust. The material flow at depth dominates the way the stresses are released in the brittle crust. Faults may cross and exist side by side at close distances.

Volumetric strain energy can be built up and stored for a longer time in the intermediate layer than the shear strain. This may be the preparation stage for sudden intrusions and phase transitions. The processes of the strain transmission along the rifting zones and transform zones at depth could then locally lead to fast build–up of strain and earthquakes.

Local lowering of viscosity as indicated below the seismic zones and in the rift zones is caused by high strain rates which destroy the crystal structure and lead to higher state of entropy and to more mobility between particles.

The fluid mobility in the intermediate layer leaves fluids trapped at the base of the brittle crust, possibly in bubbles carrying high pressures from below, which cause extensional cracking and dilatancy and earthquake triggering by "hydrofracturing mechanism". A clear example of this is found in the precursory period of a 5.8 earthquake in SISZ (Ágústsson et al., 1996) and expansion related to ongoing swarm in SW–Iceland (Rögnvaldsson et al., 1996).

Strain waves causing short term coincidences of events are related to large aseismic events in the lower crust, coinciding with fluids/magma intrusions into the brittle crust. They are channelled by rift zones and the seismic zones overlying low viscosity material triggering events on their way. The strain pulse would in such cases be much less attenuated than the instantaneous strain pulse. Only the largest strain waves are observed by their effects.

There is a crustal material flow to the SW and to N and NW of the plume center, which is faster than the general plate divergency in this part of the ridge so the stresses outside the plume center tend to be compressive. This flow is enhanced in the zones of weakness, i.e. the seismic zones and the rift zones.

The above conclusions encourage more continuous monitoring of the crustal deformation. This can be done by monitoring changes in stress orientations as inferred from microearthquakes, by continuously recording strainmeters, gravimeters and by continuous GPS.

The existence of strain waves as a trigger is of great significance for prediction. Above the Iceland hotspot this is more obvious than elsewhere because of the intensive crust/mantle interaction. But the interaction between a convective mantle and the brittle crust is everywhere. Low resistivity is found world wide in the lower crust (Hyndman et al., 1993).

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