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Volcanoes, fountains, earthquakes, and continental motion – What causes them?

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Abstract. As a common mechanism for the various volcanic phenomena on Earth we point out a severe disobeyance of Poincaré's (von Zeipel's) theorem: hot, gas-rich, high-pressure "fingers" (diatremes) can grow out of the boundary layer above a molten domain and thrust their way up from the asthenosphere toward the surface. The isobars of a planet or moon can look like the surface of a bed of nails. Linear arrays of high-pressure diatremes can drive continental motion. Moreover, we hold the tidal torque responsible for magnetic dynamo action.

Key words: Volcanism – Continental motion – Plate tectonics – Earthquakes – Outgassing – Diatremes – Magnetic dynamo

Introduction

During the past 20 years Wegener's conviction of moving continents has been corroborated by a large body of geophysical evidence (Baumann, 1984; Bonatti and Crane, 1984; Burchfiel, 1983; Closs et al., 1984; Francheteau, 1983; Jordan, 1984; McKenzie, 1983; Moorbath, 1984; Toksöz, 1984; Wilson, 1984; Worsley et al., 1984; Runcorn, 1980; Loper, 1985). At the same time, volcanism and its driving forces have been thoroughly studied and have been related to plate tectonics, continental growth, earthquakes, and mineral deposits (Williams and McBirney, 1979; Dawson, 1980; Huppert and Sparks, 1984; also Tryggvason et al., 1983; Wilson and Head, 1983; Baumann, 1984; Burbank and Reynolds, 1984). Studies of the ocean basins have revealed submarine rifting, seafloor spreading, transform faulting, hot fountains (black smokers), and bizarre ecosystems feeding on H_2S (Sclater and Tapscott, 1984; Bonatti and Crane, 1984; Macdonald and Luyendyk, 1984; Edmond and von Damm, 1984). Moreover, there are the phenomena of mud volcanoes created by escaping gases, predominantly methane and water vapor (Gold and Soter, 1980), of peatlands (Foster et al., 1983), and of combustible gas eruptions during earthquakes most of which are thought to be of biogenic origin but whose mass rates are so high and whose correlations with plate boundaries are so strong that an explanation via outgassing of the Earth's mantle may well be indicated (Gold, 1979; Gold and Soter, 1982; Giardini and Melton, 1983). Note that volcanic gases are predominantly composed of water vapor plus CO_2, H_2S, SO_2, and HCl, whereby these gases dominate at exhalation temperatures increasing from below 100 K (CO_2) up to 900 K (HCl), respectively. Instead, the Earth's interior may be in a more reduced state, with dominating contributions of CH_4, H_2, and atomic carbon – as is evidenced by occasional gas inclusions in diamonds and in quartz, by the existence of mud volcanoes, and by measurements on minerals from mafic rocks that cool at high pressures (Mathez, 1984; Freund, 1980), and as is expected in view of the composition of the carbonaceous chondrites which were probably major building blocks of the Earth, at least during late stages of its formation.

In spite of all this detailed knowledge on the geometrical, thermal, and chemical structure of volcanism, fountains, earthquakes, outgassing, and plate tectonics, there does not seem to be a complete understanding of their driving forces and causal connections such that unique predictions for other planets (Mars, Venus) and satellites (Moon, Io) could be made. For instance, it is not clear whether the huge volcanoes and valleys on Mars were formed by lava or by glaciers (cf. Wilson and Head, 1983; Gold, 1978), and the sulfuric volcanoes on Io were a surprising discovery. Furthermore, what is the connection between spin and magnetism? Parker (1983) points out that magnetic dynamos depend on circulation patterns with nonzero helicity (cyclonic convection), but that the almost synchronous planet Mercury ($P_{\text{spin}} = 2/3\, P_{\text{orbit}}$) has a polar field strength of $3.5 \cdot 10^{-7}$ T, whereas the fast-spinning planet Mars has not revealed a magnetic field of its own.

In this paper we point out a disobeyance of Poincaré's (von Zeipel's) theorem which states that in a gravitating, rigidly rotating fluid body (of simple chemistry), the level surfaces ($p = \text{const}$, $\rho = \text{const}$, $T = \text{const}$) are all identical and agree with the surfaces of constant geopotential. (Small deviations – the Eddington-Vogt meridional circulation – are caused by a non-conserved cooling flow; they are negligible for cool, slowly spinning planets like Earth. We also ignore long-time-scale rearrangements between highly viscous fluids). Instead, the boundary layer between a fluid domain and a solid crust is unstable to the growth of hot (light, vertical), gas-rich fingers (or channels, pipes, chimneys, diatremes, conduits) which allow low-lying isobars to almost touch the surface, in the shape of a fakir's bed of nails (see Fig. 1). When these fingers end in (porous) sand or clay deposits near the surface with a thickness of at least a few 10^2 m, they may create a mud volcano; when they

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end below solid rock, they will blow or melt their way out in the form of an ordinary volcano or kimberlite; and when they pierce a subsurface porous layer, they will pump it up like an air cushion until fissures open and allow a sudden discharge in the form of a shallow earthquake.

The overpressure exerted by a linear array of hot fingers – an "isobaric fence" – is argued to be strong enough to cause continental motion (plate tectonics).

We shall discuss isolated diatremes in Sect. 1, chains of diatremes in Sect. 2, and their various geophysical consequences in Sect. 4. Section 3 is devoted to tidal forces and their possible importance for the magnetic dynamo.

1. Isobars of a Geoid

In a heavy fluid body, the equations of static equilibrium imply that the surfaces of constant pressure and mass density coincide with those of constant potential; in a rigidly rotating body, the potential has to include that of the centrifugal forces.

A different situation arises when solid crust, or lithosphere, overlies a warmer partially molten asthenosphere. As soon as the fluid/solid boundary layer develops small deviations from a planar (spherical) geometry, the domes of this surface tend to be nearer to the melting point than the lower parts because the convective thermal gradient (of the fluid) is smaller than the conductive gradient (of the solid). Moreover, if this liquid contains dissolved gases like CH_4, H_2, etc. and dissolved light salts (due to their large ionic radii) like those of the radioactive elements U, Th, and K, these light ingredients will tend to concentrate near the domes.

All three properties, convection, reacting gases, and concentration of radioactive elements, tend to partially melt the ceiling of a dome, thereby enhancing the non-planar geometry: a hot finger (diatreme) starts growing upward. This is clear for the first property because a decreasing pressure (at fixed temperature) tends to induce melting. It is equally clear for radioactive heating. Concerning hydrogen-rich gases, we assume that they are stable in the high-pressure, hot mantle whereas they convert to CO_2 and H_2O near the surface (cf. Gold and Soter, 1982, Fig. 1). Le Chatelier's principle (of the yielding to changes of an intensive thermodynamic variable) then implies that heat is stored in the virtual reaction in which the gases are formed by heating and compression. By conservation of energy, the inverse reaction must be exothermic.

When a dome grows a hot vertical finger, this finger is lighter than its environment because it is hotter and contains more light ingredients. The pressure at the top (ceiling) of the finger is therefore higher than that of its environment. This local overpressure must tear the ceiling, and light gases (or fluids) of low viscosity can enter the fissures, thereby causing overhead stoping (hydraulic fracturing: large chunks of solid rock will sink down the chimney while fresh, hot magma from the mantle rises and replaces them). The longer the finger, the larger the overpressure near its ceiling and the larger the thermal and pressure contrast (see Fig. 1).

Note that our suggested instability of the lithosphere to the growth of hot fingers is different from the ordinary well-studied Rayleigh-Taylor instability of viscous fluids which can give rise to the formation of thermo-chemical plumes in the mantle (Christensen, 1984), or to the formation of salt domes in sedimentary strata (Woidt, 1978). Hot fingers grow via overhead stoping through solid rock on the time scale of days to years, whereas plumes grow hydrodynamically on the time scale of My or longer.

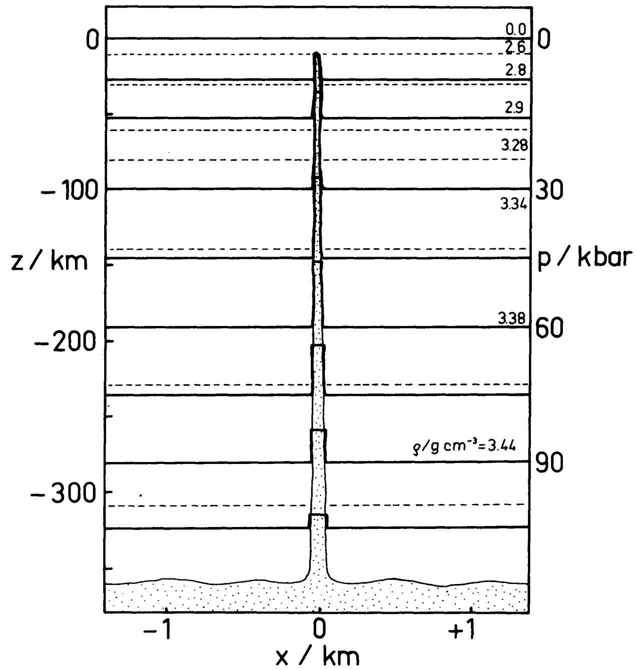

Fig. 1. Schematic isobars (heavy) and density discontinuities (thin and dashed) in the vicinity of a hot finger (diatreme). The horizontal distance scale has been enlarged by a factor of 10^2. The magma has been assumed to have an average density of $\rho = 2.7\ \mathrm{g\ cm^{-3}}$. Note that in extreme cases, isobars can jump by more than 50 km in height at the edge of the magma conduit

Why do we expect the growth of a finger to be stable, i.e., to continue up to the vicinity of the surface? Several conditions have to be satisfied. To begin with, cooling of the finger at its ceiling and along its walls must be compensated. Without exothermic reactions, heat losses per volume q at the ceiling can be overcome by convective replacements (at typical speed v) if the vertical growth rate $\dot{z}$ satisfies

$$\dot{z}/v \leq \zeta c \Delta T/q \quad \text{and} \quad \dot{z}/v \ll 1, \tag{1}$$

where c is the specific heat per volume of the rising magma, ΔT is the temperature contrast between bottom and top of the chimney, and $\zeta \lesssim 0.5$ measures the degree to which ΔT is used for heating the ceiling. We estimate $\dot{z}/v \lesssim 10^{-1}$ because $q \lesssim c\,T_{\mathrm{melt}}$ for heating and partial melting.

Secondly, heat losses $2\pi R \Delta z S$ through the wall can be overcome by convective supply $\pi R^2 c \Delta T v$ through a sufficiently large cross section of the chimney (of radius R); the condition reads (for $S = cD\Delta T \approx cD\delta T/R$, D = thermal diffusion coefficient)

$$R \gtrsim 2[(\Delta z D/v)(\delta T/\Delta T)]^{1/2} \approx 10\ \mathrm{m}(\Delta z_5/v_{-2})^{1/2}, \tag{2}$$

where Δz_5 stands for $\Delta z/10^5$ m, $v_{-2} := v/10^{-2}\ \mathrm{m\ s^{-1}}$, and where we have inserted $D = 10^{-6}\ \mathrm{m^2\ s^{-1}}$, $\delta T/\Delta T \approx 3$. Quite likely, convective velocites v can be 10 to 10^2 times larger than inserted, so that heat losses through the walls are unimportant for diameters in excess of a few meters.

A third condition to be met concerns convective instability. For this to be satisfied, the adiabatic temperature gradient $(\nabla T)_{\mathrm{ad}} = (dT/dp)_{\mathrm{ad}} |dp/dz| = \rho g \alpha / c_p$ with ρ = mass density, g = gravity acceleration, and $\alpha := |\partial \ln \rho / \partial \ln T|_p$ (Lang, 1974, 3–297) must not exceed the structural gradient $|dT/dz| \approx \Delta T/\Delta z$, whence

$$\Delta T/T \gtrsim \rho g \Delta z \alpha / c_p T = \Delta p \alpha / c_p T \approx 10^{-2} \Delta z_5 \quad (3)$$

for a chimney height of $\Delta z = 10^5$ m, temperature of $T \gtrsim 10^3$ K, $\alpha \approx 10^{-2}$, and a typical specific heat per volume $c_p = 10^{6.3}$ J/m^3 K ~ 0.2 cal/g K. In the absence of friction, a temperature drop of a few K per 10 km in the chimney is therefore well sufficient to maintain convective overturn.

As a fourth condition, wall friction limits the convective velocities v. Clearly, buoyancy forces $\pi R^2 \rho k \Delta T / \langle m \rangle$ in the chimney (with k = Boltzmann's constant, ΔT = structural minus adiabatic temperature difference, and $\langle m \rangle$ = mean atomic weight) must be larger than viscous shear forces $2\pi R \Delta z \eta \nabla v \gtrsim 2\pi \Delta z \eta v$ so that v is bounded by

$$v < R^2 k \Delta T / 2\nu \Delta z \langle m \rangle = 10^{-1}\ \mathrm{m\ s^{-1}}\ R_{0.5}^2\ \Delta T_2 / \Delta z_5 \quad (4)$$

with $R_{0.5} := R/3$ m, $\Delta T_2 := \Delta T/10^2$ K, and where we have used $\nu = \eta/\rho \lesssim 30\ \mathrm{m^2\ s^{-1}}$ for the kinematic viscosity coefficient ν (of silicic magma). Apparently, convective velocities $v \gtrsim 10^{-1}\ \mathrm{m\ s^{-1}}$ are permitted for chimneys of diameters in excess of several meters and driving temperature drops in excess of 1 K/km (cf. Wilson and Head, 1983). Considerably higher surging velocities ($R^2 \Delta T/\Delta z \gg 10^{-2}$ K m) are needed to explain the formation and survival of xenoliths (Dawson, 1980).

A fifth condition for convective overturn is Rayleigh's criterion, which wants convective heat transport to be fast compared with conductive transport. For vertical pipes, the Rayleigh number

$$Ra := \alpha g \cdot \nabla \ln T R^4 / \nu D = 10^{1.5} (\Delta T / T \Delta z)_{-6}\ R_1^4 / \nu_{1.5} \quad (5)$$

must be larger than 67 (Landau and Lifshitz, 1966, § 56, exercise 6), where $D \gtrsim 10^{-6}\ \mathrm{m^2\ s^{-1}}$ is the thermal diffusivity of the magma (= thermal conductivity per specific heat), and ν has been inserted for silicic magmas. This criterion is easily satisfied both for deep pipes of mafic magma ($\nu_{1.5} \approx 10^{-4}$) with $R \gtrsim 3$ m and for shorter pipes of silicic magma ($\nu_{1.5} \approx 1$).

Sixthly, we have to argue that the walls of a diatreme can take the enormous overpressures implied by their height and weight contrast (cf. Fig. 1). Drilling experience teaches that hydraulic fracture sets in at overpressures exceeding several 10^2 bar (Williams and Birney, 1979, p. 57). However in our case, the rising magma heats the walls to almost melting temperature so that the rocks tend to expand and flow (rather than tear, like cold rock). Moreover, silicic magma is some 10^8 times more viscous than the aqueous fluids used for hydraulic fracture so that it cannot easily penetrate thin fissures. We therefore expect the walls not to fracture or yield (on the time scale of years) even for pressure differences δp exceeding 10 kbar.

Note that pressure differences δp of this magnitude are needed to throw rock over distances of $d = 10$ km, as has been recorded for volcanic eruptions. The energy $Mv^2/2 = (Mgd/4)(\mathrm{tg}\,\vartheta + \mathrm{ctg}\,\vartheta) = Mgd/2\varepsilon$ (with ϑ = elevation angle, $\varepsilon \leqq 1$; $\varepsilon = 1$ for $\vartheta = 45°$, $h = d/4$) needed to throw a body of mass M in the Earth's gravitational field over a distance d (with peak height h of the ballistic orbit) must be supplied by a pressure difference δp between bottom and top of the body acting over a distance δx. Here we have ignored air friction. This leads to $\delta p A \delta x \geqq Mgd/2\varepsilon$ where A is the cross-sectional area. Now for $d \approx 10$ km, v must be on the order of the speed of sound in air. Such a high speed can be acquired inside a gun but not inside a diatreme. Consequently, the driving pressure δp can only act near the outlet, where the pipe widens conically, and must quickly drop to zero. We write $\xi := A \delta x / V$, where $V := M/\rho$ is the volume of the body, and estimate $\xi \lesssim 0.5$. This leads to

$$\delta p \geqq \rho g d / 2\xi\varepsilon = 3\ \mathrm{kbar}\ (d_4 / 2\xi\varepsilon). \quad (6)$$

Very likely, $2\xi\varepsilon \ll 1$ holds so that pressure differences (in the released and expanding gases) on the order of 10 kbar at the outlet of a diatreme are needed to throw rock over a distance of 10 km.

On top of meeting these elementary constraints, a hot finger cannot grow quasi-cylindrical unless the conditions at its ceiling are highly inhomogeneous, favoring a tearing and melting near the highest point; it would otherwise open up in the shape of an inverted cone. Most likely, an excess of light (and combustible) gases near the highest point helps to keep a hot finger thin.

Naively one might think that convection establishes an ordered circulation pattern with fresh, hot magma rising near the axis and colder material falling near the walls. The necessary horizontal temperature gradient would imply a horizontal density gradient (for homogeneous chemistry) which in turn would imply higher pressures near the axis than near the walls (for a uniform pressure at the bottom of the finger). The resulting horizontal pressure gradient would cause the light, hot component to move toward the walls, and the colder, descending component toward the central region, thereby wiping out the horizontal temperature gradient. Consequently, there cannot be horizontal gradients inside a chimney. Rather, the colder component is expected to sink in the form of chunks or drops through the rising, warmer component. The pattern may be stabilized by the fact that the necessary horizontal gradients near the walls will send cooler (heavier) material toward the axis, thereby preventing freezing from the sides.

The real structure of a hot finger will be further complicated by the presence of double-diffusive convection which can drive vertical mass exchange and establish compositional gradients. Huppert and Sparks (1984) speak of "convective fractionation" when they discuss the consequences of density changes during partial crystallization (see also Spera et al., 1984). In particular, thin laminar boundary-layer flows can form. Of course, the growth of a finger will not only depend on its own chemistry but also on the inhomogeneous structure of the lithosphere and crust through which it pushes its way up and on the conditions which it meets near the planetary surface. We shall discuss a number of such possibilities in the last section.

2. Plate tectonics

The lithospheric plates move with respect to each other at speeds reaching and exceeding 10^{-1} m/y. Along their divergence zones (spreading axes), hot magma surges and replenishes the crust, while at the opposite end the pushed, cooled plate is forced underneath an adjacent plate in the form of a subduction zone because plate area is conserved (see Fig. 2). What drives this motion?

It is often believed that convective motions inside the (upper) mantle are responsible for the motion of the continents so that the driving power is convective cooling of

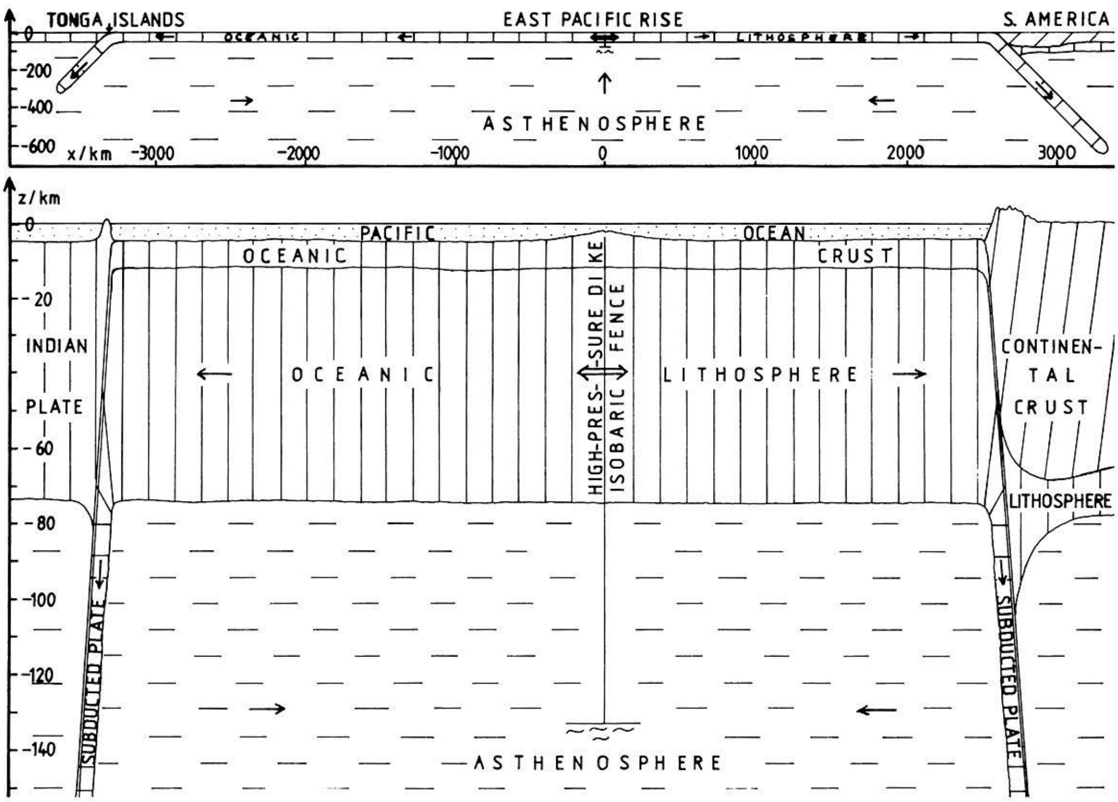

Fig. 2. Cross section through the lithosphere underneath the South Pacific ocean; upper drawing to scale, lower drawing the same with vertical scale enlarged 20-fold. Oceanic crust and lithosphere (total thickness of 70 km) move apart along the East Pacific Rise at a speed of some 15 cm/y and are subducted at the Tonga trench in the west and below the Andes in the east where they descend at an angle of $\approx 45°$. Note how thin a layer is moving. We hold a deep-rooted, high-pressure dike responsible for propelling this global circulation; the dike pushes the two plates apart with an overpressure on the order of unity

the Earth's interior (e.g., Runcorn, 1980; Loper, 1985). But why are such motions preferentially oriented east-west, whereas surface temperature gradients are oriented north-south, and why are there only some six large plates instead of thousands of small ones?

Along a divergence zone hot magma rises to the seafloor in the form of a chain of small volcanoes (hot fingers) or a sheet (dike) of rising magma. In such a hot dike, overpressures must build up intermittently which are comparable to the ones evaluated in the last section: the nailboard of isobars degrades into an "isobaric fence." Such an isobaric fence must have the same effect as an array of wooden wedges which can split stone when watered or as a number of high-pressure concrete injections into the ground that can lift a building. The fence pushes adjacent plates apart. Essentially this same mechanism has been reviewed by Jacoby (1980) and credited to Lliboutry (1972).

In order to see quantitatively that a chain of high-pressure vertical pipes can push two plates apart, we estimate the shear modulus μ of solid rock by

$$\mu \lesssim p \approx \rho g h = 10^9 \text{ Jm}^{-3}\, h_{4.5}, \tag{7}$$

where h is the height of an oceanic plate, on the order of 30 km. Somewhat more reliable would be the molecular estimate

$$\mu \lesssim e^2 (2a)^{-4} (m_e/m_p)^{1/2} = 4 \cdot 10^{10} \text{ Jm}^{-3} \tag{8}$$

for the shear modulus μ of "crud" (Press and Lightman, 1983; a = Bohr's radius) which agrees with the values determined both from the propagation of shear waves ($\mu = \rho v_s^2$) and with the reaction of the mantle to tidal forces (Jeffreys, 1970) but which tends to overestimate the long-term yield strength of realistic crust material by a factor on the order of 10^2. We shall prefer Eq. (7) because of its simplicity, yet keep in mind that it must not be applied to very shallow or very deep layers. Note that in a fluid mantle, shear forces would be due to viscosity, with $\mu = \eta |\nabla v|$, where η is the dynamic viscosity, on the order of $\lesssim 10^6$ poise ($= 10^5$ Ns/m^2) for silicic (acid) magmas (near the surface) and on the order of 10^2 poise for (deeper) mafic (alkalic) magmas (e.g., Huppert and Sparks, 1984). An equality of solid shear stresses with viscous shear stresses would therefore ask for shear velocity gradients $|\nabla v| = v/h$ on the order of

$$v \lesssim \rho g h^2/\eta = 10^{8.5} \text{ m s}^{-1} (h_{4.5}^2/\eta_5), \tag{9}$$

which correspond to relative velocities near that of light. We infer that fluid viscosity is negligible compared with that of a solid.

Let us then equate the force $\delta p h l$ – exerted by a temporary overpressure fence of height h (between 50 and 10^2 km) onto the face of a square-shaped lithospheric plate of length l ($\approx 10^3$ km) – with the shear force $\mu l b$ needed to push the plate across an effective area lb of solid resistance (the path length b is expected to be on the order of the penetration depth of the plate into the subduction zone). We find that

$$\delta p/p \lesssim b/h; \tag{10}$$

i.e., the needed overpressure $\delta p/p$ is on the order of unity. This estimate is uncertain because it is not clear whether the plate rides on a convection cell of the upper mantle, and whether its diving front end can locally crumble or melt the rock into which it submerges. Additionally, the effective shear modulus μ may be smaller than $\rho g h$ when the rock has structural flaws. However, the estimate suggests that temporary overpressures on the order of unity can push two plates apart under favorable circumstances. A glance at Fig. 1 shows that such temporary overpressures can be provided by a dense chain of rising hot fingers, each of which acts as a spreading center.

The literature often declares buoyancy forces to be responsible for causing plate motion (e.g., Loper, 1985). "Buoyancy forces" are understood as vertical forces exerted by a fluid medium on immersed objects; such forces per area are on the order of $\delta \rho g h$, where $\delta \rho$ is the difference in mass density, and h is the vertical extent of an immersed object. In contrast, the pressures $\delta p \approx \rho g h$ exerted by a deep-rooted isobaric fence can be some $\rho/\delta \rho \approx 10^2$ times higher and, moreover, act as a horizontal thrust, not a vertical pull (cf. Fig. 2).

The overpressure δp exerted by a hot finger discharges when the finger reaches the surface (seafloor). During a lava ejection, the pressure relaxes and the finger cools. An isobaric fence is therefore not static: it behaves like a collection of relaxation oscillators. Its overpressure builds up in a quasi-periodic fashion. It derives its power from thermal convection in the mantle during which hot magma rises in a divergence zone and cooler material sinks at the other end of the plate in the subduction zone.

We can therefore estimate the T-gradient of seafloor spreading from the condition that the buoyant energy gain per time, $P_{\text{therm}} \approx l b v n (f/2) k \Delta T$, in the rising column of a global convection cell balance the dissipative losses $P_{\text{diss}} \approx \mu l b v$ (mainly) at the other end of the cell (subduction zone). Here it is assumed that the plate floats on the convection cell so that the speed v of its horizontal motion equals the average convective speed of the rising magma, and that the width b of the rising column is comparable with the width of the resistive zone. ΔT is the driving temperature difference between bottom and top of the cell, $n = \rho/\langle m \rangle$ = atomic number density, f = number of thermal degrees of freedom ≈ 3, and k = Boltzmann's constant. From the condition $1 \lesssim P_{\text{therm}}/P_{\text{diss}}$ with $\mu \lesssim \rho g h$ we thus find

$$1 \approx f k \Delta T / 2 \langle m \rangle g h \approx (\Delta T)_3 / h_5 \tag{11}$$

for $\Delta T \approx 10^3$ K and a plate height h of 10^2 km, in remarkable agreement with the observations. This estimate shows that convection cells can power plate motion once they are set up.

There is, however, the yet open question of stability: How did the large-scale convection cells come into existence, i.e., why did Pangea break up some $2 \cdot 10^8$ y ago and why does continental motion persist? The (present) heat-energy content of the $3 \cdot 10^3$-K Earth is some $10 k T / \langle m \rangle R^2 \Omega^2 \approx 10^2$ times larger than its (present) rotational energy, and some $k T R / \langle m \rangle g (\Delta h)^2 = 10^6 (\Delta h)_3^{-2}$ times larger than the energy $\rho g A (\Delta h)^2/2$ of gravity anomalies due to changing surface loads (such as the melting of ice). We therefore consider volcanism (convective cooling) the only viable powerhouse for driving continental motion. Isobars in the shape of an oscillating bed of nails are likely to make the shells move wherever the former cluster in linear arrays, i.e., wherever they form fences.

At this point, smaller energy reservoirs may enter the scene in controlling the initial location of the volcanic diatremes, thereby enforcing the observed large-scale plate morphology. Such causes for large-scale order are (1) continental insulation (causing a heat accumulation and melting under central parts of a continent), (2) a (minor) shrinking of the cooling Earth (causing the thin lithosphere to wrinkle), (3) a wandering of positive gravity anomalies toward the equator (thereby changing the curvature of a continental shell; Anderson, 1984), (4) localized tidal dissipation (causing the upper mantle to melt preferentially in inelastic regions of the equatorial belt); and (5) cratering by impact: their circular edges are natural sites for subduction zones. Either one or a combination of these processes is likely to have caused an inhomogeneous shape of the initial "isobaric nailboard," with high-pressure volcanic fences being preferentially arranged in strips aligned with meridional circles, predominantly so along an equatorial belt.

Once the lithosphere starts tearing along a few fissures filled with hot (light) magma, a convective circulation pattern in the mantle can be started that has local ($\gtrsim 10$ km) negative gravity anomalies above the rising columns (= spreading axes; Macdonald and Luyendyk, 1984) and extended ($\gtrsim 300$ km) positive gravity anomalies above the sinking columns (= subduction zones; Toksöz, 1984). The likewise observed extended ($\gtrsim 300$ km) positive gravity anomalies around spreading axes (Dixon and Parke, 1983) can be understood as the result of horizontal thrusting: the pushed plate bulges up. Enhanced cooling of an aging oceanic plate guarantees that sinking columns of a convection cell are kept cooler than rising ones, thereby keeping the heat engine at work. The idealized pattern of hot, upwelling magma closing the gaps between diverging plates whose cool opposite ends sink and fuse with the upper mantle appears to be dynamically stable when one accepts the presence of high-pressure volcanic conducts which push the plates apart. Note that in the absence of such quasi-permanent over-pressure along the spreading axes, the moving plates are likely to be arrested by friction forces and the fissures likely to freeze. Continental shell motion owes its existence to a drastic violation of Poincaré's law.

3. Tidal torque and the magnetic dynamo

The present spin of Earth is controlled by tidal forces; let us compare them with the forces controlling plate tectonics. The tidal torque T exerted by one body of a celestial binary system onto its companion is given by (cf. Goldreich and Peale, 1968)

$$T = (3k/2)(G M_*^2 R^5 / a^6) \sin 2\delta. \tag{12}$$

Here, M_* is the mass of the disturbing body, R is the radius of the disturbed body, a is their separation, $k := (3/2)(1 + 19\mu / 2 g \rho R)^{-1}$ is the "tidal Love number," and δ is the (small) phase lag angle of the tidal bulge. δ is related to the Q factor of the tidal oscillations by $\tan 2\delta = Q^{-1}$, where Q for the Earth's mantle is probably of order 10^3 (Lambeck, 1980).

However, oceanic tides achieve much stronger dissipation, corresponding to a (present) effective $Q = 12$ or $\sin 2\delta = 0.08$. Together with an empirical $k = 0.3$, one thus obtains for the present tidal torque exerted on Earth

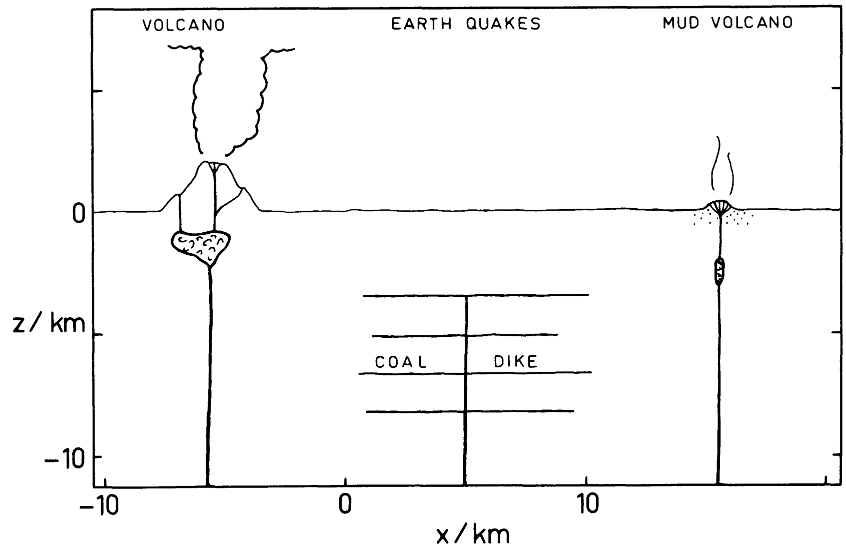

Fig. 3. Sketch of the different expected surging modes of a hot finger (diatreme) depending on the ground structure near the surface. A cap rock gives rise to volcano formation, a sand layer to the formation of a mud volcano, and a porous subsurface layer (of biogenic origin) can lead to (shallow) earthquakes and to the formation of coal seams

$$T_{\oplus} = 5.3 \cdot 10^{16}\ \mathrm{N\,m}, \tag{13}$$

in which the Sun has a share of 21%. For a moment of inertia $I \approx MR^2/3 = 0.8 \cdot 10^{38}$ kg m^2, this torque corresponds to the observed present average growth rate of $2 \cdot 10^{-5}$ s/y of the length of the day. Note that this rate would be some 10^2 times smaller if the effective Q were $\approx 10^3$, i.e., if there were no oceans on Earth.

The tidal forces are much smaller on the average than the volcanic forces which are thought to push the continental plates: the volcanic forces acting east-west across one midoceanic ridge exert a partial torque $phlR \lesssim \rho g h^2 lR \approx 10^{27.5}$ N m h_5^2 which is some 10^{11} times stronger than the tidal torque. Nevertheless, tidal forces may not be negligible for three reasons: (1) they can be much stronger locally when oceans are excited not too far from resonance, (2) their dissipated energy is an inhomogeneous heat source for the Earth's crust (as noted above), and (3) their net torque does not vanish so that they brake the spin motion.

If essentially the whole tidal torque on Earth is exerted through its oceans, the interior of the planet must be decelerated through internal shear forces rather than through smooth gravitational (body) forces. The implied average shear modulus $\langle\mu\rangle$ is given by

$$\langle\mu\rangle = T/2\pi R^3 = 3 \cdot 10^{-5}\ \mathrm{N\,m^{-2}}. \tag{14}$$

When compared with Eqs. (7) and (8), this needed shear modulus is some 10^{13} times smaller than that of solid rock. A layer of mafic magma, on the other hand, would need a (large!) shear-velocity gradient of $|\nabla v| \geqq 3 \cdot 10^{-6}\ \mathrm{s^{-1}}$ in order to transfer the torque, which corresponds to a velocity profile of 1 mm s^{-1} across 300 m. We conclude that a solid Earth would rotate rigidly whereas the existence of a fluid layer implies differential rotation.

Now it is known from the propagation of shear waves, the weak response of Earth to tidal forces, and its near-rigid free precession that the mantle behaves almost like a solid body with a (large) effective viscosity $\eta = \mu\tau$ in excess of several 10^{20} poise (cf. Jeffreys, 1970, p. 345). On the other hand, the closeness of the mantle to isostatic equilibrium and the feasibility of continental motion set a rough upper limit of $\eta \lesssim 10^{23}$ poise, because Maxwell's relaxation time scale τ (for plastic flow of the asthenosphere) is on the order of $\tau = 10^{5.5\pm1}$ y (see also Vetter et al., 1980). However, it is equally known that the outer core is fluid. Consequently, we expect the Earth's core to be less decelerated than the mantle, i.e., to spin faster. Instead, magnetic anomalies are known to drift westward during the past (at least) 300 y at an equatorial speed of 0.64 mm s^{-1} (Lambeck, 1980; Morrison, 1985; Bloxham and Gubbins, 1985).

If we interpret the westward drift of the magnetic anomalies as a relative rotation between the flux-generating core and the mantle, we are forced to conclude that the spin of the core oscillates, coupled to the mantle by the magnetic field. According to Eq. (14), the necessary toroidal magnetic field component B_φ amounts to

$$B_\varphi = 8\pi\langle\mu\rangle/B = 10^{-6}\ \mathrm{T} \tag{15}$$

for a total field B of 1 Gauß ($= 10^{-4}$ T). Its oscillation period P would be given by

$$P = 2\pi[I_{\mathrm{core}}/(dT/d\theta)]^{1/2} = 2 \cdot 10^3\ \mathrm{y}\ (d \ln T/d\theta)^{-1/2}, \tag{16}$$

where $I_{\mathrm{core}} \approx 5 \cdot 10^{36}$ kg m^2 is the estimated moment of inertia of the core, and $dT/d\theta$ is the change in the torque per change in the torsion angle between core and mantle (= torsion constant). $d \ln T/d\theta$ would be on the order of 10^2 if the magnetic fields were also anchored in the (decelerated) mantle, but may well be $\lesssim 10$ due to the poor electrical conductivity of the latter. We thus arrive at the prediction that the magnetic anomalies fluctuate with a period on the order of 10^3 y.

This interpretation of a magnetic coupling between core and mantle of a spun-down planet suggests that the tidal torque may be driving the magnetic dynamo. It is supported by the fact that the magnetic dipole moments of Mercury, Venus, Earth, Moon, and Mars do not scale as their angular velocities but rather as their tidal decelerations (cf. Parker, 1983).

4. Volcanic activities

The surfaces of several planets and satellites are controlled by volcanism. This is already indicated by the fact that the highest mountains tend to be as high as they can be given the finite yield strength of the underlying lithosphere (cf. Weisskopf, 1975) and despite efficient erosive processes. On the present Earth, crustal volume is newly created at a rate of $\dot{V} \lesssim 10^{2.5}\ \mathrm{m^3\,s^{-1}}$ by seafloor spreading along the oceanic ridges. The ejection by one volcano like Tambora (1815) every 10^2 y would amount to $\dot{V} \approx 10^2\ \mathrm{m^3\,s^{-1}}$ (Stothers, 1984), and the remaining $\gtrsim 600$ active volcanoes together do not fall behind by a large factor (cf. Williams and McBirney, 1979; LaMarche and Hirschboeck, 1984). The erosive volume rate is (independently) estimated as $\dot{V} \approx 10^2\ \mathrm{m^3\,s^{-1}}$, or $\dot{M} = 10^{5.5}\ \mathrm{kg\,s^{-1}}$, as is to be expected for a steady state.

The Earth has a volume V of $10^{21}\ \mathrm{m^3}$. A present volcanic volume rate of $\dot{V} \lesssim 10^{2.5}\ \mathrm{m^3\,s^{-1}}$ therefore means that $\lesssim 0.05\ V$ has been turned over throughout its history, ignoring a likely higher rate at earlier times. If only the upper mantle (with a volume of $\lesssim 0.3\ V$) has participated in this overturn, its average overturn probability exceeds 10%. These numbers illustrate the importance of volcanism for the surface structure of our planet. What are the causes for such an activity?

We propose that the ejection of mantle material to the surface is brought about by the formation of hot fingers, due to the instability of a liquid/solid boundary layer. Clearly, such volcanic activity culminates in divergence zones. Note that there are no mass ejections along transform faults even though crustal rocks have slid past each other: without a local overpressure, torn crust is welded again and no ejection occurs. On the other hand, volcanoes can form in the middle of a plate in a static region through the action of a hot spot which we interpret as a large (old) hot finger (cf. McKenzie, 1983; Anderson, 1984). Given our above estimates, such fingers very likely cannot pierce a lithosphere of arbitrary thickness, nor could they have pierced the present lithosphere of Earth without the help of dissolved gases and/or radioactive elements. It is not clear to us whether past measurements have been able to discriminate between radioactive heating of the crust and heating by nonemerged hot fingers (cf. O'Nions and Oxburgh, 1983).

At this point, it is time to discuss the different ways in which a hot finger can make its appearance at the surface. The most spectacular form of volcanism is encountered when the rising magma does not intersect a loose or porous layer so that it conserves its enormous overpressure until very near the surface ($\gtrsim 300$ m). It can then lift the ground, blow off the top of a mountain, and eject material at velocities of $\lesssim \mathrm{km\,s^{-1}}$ to heights in excess of 30 km. Kimberlites are another form of remnant of such pyroclastic events. How much material is ejected will depend on the geometry of the diatreme (hot finger), i.e., on its diameter, length, and in particular on the existence of large high-pressure magma chambers which can form on encounter with an easily melting zone. Another necessary condition for explosive ejection is a (silicic, acid) magma of high viscosity ($\eta \lesssim 10^6$ poise) so that the dissolved gases (H_2O, CO_2, etc.) cannot escape beforehand through thin fissures. Note that during the explosion, their volume increases by a factor of $\lesssim 10^3$. If such a discharge happens under the surface of the ocean, a tsunami will result.

In the case of (mafic, alkalic) magmas of low viscosity ($\eta \approx 10^2$ poise), on the other hand, the formerly dissolved gases will largely escape before outburst, thereby releasing the enormous overpressure and giving rise to fire fountains and/or magma flows. Near the surface of a volcano, a hot finger can branch into several and give rise to independent outbursts. A stagnating side branch may thereby freeze and lock itself off from the main conduct.

A rising, mafic magma finger may also grow into a porous surface layer, with a thickness of several 10^2 m, through which the gases can easily escape. Its first outburst may be violent, but subsequent outbursts need no longer supply hot, burnt gases. We get a cool mud volcano, inside of which a substantial fraction of the dissolved H_2O vapor has condensed out (whence the mud).

When a mafic magma finger pierces a subsurface porous layer (of biogenic origin) its partially dissolved gases will penetrate this layer and pump it up like an air cushion. This process may take several years and extend horizontally over several 10^2 km. When a critical thickness of the cushion is reached, the gas will eventually find a way out, i.e., we get an earthquake. Another possibility is peaceful, steady outgassing observed particularly in lakes and valleys. In all these cases, however, we think that the escaping gases made most of their way up through one of those diatremes in the form of channelled surging which is much faster than any diffusive process. There may be many more diatremes than volcanoes.

If this scheme is correct, it implies that an earthquake which is unrelated to a moving plate boundary marks the momentary end of an episode during which a layer of biogenic origin is exposed to high-pressure gases of hydrocarbonic composition. The regularity, thickness, large extent, and carbon enrichment of coal seems may thus find a natural explanation (cf. Fig. 3).

Another corollary of our thesis is that crystals which only form under high pressure, like diamonds, need not have formed at a depth h given by $h = p/\rho g$: they may well have formed nearer to the surface in a high-pressure diatreme and may have been convected upward and ejected during one of the subsequent eruptions. The isobars of a partially solid planet or moon can grossly deviate from equipotentials.

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