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1 Laboratory experiments on rain-driven convection:  
2 implications for planetary dynamos

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5 **Abstract**

6 Compositional convection driven by precipitating solids or immiscible liquids has been  
7 invoked as a dynamo mechanism in planets and satellites throughout the solar system,  
8 including Mercury, Ganymede, and the Earth. Here we report laboratory experiments  
9 on turbulent rain-driven convection, analogs for the flows generated by precipitation  
10 within planetary fluid interiors. We subject a two-layer fluid to a uniform intensity  
11 rainfall, in which the rain is immiscible in the upper layer and miscible in the lower  
12 layer. Rain falls through the upper layer and accumulates as a two-fluid emulsion in the  
13 interfacial region between the layers. In experiments where the rain is denser than the  
14 lower fluid, rain-injected vortices evolve into small-scale plumes that rapidly coalesce  
15 into larger structures, resulting in turbulent convection throughout the lower layer.  
16 The turbulent convective velocity in our experiments increases approximately as the  
17 cube root of the rain buoyancy flux, implying little or no dependence on viscous and  
18 chemical diffusivities. Applying diffusion-free scaling laws for magnetic field generation,  
19 we find that precipitation-driven convection can be an effective dynamo mechanism in  
20 planetary cores provided the precipitation buoyancy flux is large and the convecting  
21 region is deep and nearly adiabatic.

22 **Key words:** Rain-driven convection, planetary dynamos, iron snow, magnesium precip-  
23 itation, convection experiments, scaling laws

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# 1 Introduction

Multi-phase flows involving liquids plus solids or several immiscible liquids have been proposed as power sources for a number of planetary dynamos, both large and small (Breuer et al., 2015). Examples include iron-snow precipitation in the iron alloy cores of Mercury (Villim et al., 2010; Dumberry and Roldovini, 2015) and Ganymede (Hauck et al., 2006; Rückriemen et al., 2015; Christensen, 2015), in terrestrial exoplanets (Gaidos et al., 2010), precipitation of low-density constituents near the top of Earth’s iron-rich core for the early geodynamo (Buffett et al., 2000; O’Rourke and Stevenson, 2016; Badro et al., 2016), and helium rain in Saturn and other hydrogen-rich giant planets (Stevenson, 1980; Fortney and Hubbard, 2004).

Although the compositions of both the precipitate and the core fluid, and the ways envisioned to generate fluid motions differ among the planets and satellites, the underlying mechanics are fundamentally similar in each case, as Figure 1 illustrates. First, cooling of the planet leads to saturation of one or more components of the conducting fluid. Nucleation of that component produces liquid drops or solid grains, which precipitate as the equivalents of rain, sleet, graupel, or snow.

In Mercury and Ganymede, the usual assumption is that dense iron-snow precipitates downward through the upper, cooler portions of their molten iron alloy cores (Dumberry and Roldovini, 2015; Christensen, 2015). Falling into the deeper and warmer portions of the core, the iron-snow melts, increasing the density and destabilizing the fluid there. The core then consists of a two-layer system, with a precipitation-dominated region overlying a convection-dominated region, shown in Figure 1a. More complex layering is possible, including staircase structures, particularly if the melting curve and the temperature profile are irregular (Vilim

48 et al., 2010; Dumberry and Roldovini, 2015), or if double diffusive process occur. In general,  
49 however, it is usually concluded that the precipitating regions are stable and that dynamo  
50 action is concentrated in the convective region (Christensen and Wicht, 2008). The process  
51 is similar to convective turbulence in Earth’s troposphere, when precipitation falling from  
52 clouds as virga snow sublimates before reaching the ground, cooling and destabilizing the  
53 air below the cloud base (Kudo, 2013).

54 A related but somewhat different precipitation scenario has been proposed for the Earth’s  
55 core in the deep past. Evidence for a geomagnetic field of similar strength as the present-  
56 day field extends to 3.4 Ga (Tarduno et al. 2010) and possibly 4.2 Ga (Tarduno et al.,  
57 2015), whereas current estimates place the age of the inner core at 1 Ga or less (Olson  
58 et al., 2015), highlighting the need for geodynamo energy sources other than inner core  
59 growth. It has been hypothesized that, as the core cooled from an initially high temperature  
60 state, nucleation of weakly soluble magnesium-bearing grains or drops occurred (O’Rourke  
61 and Stevenson, 2016; Badro et al., 2016). Positively buoyant in Earth’s iron-rich core, the  
62 magnesium-bearing compounds precipitated upward and accumulated at the core-mantle  
63 boundary (CMB), as shown schematically in Figure 1b. Removal of magnesium left the  
64 residual core fluid denser and therefore unstable, enabling a compositional convective region  
65 to develop, thereby helping to maintain the early geodynamo.

66 Another example of precipitation-driven flow has been proposed for gas giant planets,  
67 Saturn in particular (Stevenson, 1980), but also giant exoplanets (Fortney and Hubbard,  
68 2004). There, cooling produces supersaturated conditions for helium in the outer portion of  
69 the hydrogen-rich fluid envelope of the planet, a situation broadly similar to Figure 1a. It is  
70 hypothesized that helium rainfall stabilizes the density profile in a precipitation-dominated  
71 layer, allowing strong horizontal shear flows to develop. It has been shown that dynamo

72 action driven by strong shear constrains the external magnetic field to be highly axisymmetric  
73 (Stanley, 2010; Cao et al., 2012), as observed on Saturn (Cao et al., 2011).

74 Thermochemical evolution calculations reveal that the rate of gravitational potential  
75 energy release in the saturation-precipitation process is proportional to the cooling rate of  
76 the fluid (Rückriemen et al., 2015). For plausible planetary cooling rates, the precipitation  
77 mechanisms described above are expected to release substantial amounts of gravitational  
78 potential energy per unit time, and therefore hold potential for dynamo action. Yet, little  
79 is known about the multi-phase flows involved in precipitation-driven convection, much less  
80 their ability to produce efficient planetary dynamos. In particular, there are questions about  
81 whether the potential energy released by precipitation converts efficiently to kinetic energy  
82 of fluid motion, and whether the kinetic energy is produced at scales that are large enough  
83 for dynamo action.

84 Precipitation may also play an important role during late-stage planetary accretion. It  
85 has been hypothesized that late-stage giant impacts fragment core-forming metals in a deep  
86 magma ocean (Tonks and Melosh, 1993; Nakajima and Stevenson, 2015) that might extend  
87 to the core (Labrosse et al., 2007). For some impacts, complete fragmentation could produce  
88 iron rainfall from the mantle directly into the core (Ichikawa et al., 2010; Deguen et al.,  
89 2014; Kraus et al., 2015). In this scenario, high-pressure, high-temperature metal-silicate  
90 interactions predict that the metallic rain absorbs large concentrations of lighter elements  
91 from the magma (Takafuji et al., 2005; Siebert et al., 2011) and would enter the core with a  
92 large density deficit, rather than a density excess, possibly contributing to the stable layering  
93 inferred at the top of the present-day outer core (Helffrich and Kaneshima, 2010; Landeau  
94 et al., 2016).

95 In this study we investigate these issues using analog laboratory experiments of rain-  
96 driven convection. We exploit differences in interfacial tension to create a transition from  
97 a precipitation-dominated region to a convection-dominated region in a two-layer fluid with  
98 geometry shown in Figure 1c. In both the laboratory and in nature, precipitation-driven  
99 convection is intrinsically non-uniform on multiple scales. On the smallest scales, it is gran-  
100 ular in space and time because it originates from the dissolution of individual drops or  
101 particles. It is also heterogeneous at intermediate time and length scales, due to variability  
102 in the local precipitation rate, which in our experiments comes from random fluctuations  
103 in the rain production apparatus in Figure 1c, and in natural systems comes from lateral  
104 heterogeneity in temperature, composition, and the velocity of the fluid through which the  
105 precipitation falls. A fundamental assumption is that these small and intermediate scale  
106 heterogeneities average out, so that on larger scales the precipitation induces a constant and  
107 horizontally uniform buoyancy flux, analogous to turbulent Rayleigh-Benard convection with  
108 fixed heat flux boundary conditions (Verzicco and Sreenivasan, 2008; Johnston and Doering,  
109 2009; Huang et al., 2015). However, because precipitation-driven convection has received so  
110 little attention, the validity of this assumption as applied to planetary dynamos is an open  
111 question.

## 112 **2 Rain-driven convection experiments**

113 Figures 2-5 show results of laboratory experiments in which a two-layer fluid is subject to a  
114 uniform intensity rainfall of a third fluid. The rain is a dyed aqueous solution, immiscible in  
115 the upper fluid, a low-density, low-viscosity silicone oil, but miscible in the lower fluid, pure  
116 water in these experiments. The two fluids are confined in a rectangular plexiglass tank with  
117 a square 25x25 cm cross-sectional area shown in Figure 1c, above which a 25 cm diameter

118 circular shower head consisting of 100 equally-spaced, 0.75 mm diameter spigots connected  
119 via a flow meter and control valve to a head tank containing the rain fluid. Back lighting  
120 with a rectangular diode array is used to illuminate the tank. The stochastic drop pattern  
121 from the shower head yields a statistically uniform precipitation in the central region of the  
122 tank, where we confine our measurements, and slightly reduced precipitation in the corner  
123 regions.

124 We consider two types of rain, distinguished by their density relative to the lower fluid.  
125 In high-density rain experiments, hereafter referred to as Type 1 experiments, the rain fluid  
126 consists of a water-NaCl solution with organic dye, the salt and dye concentrations adjusted  
127 to produce the desired density excess with respect to the lower fluid. In low-density rain  
128 experiments, referred to as Type 2, the rain fluid consists of a water-ethanol solution with  
129 organic dye, their concentrations adjusted to produce the desired density deficit with respect  
130 to the lower fluid. Properties of the experimental fluids and a comparison of the dimensionless  
131 parameters in the experiments and in planetary cores are given in Table 1.

132 In the high-density Type 1 rain experiment shown in close-up in Figures 2a and 2b,  
133 the rain has an initial density excess of  $\Delta\rho/\rho_L=+1.0\%$  with respect to the lower fluid, the  
134 average rainfall intensity is  $R=4.5\times 10^{-4}$  m s<sup>-1</sup> (approximately 28 ml s<sup>-1</sup> total rain rate)  
135 applied over approximately 120 s. The corresponding rain buoyancy flux, defined as (see  
136 Appendix)

$$F = \frac{g\Delta\rho}{\rho_L}R \quad (1)$$

137 where  $g$  is gravity, is  $F = 4.5 \times 10^{-5}$  m<sup>2</sup> s<sup>-3</sup> in this experiment. In the low-density Type  
138 2 rain experiment shown in close-up in Figures 2c and 2d, the rain density deficit is -1.0%  
139 and the average rainfall intensity is the same as in the accompanying Type 1 experiment, so  
140 that the rain buoyancy flux in that case is  $F = -4.5 \times 10^{-5}$  m<sup>2</sup> s<sup>-3</sup>.

141 In both Type 1 and 2 experiments, rain precipitates through the upper layer and accumu-  
142 lating as a binary emulsion at the interface with the lower fluid. The drops fall along nearly  
143 vertical trajectories in the upper layer. We do not observe substantial lateral deflections or  
144 clustering of the drops as they fall, even in the highest rain intensity experiments, as would  
145 be the case if the rainfall generated coherent large-scale motions in the upper layer. We  
146 emphasize this point because it relates to the question of the capacity for dynamo action in  
147 precipitation-dominated regions of planetary cores.

148 Gravitational separation of the fallen rain and the entrained upper layer fluid occurs  
149 within the two-phase emulsion zone shown in Figure 2. The various steps in the gravita-  
150 tional separation processes within the emulsion are the same as described in detail by Sato  
151 and Sumita (2007) and Gilet et al., (2007) for the case in which the negatively buoyant com-  
152 ponent is discontinuous (localized in drops) and the neutral component forms the connected  
153 pore network. Drop dissolution ejects small volumes of the rain fluid downward, forming  
154 millimeter-sized ring vortices below the emulsion interface, while the less dense silicone oil  
155 percolates back into the upper layer.

156 The behavior of the rain fluid injected into the lower layer depends on its density relative  
157 to the lower fluid. In the case of excess density rain in Type 1 experiments, the vortex  
158 rings formed by drop dissolution evolve into small-scale, negatively buoyant compositional  
159 plumes, as shown in Figure 2a. The time sequence in Figures 2a and 2b shows how these  
160 small-scale compositional plumes coalesce into larger plumes that drive a still larger-scale  
161 turbulent convective overturn, mixing and homogenizing the entire lower fluid. In contrast,  
162 in Type 2 experiments with density-deficient rain, the small-scale compositional plumes are  
163 positively buoyant and barely penetrate into the lower fluid, as shown in Figures 2c and 2d.  
164 Instead of producing turbulence throughout the lower layer, the rain products accumulate



165 below the emulsion interface, producing a stratified rain product layer that grows with time.

166 The intensity of convection in the lower layer is governed by a flux Rayleigh number  
167 defined in terms of  $F$  (see Appendix)

$$Ra_F = \frac{F d_L^4}{\nu^3} \quad (2)$$

168 where  $d_L$  and  $\nu$  are the initial lower fluid depth and kinematic viscosity, respectively. For  
169 the Type 1 experiment shown in Figures 2a and 2b, using the lower fluid depth  $d_L$  given in  
170 Table 1, the flux Rayleigh number is  $Ra_F = 4.5 \times 10^{10}$ . We measure the intensity of the  
171 convective turbulence in terms of a Reynolds number based on the lower fluid depth and the  
172 average velocity of the plumes  $w$  (see Appendix)

$$Re = \frac{w d_L}{\nu}, \quad (3)$$

173 The average plume velocity is defined as  $w = d_L/\tau$ , where  $\tau$  is the plume travel time, the  
174 average of the measured time intervals between the formation of plumes below the interfacial  
175 emulsion and the arrival of those plumes at the base of the tank. For the Type 1 experiments  
176 shown in Figures 2a and 2b,  $Re \simeq 1.8 \times 10^3$ .

177 The image sequence in Figure 3 shows large-scale views of the development of the con-  
178 vecting region in an 0.1% excess density (Type 1) rain experiment at  $Ra_F = 4.5 \times 10^9$ . The  
179 emergence of large-scale flows through plume growth and coalescence is evident in this se-  
180 quence. At 9 s after rain injection begins (Figure 3a) the small size of the plumes corresponds  
181 to the scale at which vorticity is injected into the lower layer following the coalescence of  
182 individual rain drops. At 24 s after the initial injection (Figure 3b), clusters of the small  
183 plumes have merged to form larger plumes. Plume mergers continue as the mixing layer and  
184 the large-scale circulation approach the base of the lower layer at 39 s (Figure 3c). The final

185 image Figure 3d shows nearly complete homogenization on the lower layer by the overturn-  
 186 ing circulation. The progressive upward displacement of the emulsion at the rain-convection  
 187 interface in Figures 3a-d is due to filling of the lower layer by the rain fluid.

188 The two images in Figure 4 compare Type 1 rain-driven turbulent convection with large  
 189 and small density excess, respectively. In Figure 4a, the rain density excess is 0.03% and the  
 190 flux Rayleigh number is  $Ra_F = 1.35 \times 10^9$ , whereas in Figure 4b the rain density excess is  
 191 2.5% and the flux Rayleigh number is approximately  $Ra_F = 1.13 \times 10^{11}$ , nearly two orders  
 192 of magnitude larger. Note the increase in turbulent mixing from the higher density rainfall  
 193 and the shorter growth time of the plumes in Figure 4b compared to Figure 4a. But in terms  
 194 of their large-scale structure, the two flows are remarkably similar, despite the nearly two  
 195 orders of magnitude difference in  $Ra_F$ .

196 Figure 5 shows the relationship between the plume Reynolds number defined in (3) and  
 197 the flux Rayleigh number, from five different Type 1 rain convection experiments. Figure 5  
 198 also shows a power-law fit to the data of the form

$$Re = cRa_F^\beta \quad (4)$$

199 in which  $c = 0.17 \pm 0.02$  and  $\beta = 0.32 \pm 0.02$ . Within experimental error, (4) is indis-  
 200 tinguishable from the  $\beta=1/3$ -power law exponent that corresponds to ideal inertial scaling  
 201 (see Appendix). In terms of dimensional variables, a 1/3-power law relationship between  
 202  $Re$  and  $Ra_F$  implies  $w \propto (Fd_L)^{1/3}$ , i.e., the plume velocity is independent of the viscosity  
 203 and the diffusivity, and depends only on the buoyancy flux and the convective layer depth.  
 204 In our experiments, we mostly varied the buoyancy flux, by independently varying the rain  
 205 density and the precipitation rate. In contrast, the lower layer depth was varied only by  
 206 small amounts, so the implied  $d_L^{1/3}$ -dependence of  $w$  was not actually tested.

207 We can compare our plume Reynolds number scaling with the plume Reynolds numbers  
 208 measured in thermal convection experiments of the classical Rayleigh-Benard type, in which  
 209 the top and bottom boundaries are rigid and isothermal. Figure 5 shows a semi-theoretic  
 210 scaling law for the plume Reynolds number derived by Grossmann and Lohse (2001), which  
 211 has the form

$$Ra_F = (Nu - 1)Pr^{-2}Ra = c_1 \frac{Re^2}{f(Re)} + c_2 Re^3, \quad (5)$$

212 in which  $Ra$  and  $Nu$  are the Rayleigh number and the Nusselt number in Rayleigh-Benard  
 213 convection defined in terms of the temperature difference across the fluid (defined in the  
 214 Appendix),  $Pr$  is the Prandtl number,  $f(x) = (1 + x^2)^{1/4}/x^{1/2}$ , and  $c_1$  and  $c_2$  are empirical  
 215 coefficients that depend on the experimental geometry and boundary conditions (Grossmann  
 216 and Lohse, 2003), and also on the technique used to measure the plume velocities (Ahlers et  
 217 al., 2009).

218 Fitting (5) to turbulent Rayleigh-Benard convection experiments in a variety of fluids,  
 219 Grossmann and Lohse (2002) obtain  $c_1 \simeq 8.7$  and  $c_2 \simeq 1.45$ . The resulting scaling law is  
 220 indicated by the dashed line in Figure 5 . A similar scaling has been found to hold for  
 221 Rayleigh-Benard convection in cylindrical domains, but with some variations in the coeffi-  
 222 cients (Ahlers and Xu, 2001; Funfschilling et al., 2005). In Figure 5 we indicate the sensitivity  
 223 to geometry and boundary conditions with the background gray shading. We note that the  
 224 trend of our data is hardly distinguishable from (5), and the offset is only about 40%, an  
 225 amount which can be expected in light of the differences in boundary conditions, geometry,  
 226 and the fact that we measure the overall plume travel time in our rain convection experi-  
 227 ments versus mid-depth plume travel times in the conventional Rayleigh-Benard convection  
 228 experiments.

### 3 Implications for precipitation-driven dynamos

We quantify the dynamo capacity of precipitation-driven convection in planetary core using scaling laws for the fluid velocity and the magnetic field intensity generated by thermochemical convection obtained from numerical dynamo systematics (Christensen and Aubert, 2006; Aubert et al. 2009). These scaling laws include the effects of sphericity, self-gravitation, and planetary rotation, which are missing from our experiments, and assume that the fluid velocity and the magnetic induction are independent of diffusive effects, including the viscosity. The fact that our experimental convective velocities are consistent with inertial (i.e., viscosity-independent) scaling provides some justification for using such diffusion-free dynamo scaling laws in this application.

We assume that the planet's core is divided into an outer precipitation-dominated region and a central convection-dominated region, separated at a radius  $r_c$ . We can write the precipitation buoyancy flux at the top of the convective region as (1), where now  $g$  is gravity at radius  $r_c$ ,  $\rho_L$  is the convecting region density,  $\Delta\rho$  is the density difference between the convecting region and the precipitation, and  $R$  is the rate of precipitation, solid or liquid, evaluated at the top of the convecting region. Magnetic field generation depends on  $F$  through the magnetic Reynolds number of the convection  $Rm$ . In an entirely molten convecting sphere with radius  $r_c$ ,

$$Rm = \frac{ur_c}{\eta} \quad (6)$$

where  $\eta$  is the magnetic diffusivity of the fluid and  $u$  is the rms fluid velocity. For full-sphere geometries and assuming an adiabatic heat flux at the CMB, dynamo scaling relationships (Aubert et al., 2009) yield

$$Rm \simeq 1.3p^{0.42}PmE^{-1} \quad (7)$$

250 where  $Pm = \nu/\eta$  is the magnetic Prandtl number,  $E = \nu/\Omega r_c^2$  is the Ekman number,

$$p = \frac{3F}{5\Omega^3 r_c^2} \quad (8)$$

251 is the convective power, and  $\Omega$  is angular velocity of planetary rotation. The critical magnetic  
 252 Reynolds number for convective dynamo onset is  $Rm_{crit} \simeq 40$  (Christensen et al., 1999).

253 We first consider magnesium precipitation in Earth’s core prior to inner core nucleation,  
 254 recently proposed by O’Rourke and Stevenson (2016) and later considered by Badro et al.  
 255 (2016). Rising precipitation in the form of buoyant magnesium oxides and silicates would  
 256 leave behind a dense iron-rich residual fluid, promoting compositional convection deeper in  
 257 the core. If we assume that the background state of the core is well mixed (i.e., adiabatic),  
 258 then very low intensity precipitation of this type could initiate the geodynamo. According  
 259 to (1), (7), (8) and the data in Table 2, a precipitation rate of just a few  $\text{m Gyr}^{-1}$  produces  
 260  $Rm_{crit}$  under these conditions.

261 More intense magnesium precipitation is needed to maintain a finite intensity magnetic  
 262 field, even if the background state of the core is well mixed. According to dynamo model  
 263 systematics (Olson and Christensen, 2006) the dipole magnetic moment  $\mathcal{M}$  in the dipole-  
 264 dominated regime depends on the buoyancy flux from precipitation  $F$  approximately as

$$\mathcal{M} \simeq 1.8 \left( \frac{\rho}{\mu_0} \right)^{1/2} r_c^{10/3} F^{1/3}, \quad (9)$$

265 where  $\mu_0$  is free-space magnetic permeability. In order to sustain a magnetic field with dipole  
 266 moment  $\mathcal{M}=50 \text{ ZAm}^2$ , comparable to the time-averaged Proterozoic geomagnetic field inten-  
 267 sity (Biggin et al., 2015), (9) along with the data in Table 2 predict that a precipitation rate  
 268 of  $R \simeq 400 \text{ m Gyr}^{-1}$  is required. This is a negligibly small magnesium addition to the mantle  
 269 and would not materially affect geochemical tracers such as Hf-W isotopic ratios (Rudge et  
 270 al., 2010) or mantle siderophiles (Chabot et al., 2005). It would slightly increase Earth’s

271 rotation rate, but by less than 1%, far too small to counteract the angular deceleration from  
 272 tidal braking (Denis et al., 2011).

273 Far more intense precipitation would be needed for dynamo action if the core was strat-  
 274 ified, either thermally or by composition. If the heat loss from the convective region is less  
 275 than the heat conducted down the core adiabat, a plausible situation in view of the high  
 276 thermal conductivity of the core (Gomi et al., 2013; Labrosse, 2015), then an additional  
 277 compositional buoyancy flux must be added to balance the negative thermal buoyancy flux  
 278 due to the subadiabatic thermal gradient, in order to homogenize the core fluid. In terms  
 279 of the total heat loss from the convecting region  $Q$ , the additional thermal buoyancy flux  
 280 is given by  $F_T = (Q - Q_{ad})/\rho A_c H_c$ , where  $Q_{ad}$  is the adiabatic heat loss,  $A_c = 4\pi r_c^2$  is the  
 281 surface area of the convecting region, and  $H_c$  is its temperature scale height. The net buoy-  
 282 ancy flux that enters into the expression for the convective power (8) then becomes  $F + F_T$ .  
 283 If the thermal buoyancy flux  $F_T$  is negative because  $Q < Q_{ad}$ , then a larger rain-produced  
 284 buoyancy flux  $F$  is required in order to compensate for the subadiabatic heat loss.

285 The effects of core heat flux and magnesium precipitation rate on the dynamo structure  
 286 and the induced dipole moment are shown in Figure 6, where we have assumed a convecting  
 287 zone radius of  $r_c=3480$  km, equal to the present-day core radius, and an adiabatic heat  
 288 flux of  $0.1 \text{ W m}^{-2}$ , equivalent to  $Q_{ad} \simeq 15 \text{ TW}$ . Line contours indicate the induced dipole  
 289 moment in  $\text{ZAm}^2$  from (9) and the local Rossby number  $Ro_\ell$ , which controls the transition  
 290 from dipolar to multipolar dynamo states (Christensen and Aubert, 2006; Aubert et al.,  
 291 2009):

$$Ro_\ell \simeq 0.54p^{0.48} E^{-0.32} (PrPm)^{0.19}, \quad (10)$$

292 where  $Pr = \nu/\kappa$  is the Prandtl number. Shadings indicate dynamo regimes that are subcrit-  
 293 ical (labeled no-dynamo), convection driven by precipitation against stable thermal strat-

294 ification (labeled precipitation dynamo), and convection driven by precipitation acting in  
295 concert with unstable thermal stratification (labeled thermo-precipitation dynamo).

296 Magnesium precipitation can maintain a strong dipolar dynamo even under subadiabatic  
297 conditions, provided the precipitation rate is high enough. Suppose, for example, that  $Q = 12$   
298 TW, equivalent to  $q_{cmb} \simeq 0.08 \text{ W m}^{-2}$  for the average local heat flux at the CMB. To maintain  
299 a dipole moment of  $50 \text{ ZAm}^2$  with this core heat flux, Figure 6 indicates that a magnesium  
300 rain rate of approximately  $R=2.8 \text{ km Gyr}^{-1}$  is required. For purposes of comparison, this  
301 is approximately the same as the present-day mean rate of sediment accumulation on the  
302 Pacific ocean floor (Olson et al., 2016), a modern analog that demonstrates how sparse this  
303 type of precipitation would be in the Earth’s core.

304 With these properties,  $Ro_\ell < 0.1$ , indicating the dynamo would be dipolar, consistent  
305 with most Proterozoic paleomagnetic data (Biggin et al., 2015). However, a precipitation  
306 dynamo with these parameters lies perilously close to the subcritical regime in Figure 6,  
307 such that a modest reduction in core heat flux would cause it to fail. The situation is made  
308 worse by the fact that the precipitation rate  $R$  is coupled to  $Q$  through the cooling rate of  
309 the core, so that a reduction in  $Q$  also reduces  $R$ , further increasing the chances of dynamo  
310 failure.

311 The results in Figure 6 assume the convective region occupies the entire Earth’s core.  
312 In other planets the convective region may occupy only a fraction of the core, and if so,  
313 the precipitation rate required for dynamo action can be very high. For example, Figure 7  
314 shows contours of magnetic Reynolds number from (7) and local Rossby number from (10)  
315 as functions of convecting radius and iron-snow precipitation rate for Mercury, based on the  
316 parameter values in Table 2 and assuming a solid inner core is absent or negligibly small.

317 If the convective radius exceeds 200 km, a minimum precipitation rate near  $30 \text{ km Gyr}^{-1}$   
318 is needed for dynamo onset, but in a smaller convective region, with  $r_c < 100 \text{ km}$ , dynamo  
319 action would require implausibly high precipitation rates.

320 Another implication of Figure 7 is that iron-snow precipitation can generate high mag-  
321 netic Reynolds number conditions in a deep convective region in Mercury, but according  
322 to (10), this convection would induce a multi-polar field because of the planet's slow ro-  
323 tation. This is seemingly in contradiction with the offset dipolar field observed in orbit  
324 around Mercury (Anderson et al., 2011). It is, however, consistent with a proposed deep dy-  
325 namo mechanism for Mercury by Christensen (2006), which postulates that the non-dipolar  
326 components of the dynamo magnetic field generated deep within the Mercury's core are elec-  
327 tromagnetically screened in an overlying stable layer. According to our experiments, such  
328 electromagnetic screening might occur in the precipitation-dominated layer because it lacks  
329 large-scale circulation.

330 Figure 8 shows the dynamo regimes for Ganymede, calculated according to the same  
331 scaling laws as for Earth and Mercury, but using the property values for Ganymede in Table  
332 2. The model for Ganymede differs from Mercury in having a smaller convective region and  
333 a lower thermal conductivity, following Christensen (2015). A convective dynamo driven  
334 by iron snow precipitation, possibly aided by an unstable thermal gradient, is predicted for  
335 most precipitation rates, including the nominal precipitation rate of  $10 \text{ km Gyr}^{-1}$ . The no-  
336 dynamo regime in Figure 8 consists of two part. In region (a) the convection is subcritical,  
337 i.e., the magnetic Reynolds number is below 40, whereas in the no-dynamo region (b) the  
338 heat loss is less than the latent heat released by melting the precipitation (assumed to be  
339  $5 \times 10^5 \text{ J kg}^{-1}$ ), so that steady-state convection is not sustainable.



## 340 4 Conclusions

341 In this paper we describe a new class of convective flows, driven by buoyancy acquired  
342 through precipitation. This form of convection has been proposed for evolving planetary  
343 cores, where cooling results in nucleation, precipitation, and dissolution of buoyant con-  
344 stituents. Our rain-driven experiments are analogous to precipitation-driven systems in  
345 planetary interiors in several respects, although there are important differences. Similarities  
346 include (1) the basic two-layer dynamical structure, with a precipitation region separated  
347 from a region where the precipitation is dissolved, with the possibility of compositional  
348 convection; (2) a thin multi-component transition zone separating these regions where the  
349 buoyancy for convection originates; and (3) buoyancy production on the scale of the indi-  
350 vidual precipitates, followed by plume coalescence. Significant differences include, for the  
351 experiments (1) lack of phase changes (solidification and melting) and latent heat exchange,  
352 which could contribute to the convection; (2) absence of rotational and magnetic field effects,  
353 which tend to stabilize the convection; and (3) the idealized plane-layer geometry, which ig-  
354 nores the changes in fluid volume and gravity with radius. In addition, there are structural  
355 differences between our experiments and some of the scenarios envisioned for planetary in-  
356 teriors.

357 In Earth's core, for example, it is supposed that extraction of upward precipitating mag-  
358 nesium or other light constituents would yield a negatively buoyant residual liquid, whereas  
359 in our experiments, the negatively buoyant liquid is produced through addition of downward  
360 precipitation. In cores dominated by the Fe-FeS system, the geometry might be inverse of  
361 our experiment. For example, for core compositions on the FeS side of the eutectic, convec-  
362 tion might be due to sulfur-rich fluid rising from a deep iron snow zone (Hauck et al. 2006;

363 Chen et al. 2008). In spite of these differences, we claim that the fundamental effects of  
364 precipitation on the dynamics of these system are captured in our experiments.

365 A key finding of our study is the process by which stochastic precipitation self-organizes  
366 to generate larger-scale convective motions. In our experiments, this organization begins  
367 during gravitational separation within multi-component emulsion layers, whereas in plan-  
368 etary interiors it is generally hypothesized to occur near the pressure-temperature horizon  
369 that corresponds to a phase transition. In both situations, however, the fluid buoyancy  
370 develops on the scale of the isolated particles or drops, generating micro-scale plumes that  
371 must coalesce in order to form the larger plume structures that induce convective overturn.

372 Although substantial gravitational potential energy may be released in the precipitation-  
373 dominated region of a planetary core, this does not guarantee that large scale magnetic  
374 induction occurs there. If the only fluid motions in the precipitation zones are in particle  
375 wakes, which have lateral dimensions on the scale of the particles themselves ( $\sim 1$  mm),  
376 these will be quickly dissipated by viscous or Lorentz forces, and will contribute little (if  
377 anything) to dynamo action. In contrast, our experiments show that the convective regions  
378 have far more capacity for dynamo action. In planetary cores, the rates of precipitation  
379 needed to maintain a dynamo are not excessive, especially if the precipitation is intense and  
380 strongly buoyant and the convective region is deep.

381 Lastly, we note that previous studies have found that diffusive boundary layer effects,  
382 including viscosity, remain controlling factors in heat transfer in Rayleigh-Benard convection  
383 when planetary rotation is included, even at high Rayleigh numbers (King et al., 2009,  
384 Stellmach et al., 2014). It would be revealing, therefore, to determine how the velocity  
385 scaling for precipitation-driven convection changes when rotation is added.

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## 394 Appendix

395 The steady-state mechanical energy equation for non-rotating, non-magnetic Boussinesq  
 396 convection can be written

$$\frac{1}{2}\nabla \cdot (u^2\mathbf{u}) = \mathbf{F} + \nu\mathbf{u} \cdot (\nabla^2\mathbf{u}) \quad (11)$$

397 where  $\rho$  is the mean density,  $\mathbf{u}$  is the fluid velocity,  $\nu$  is kinematic viscosity, and

$$F = \rho'(\mathbf{g} \cdot \mathbf{u})/\rho \quad (12)$$

398 is the buoyancy flux, in which  $\mathbf{g}$  is the vector gravity and  $\rho'$  is the density perturbation,  
 399 respectively. From left to right, the terms in (11) represent fluid inertia, production of  
 400 kinetic energy by buoyancy, and dissipation of kinetic energy by viscosity, respectively. At  
 401 convective onset, buoyancy production balances viscous dissipation. Using  $d$  and  $\nu/d$  to  
 402 scale length and velocity in the dissipation term, the ratio of the second to the third term  
 403 in (11) defines the Rayleigh number used in our experiments:

$$Ra_F = Fd^4/\nu^3 \quad (13)$$

404 Ideal inertial scaling, appropriate for fully-developed turbulent convection corresponds  
 405 to a balance between the inertia of the fluid and the buoyancy force driving the motion,  
 406 the first and second terms in (11). For this balance, by choosing  $w$  and  $d$  for the velocity  
 407 and length scales, (11) can be written in terms of non-dimensional variables (denoted by  
 408 asterisks) as

$$\frac{w^3}{2d}\nabla^* \cdot (u^{*2}\mathbf{u}^*) = F \quad (14)$$

409 where . In order for this force balance to hold, the convective velocity must scale like

$$w \sim (Fd)^{1/3}, \quad (15)$$

410 as given in the text.

411 Traditional Rayleigh-Benard thermal convection experiments use a different Rayleigh  
412 number, defined as

$$Ra = \alpha g \Delta T d^3 / \kappa \nu \quad (16)$$

413 where  $\alpha$  and  $\kappa$  are thermal expansion and thermal diffusivity, respectively, and  $\Delta T$  is the  
414 temperature difference across the fluid. Heat transfer in Rayleigh-Benard convection is  
415 usually measured in terms of the Nusselt number, defined as

$$Nu = qd/k\Delta T \quad (17)$$

416 where  $q$  and  $k$  are heat flux (conventionally measured on the fluid layer boundary) and  
417 thermal conductivity of the fluid, respectively. The definitions (16) and (17) are used in the  
418 r.h.s. of equation (5).

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Table 1: Rain-Driven Convection Experiment Properties

Property	Notation	Units	Experiment Type 1	Experiment Type 2
upper fluid density	$\rho_U$	$\text{kg m}^{-3}$	$0.82 \times 10^3$	$0.82 \times 10^3$
lower fluid density	$\rho_L$	$\text{kg m}^{-3}$	$1.0 \times 10^3$	$1.0 \times 10^3$
rain-lower fluid density difference	$\Delta\rho$	$\text{kg m}^{-3}$	0.3-25	-10
rain viscosity	$\nu_r$	$\text{m}^2 \text{s}^{-1}$	$1 \times 10^{-6}$	$1.6 \times 10^{-6}$
upper fluid viscosity	$\nu_U$	$\text{m}^2 \text{s}^{-1}$	$1.2 \times 10^{-6}$	$1.2 \times 10^{-6}$
lower fluid viscosity	$\nu_L$	$\text{m}^2 \text{s}^{-1}$	$1 \times 10^{-6}$	$1 \times 10^{-6}$
upper fluid depth	$d_U$	m	0.192	0.192
lower fluid depth, initial	$d_L$	m	0.178	0.178
rain interfacial tension	$\sigma_r$	$\text{N m s}^{-1}$	$2.2 - 2.4 \times 10^{-2}$	$2.6 \times 10^{-2}$
rain rate	$R$	$\text{m s}^{-1}$	$3.4 - 4.6 \times 10^{-4}$	$4.5 \times 10^{-4}$
rain buoyancy flux	$F$	$\text{m}^2 \text{s}^{-3}$	$1.35 - 110 \times 10^{-6}$	$-4.5 \times 10^{-5}$
rain drop radius	$r_r$	m	$\sim 10^{-3}$	$\sim 10^{-3}$
plume velocity	$w$	$\text{m s}^{-1}$	$3.4 - 14.3 \times 10^{-3}$	$< 4 \times 10^{-4}$
Dimensionless Parameter	Definition	Experiment Type 1	Experiment Type 2	Early Core (Estimated)
Bond Number, $B$	$g\Delta\rho r_r^2/\sigma$	0.012-0.018	0.015	$\sim 0.1$
Ohnesorge Number, $Oh$	$\rho_r\nu_r/\sqrt{\sigma r_r}$	$\sim 0.15$	$\sim 0.17$	$\sim 0.1$
Density Ratio, $\epsilon$	$\Delta\rho/\rho_L$	$0.3-25 \times 10^{-3}$	-0.01	$\sim 4.6$
Rayleigh Number, $Ra_F$	$Fd_L^4/\nu_L^3$	$1.35-112 \times 10^9$	$-4.5 \times 10^{10}$	$\sim 10^{29}$
Plume Reynolds Number, $Re$	$wd_L/\nu_L$	$6.05-25.5 \times 10^2$	$< 2 \times 10^1$	$\sim 2 \times 10^7$
Magnetic Reynolds Number, $Rm$	$ud_L/\eta$	na	na	$\sim 2 \times 10^3$

Table 2: Core Precipitation Properties (Nominal Values)

Notation	Definition, Units	Early Earth	Mercury	Ganymede
$r_c$	convection radius, km	3480 <sup>a</sup>	500 <sup>b</sup>	650 <sup>j</sup>
$\Omega$	rotation rate, $10^{-5}$ rad s <sup>-1</sup>	5 <sup>c</sup>	0.124 <sup>d</sup>	1.02
$\eta$	magnetic diffusivity, m <sup>2</sup> s <sup>-1</sup>	0.8 <sup>e</sup>	$\sim 1$	0.7 <sup>k</sup>
$\rho$	average density, $10^3$ kg m <sup>-3</sup>	11 <sup>a</sup>	7.7 <sup>b</sup>	<sup>j</sup>
$q_s$	adiabatic heat flux, mW m <sup>-2</sup>	100 <sup>f</sup>	10 <sup>b</sup>	4 <sup>j</sup>
$q$	CMB heat flux, mW m <sup>-2</sup>	50-100 <sup>f</sup>	3-10 <sup>b</sup>	2-4 <sup>l</sup>
$H_c$	temperature scale height, km	6500 <sup>g</sup>	1400 <sup>b</sup>	1100 <sup>b</sup>
$g\Delta\rho/\rho$	precipitation buoyancy, m s <sup>-2</sup>	4.6 <sup>h</sup>	0.4 <sup>e</sup>	0.4 <sup>e</sup>
$R$	precipitation rate, km Gyr <sup>-1</sup>	2.5	50	10
$F$	precipitation buoyancy flux, $10^{-13}$ m <sup>2</sup> s <sup>-3</sup>	6	4	1.3
$\nu$	kinematic viscosity, $10^{-6}$ m <sup>2</sup> s <sup>-1</sup>	1	1	1 <sup>i</sup>
$k$	thermal conductivity, W m <sup>-1</sup> k <sup>-1</sup>	130 <sup>f</sup>	120 <sup>k</sup>	30 <sup>j</sup>

<sup>a</sup> PREM Dziewonski and Anderson (1981); <sup>b</sup> Dumberry and Rivoldini (2015); <sup>c</sup> Williams (2000); <sup>d</sup> Ward et al. (1976); <sup>e</sup> Hirose et al. (2013); <sup>f</sup> Labrosse (2015); <sup>g</sup> Vocadlo et al. (2003); <sup>h</sup> PREM CMB buoyancy contrast; <sup>i</sup> Perrialt et al. (2010); <sup>j</sup> Christensen (2015) <sup>k</sup> Deng et al. (2013) <sup>l</sup> Hauk et al. (2006) <sup>m</sup> Chen et al. (2008)

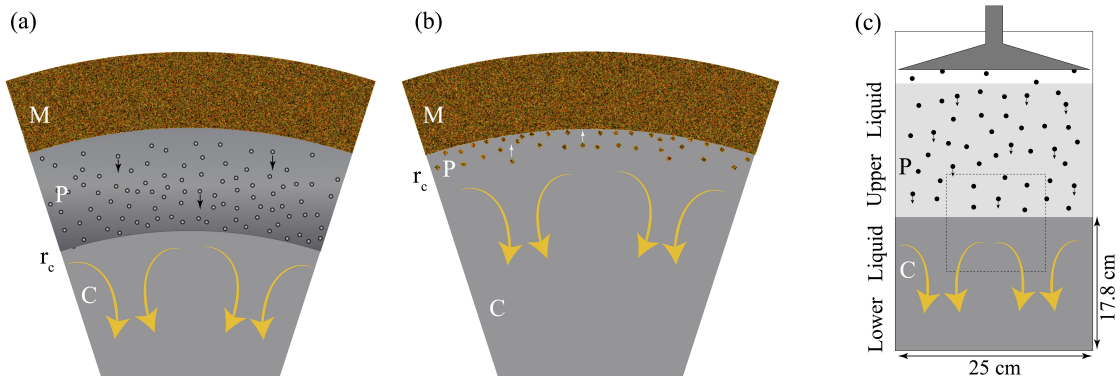


Figure 1: **Precipitation-Driven Convection.** Schematics of precipitation-driven convection in cooling planetary cores and laboratory analog. (a) Iron-snow; (b) Magnesium precipitation; (c) Laboratory apparatus, with the dashed box indicating the region imaged in Figure 2. M,P,C denote the mantle, the precipitation-dominated region, and the convection-dominated region, respectively.  $r_c$  denotes convective region radius. Small arrows indicate precipitation directions; large arrows denote convection.

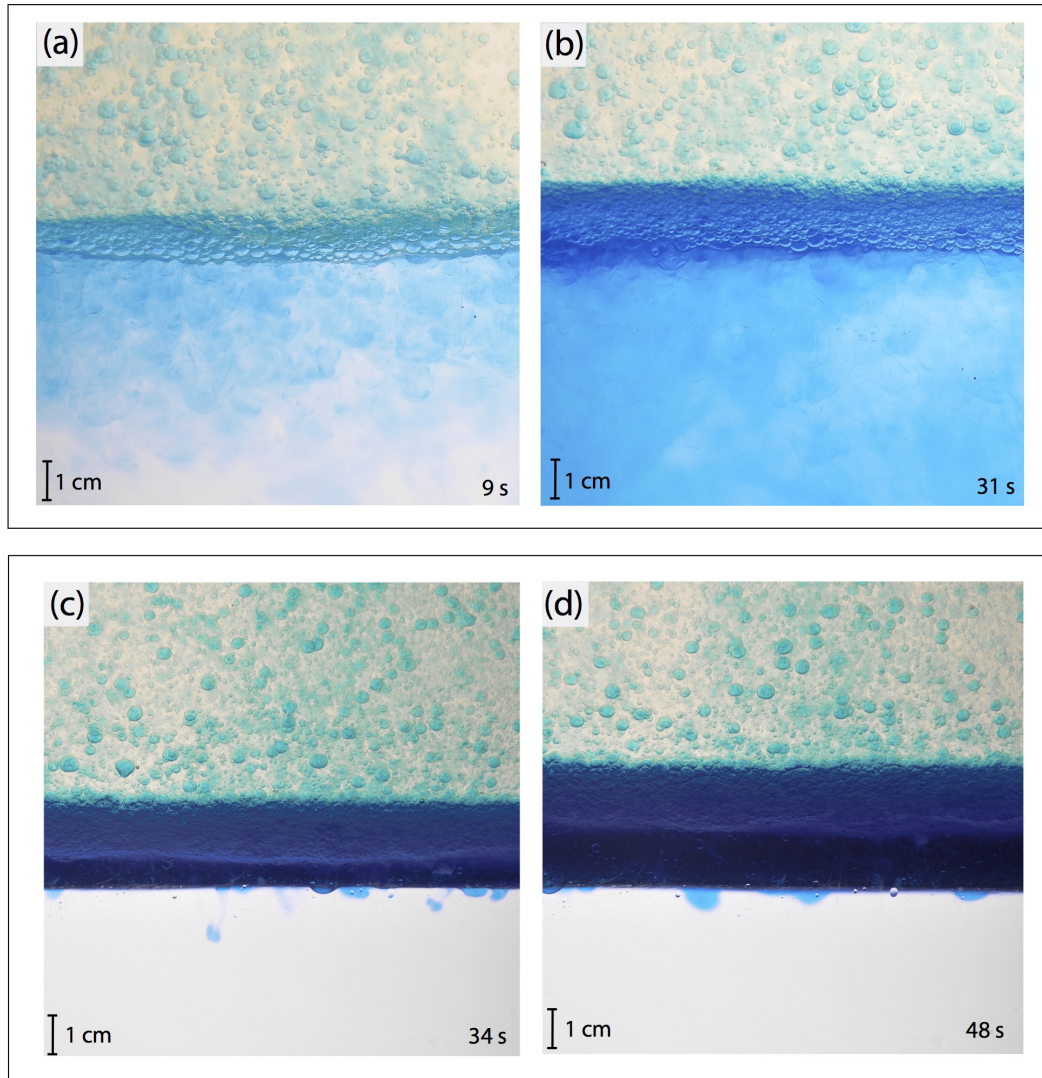


Figure 2: **Rain-Driven Convection Experiments.** In each case, raindrops (an aqueous solution) released at a constant and spatially uniform rate, fall through the upper liquid (silicone oil) into the lower liquid (another aqueous solution). In images (a) and (b) the blue-dyed rain is initially 1% denser than the lower liquid (Type 1 experiments). This close-up sequence, corresponding to the dashed box in Figure 1 with time given in seconds after convective onset, shows the development of a two-phase emulsion layer, phase separation producing compositional plumes and turbulent convection, resulting in homogenization of the lower liquid. In images (c) and (d) the rain is initially 1% less dense than the lower liquid (Type 2 experiments). In this case the rain products accumulate in the interfacial region and stratify the lower liquid. Experimental parameters are given in Table 1.



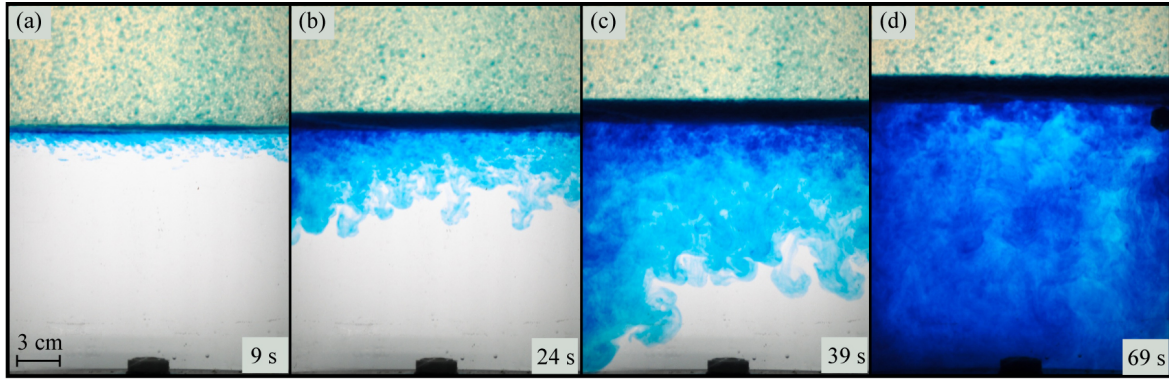


Figure 3: **Development of a Rain-Driven Convection Region.** This sequence shows the development of rain-driven convection in a Type 1 experiment with 0.1% excess rain density. (a): Onset of small scale compositional plumes; (b): Growth and turbulent merging of compositional plumes; (c): Large-scale convection in the lower layer; (d) mixing and homogenization. Time is given in seconds after rain injection begins.

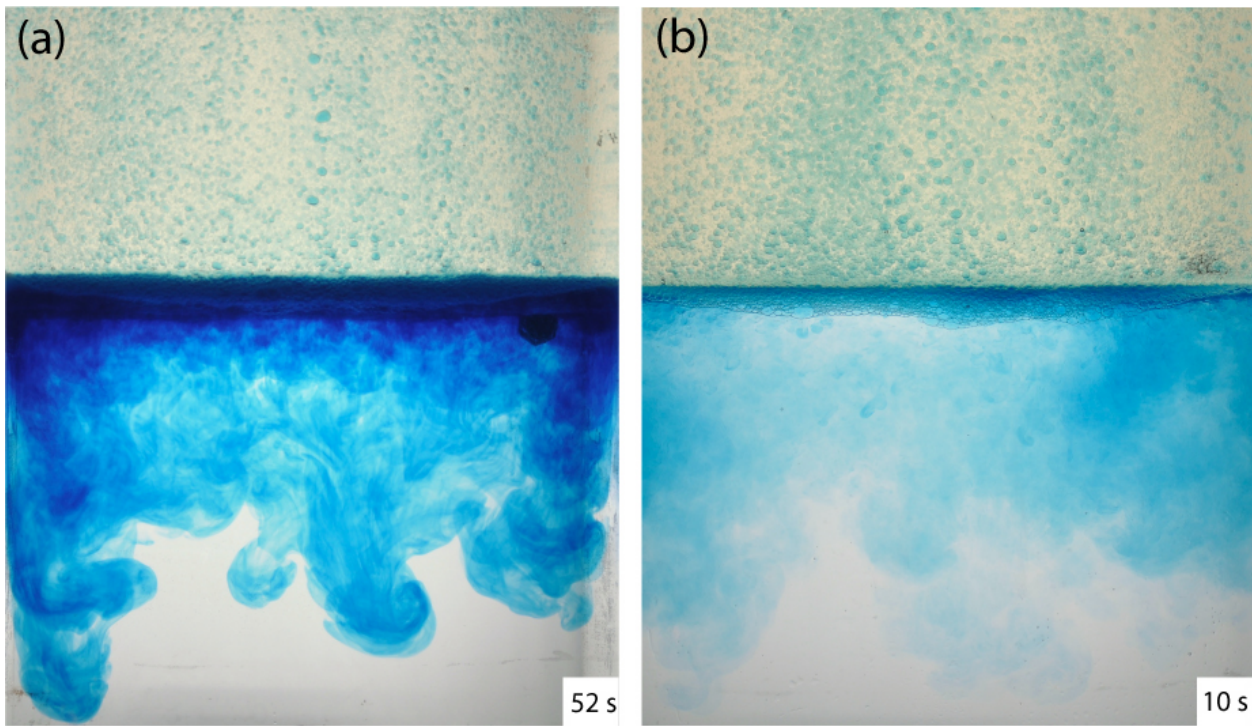


Figure 4: **Density Effects in Rain-Driven Convection.** Comparison of rain-driven turbulent convection in Type 1 experiments with large and small density contrasts. (a): Convection with initial rain excess density of 0.03%; (b): Convection with initial rain excess density of 2.5%. Time is given in seconds after convective onset.

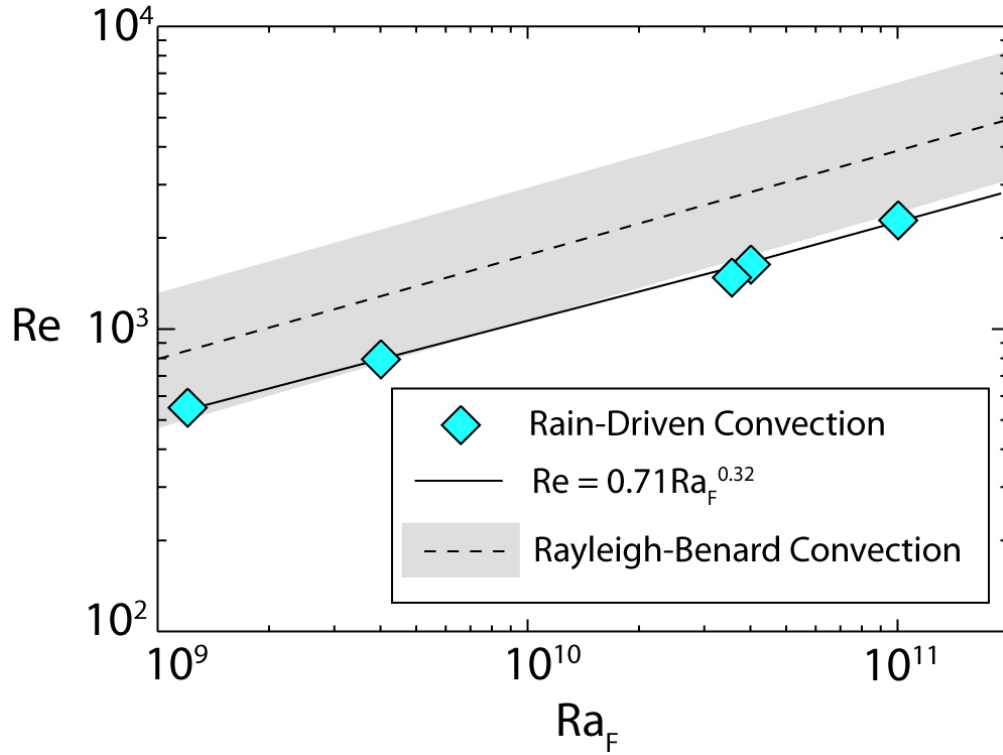


Figure 5: **Rayleigh Number-Reynolds Number Relation in Rain-Driven Convection** Reynolds number  $Re$  based on plume descent speed versus flux Rayleigh number  $Ra_F$  from five Type 1 (dense rain) experiments, in which the Rayleigh number increases with increasing excess density of the rainfall. Symbol size is representative of 1 s.d. uncertainties in  $Re$  and  $Ra_F$ . Power law fit exponent and coefficient are given in the plot. Dashed line is a fit by Grossmann and Lohse (2002) to plume velocities in Rayleigh-Benard experiments.

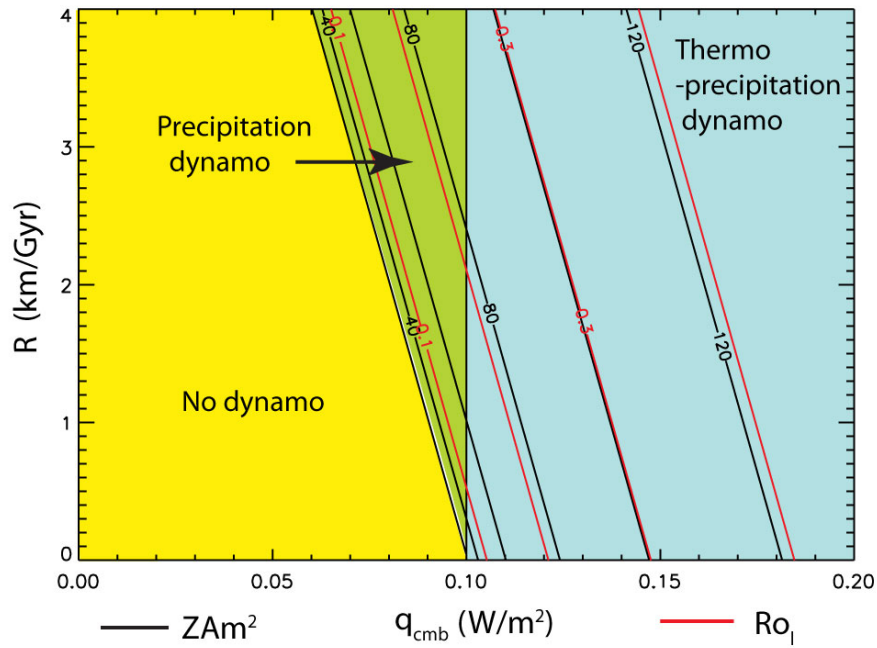


Figure 6: **Dynamo Regimes from Magnesium Precipitation in Earth’s Core** Dipole magnetic moment and local Rossby number for dynamo action in the totally fluid core (prior to inner core nucleation) driven by magnesium precipitation. Axes are core heat flux and precipitation rate. An adiabatic heat flux of  $100 \text{ mW m}^{-2}$  is assumed. Additional parameters are given in Table 2.

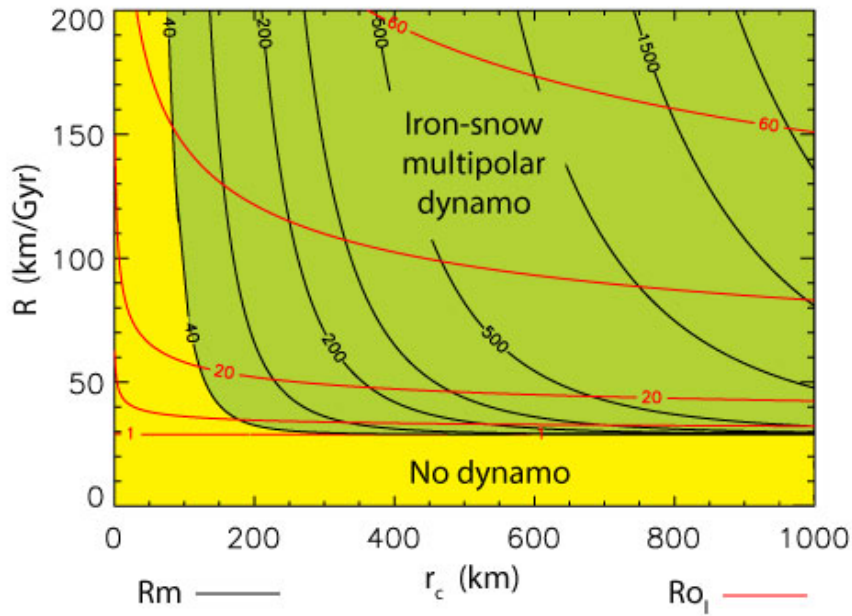


Figure 7: **Dynamo Regimes from Iron-snow Precipitation in Mercury** Magnetic Reynolds number  $Rm$  and local Rossby number  $Ro_\ell$  for dynamo action in the assumed totally fluid iron alloy core of Mercury, driven by iron-snow precipitation. Axes are iron-snow precipitation rate and convective radius. Total and adiabatic heat fluxes of  $5$  and  $50 \text{ mW m}^{-2}$  are assumed. Additional parameters are given in Table 2.

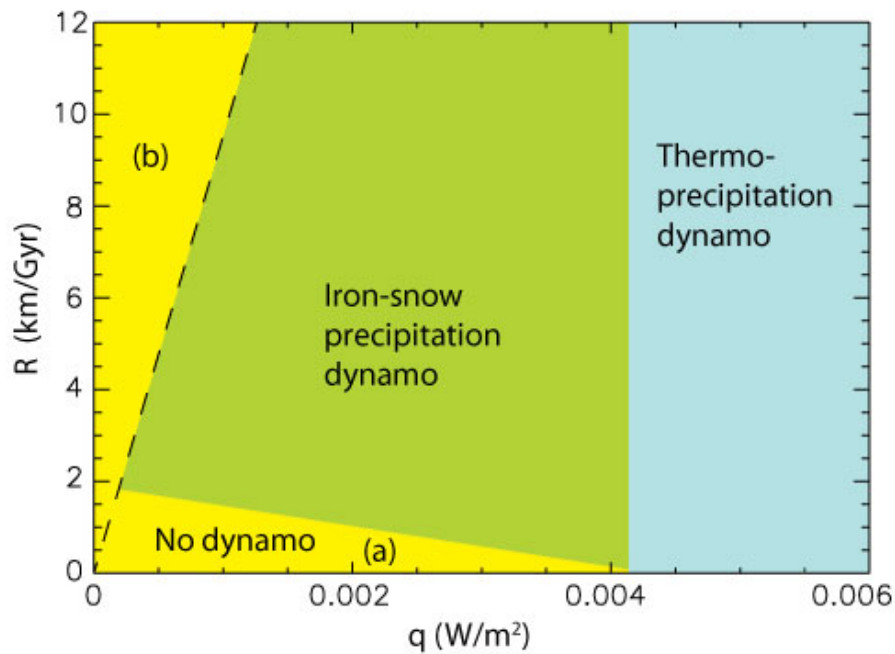


Figure 8: **Dynamo Regimes from Iron-snow Precipitation in Ganymede** Dynamo regimes in the assumed totally fluid iron alloy core of Ganymede, driven by iron-snow precipitation. Axes are heat flux at the top of the convecting region and precipitation rate. An adiabatic heat flux of  $5 \text{ mW m}^{-2}$  is assumed. No-dynamo region (a) has subcritical magnetic Reynolds number; region (b) has subcritical heat flux. Additional parameters are given in Table 2.