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

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Abstract

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

Key words: Rain-driven convection, planetary dynamos, iron snow, magnesium precipitation, 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

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

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

Another example of precipitation-driven flow has been proposed for gas giant planets, Saturn in particular (Stevenson, 1980), but also giant exoplanets (Fortney and Hubbard, 2004). There, cooling produces supersaturated conditions for helium in the outer portion of the hydrogen-rich fluid envelope of the planet, a situation broadly similar to Figure 1a. It is hypothesized that helium rainfall stabilizes the density profile in a precipitation-dominated 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-Bénard 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, into which a 25 cm diameter

118 circular shower head is fitted, consisting of 100 equally-spaced, 0.75 mm diameter spigots
 119 connected via a flow meter and control valve to a head tank containing the rain fluid. With
 120 this setup, the corner regions of the tank receive less direct precipitation compared to the
 121 central region. In our experiments, we concentrate on the action in the central region, where
 122 the rain is nearly uniform. Back lighting with a rectangular diode array is used to illuminate
 123 the tank.

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 $R=4.5\times10^{-4}$ m s⁻¹ (approximately 28 ml s⁻¹ 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 $F = 4.5 \times 10^{-5}$ m² s⁻³ 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 $F = -4.5 \times 10^{-5}$ m² s⁻³.

141 In both Type 1 and 2 experiments, rain precipitates through the upper layer and accumu-
142 lates 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 upper layer fluid occurs within the
149 two-phase emulsion zone shown in Figure 2 for both Type 1 and Type 2 experiments. The
150 various steps in the gravitational separation processes within the emulsion are the same as
151 described in detail by Sato and Sumita (2007) and Gilet et al., (2007) for the case in which
152 the negatively buoyant component is discontinuous (localized in drops) and the neutral
153 component forms the connected pore network. Drop dissolution ejects small volumes of the
154 rain fluid downward, forming millimeter-sized ring vortices below the emulsion interface,
155 while the less dense silicone oil percolates back into the upper layer.

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

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

167 The intensity of convection in the lower layer in Type 1 experiments is governed by a
 168 flux Rayleigh number defined in terms of F (see Appendix)

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

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

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

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

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

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

189 The two images in Figure 4 compare Type 1 rain-driven turbulent convection with large
 190 and small density excess, respectively. In Figure 4a, the rain density excess is 0.03%, the flux
 191 Rayleigh number is $Ra_F = 1.35 \times 10^9$ and the plume Reynolds number is $Re = 605$, whereas
 192 in Figure 4b the rain density excess is 2.5%, the flux Rayleigh number is approximately
 193 $Ra_F = 1.13 \times 10^{11}$, nearly two orders of magnitude larger, and the plume Reynolds number
 194 is approximately $Re = 2500$. Note the increase in turbulent mixing from the higher density
 195 rainfall and the shorter growth time of the plumes in Figure 4b compared to Figure 4a.
 196 But in terms of their large-scale structure, the two flows are remarkably similar, despite the
 197 nearly two orders of magnitude difference in Ra_F .

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

$$Re = c Ra_F^\beta \quad (4)$$

201 in which $c = 0.17 \pm 0.02$ and $\beta = 0.32 \pm 0.02$. Within experimental error, (4) is indis-
 202 tinguishable from the $\beta=1/3$ -power law exponent that corresponds to ideal inertial scaling
 203 (see Appendix). In terms of dimensional variables, a $1/3$ -power law relationship between
 204 Re and Ra_F implies $w \propto (Fd_L)^{1/3}$, i.e., the plume velocity is independent of the viscosity
 205 and the diffusivity, and depends only on the buoyancy flux and the convective layer depth.
 206 In our experiments, we mostly varied the buoyancy flux, by independently varying the rain
 207 density and the precipitation rate. In contrast, the lower layer depth was varied only by

208 small amounts, so the implied $d_L^{1/3}$ -dependence of w was not actually tested.

209 We can compare our plume Reynolds number scaling with the plume Reynolds numbers
210 measured in thermal convection experiments of the classical Rayleigh-Bénard type, in which
211 the top and bottom boundaries are rigid and isothermal. Figure 5 shows a semi-theoretic
212 scaling law for the plume Reynolds number derived by Grossmann and Lohse (2002), which
213 has the form

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

214 in which Ra and Nu are the Rayleigh number and the Nusselt number in Rayleigh-Bénard
215 convection defined in terms of the temperature difference across the fluid (see the Appendix),
216 where $Pr = \nu/\kappa$ is the Prandtl number (κ being thermal diffusivity), $f(x) = (1+x^2)^{1/4}/x^{1/2}$,
217 and c_1 and c_2 are empirical coefficients that depend on the experimental geometry and bound-
218 ary conditions (Grossmann and Lohse, 2003), and also on the technique used to measure the
219 plume velocities (Ahlers et al., 2009).

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

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

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

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

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

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

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

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

where μ_0 is free-space magnetic permeability. In order to sustain a magnetic field with dipole moment $\mathcal{M}=50 \text{ ZAm}^2$, comparable to the time-averaged Proterozoic geomagnetic field intensity (Biggin et al., 2015), (9) along with the data in Table 2 predict that a precipitation rate of $R \simeq 400 \text{ m Gyr}^{-1}$ is required. This is a negligibly small magnesium addition to the mantle and would not materially affect geochemical tracers such as Hf-W isotopic ratios (Rudge et al., 2010) or mantle siderophiles (Chabot et al., 2005). It would slightly increase Earth’s

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

Far more intense precipitation would be needed for dynamo action if the core was stratified, either thermally or by composition. If the heat loss from the convective region is less than the heat conducted down the core adiabat, a plausible situation in view of the high thermal conductivity of the core (Gomi et al., 2013; Labrosse, 2015), then an additional compositional buoyancy flux must be added to balance the negative thermal buoyancy flux due to the subadiabatic thermal gradient, in order to homogenize the core fluid. In terms of the total heat loss from the convecting region Q , the additional thermal buoyancy flux 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 surface area of the convecting region, and H_c is its temperature scale height. The net buoyancy flux that enters into the expression for the convective power (8) then becomes $F + F_T$. If the thermal buoyancy flux F_T is negative because $Q < Q_{ad}$, then a larger rain-produced buoyancy flux F is required in order to compensate for the subadiabatic heat loss.

The effects of core heat flux and magnesium precipitation rate on the dynamo structure and the induced dipole moment are shown in Figure 6, where we have assumed a convecting region outer radius of $r_c=3480$ km, equal to the present-day CMB radius, and an adiabatic heat flux of 0.1 W m^{-2} , equivalent to $Q_{ad} \simeq 15 \text{ TW}$. Line contours indicate the induced dipole moment in ZAm^2 from (9) and the local Rossby number Ro_ℓ , which has been proposed as a control parameter for the transition from dipolar to multipolar dynamo states (Christensen and Aubert, 2006; Aubert et al., 2009):

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

The dipolar-multipolar transition occurs near $Ro_\ell \sim 0.1$, according to many previous studies

295 (Christensen and Aubert, 2006; Aubert et al., 2009; Oruba and Dormy, 2014). Shadings
 296 in Figure 6 indicate dynamo regimes that are subcritical (labeled no-dynamo), convection
 297 driven by precipitation against stable thermal stratification (labeled precipitation dynamo),
 298 and convection driven by precipitation acting in concert with unstable thermal stratification
 299 (labeled thermo-precipitation dynamo).

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

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

315 The results in Figure 6 assume the convective region occupies the entire Earth’s core.
 316 In other planets the convective region may occupy only a fraction of the core, and if so,
 317 the precipitation rate required for dynamo action can be very high. For example, Figure 7

318 shows contours of magnetic Reynolds number from (7) and local Rossby number from (10)
 319 as functions of convecting radius and iron-snow precipitation rate for Mercury, based on the
 320 parameter values in Table 2 and assuming a solid inner core is absent or negligibly small.
 321 If the convective radius exceeds 200 km, a minimum precipitation rate near 30 km Gyr^{-1} is
 322 needed for dynamo onset, but in a smaller convecting region, with $r_c < 100 \text{ km}$, say, dynamo
 323 action would require far higher precipitation rates.

324 Another implication of Figure 7 is that iron-snow precipitation can generate high mag-
 325 netic Reynolds number conditions in a deep convective region in Mercury, but the large
 326 values of Ro_ℓ imply that this convection would induce a multipolar field, rather than a
 327 dipole dominant field. Accordingly, the scaling relation (9) does not apply here. Our predic-
 328 tion of a multipolar dynamo field is seemingly in contradiction with the offset dipolar field
 329 observed in orbit around Mercury (Anderson et al., 2011). It is, however, consistent with
 330 a proposed deep dynamo mechanism for Mercury by Christensen (2006), which postulates
 331 that the non-dipolar components of the dynamo magnetic field generated deep within the
 332 Mercury’s core are electromagnetically screened in an overlying stable layer. According to
 333 our experiments, such electromagnetic screening might occur in the precipitation-dominated
 334 layer because it lacks large-scale circulation.

335 Figure 8 shows the dynamo regimes for Ganymede, calculated according to the same
 336 scaling laws and assumptions as used for Earth and Mercury but with the property values
 337 for Ganymede in Table 2. The model for Ganymede differs from Mercury in having a smaller
 338 convecting region and a lower assumed thermal conductivity. A convective dynamo driven
 339 by iron snow precipitation, possibly aided by an unstable thermal gradient, is predicted
 340 for most precipitation rates. The contours show the local Rossby number Ro_ℓ . The range
 341 includes the dynamo configurations described by Christensen (2015): a dipole-dominated

dynamo at low Ro_ℓ and multipolar dynamo at high Ro_ℓ .

4 Conclusions

In this paper we describe a new class of convective flows, driven by buoyancy acquired through precipitation. This form of convection has been proposed for evolving planetary cores, where cooling results in nucleation, precipitation, and dissolution of buoyant constituents. Our rain-driven experiments are analogous to precipitation-driven systems in planetary interiors in several respects, although there are important differences.

Similarities include (1) the basic two-layer dynamical structure, with a precipitation region separated from a region where the precipitation is dissolved, with the possibility of compositional convection; (2) a thin multi-component transition zone separating these regions where the buoyancy for convection originates; (3) buoyancy production on the scale of the individual precipitates, followed by plume coalescence.

Significant differences include, for the experiments (1) lack of phase changes (solidification and melting) and latent heat exchange, which could contribute to the convection; (2) absence of rotational and magnetic field effects, which tend to stabilize the convection; (3) the idealized plane-layer geometry, which ignores the changes in fluid volume and gravity with radius; and (4) double-diffusive convection, which could increase the dynamo capacity of this system. Each of these merit additional study.

In addition, there are internal structure differences between our experiments and some of the scenarios envisioned for planetary interiors. In Earth's core, for example, it is supposed that extraction of upward precipitating magnesium or other light elements would yield a negatively buoyant residual liquid, whereas in our experiments, the negatively buoyant liquid

364 is produced through addition of downward precipitation. In cores dominated by the Fe-FeS
365 system, the geometry might be inverse of our experiment. For example, for core compositions
366 on the FeS side of the eutectic, convection might be due to sulfur-rich fluid rising from a
367 deep iron snow zone (Hauck et al. 2006; Chen et al. 2008). In spite of these differences,
368 we claim that the fundamental effects of precipitation on the dynamics of these systems are
369 captured in our experiments.

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

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

386 Lastly, we note that previous studies have found that diffusive boundary layer effects,

387 including viscosity, remain controlling factors in heat transfer in Rayleigh-Bénard convection
388 when planetary rotation is included, even at high Rayleigh numbers (King et al., 2009,
389 Stellmach et al., 2014). It would be revealing, therefore, to determine how the velocity
390 scaling for precipitation-driven convection changes when rotation is added.

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394 **Author contributions:** BH, ML, and PO jointly conducted the experiments; PO and ML
395 jointly made the calculations; PO and ML prepared the figures, tables, and references; PO
396 and ML jointly wrote the paper.

399 Appendix

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

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

402 where ρ is the mean density, $\mathbf{u}$ is the fluid velocity, ν is kinematic viscosity, and

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

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

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

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

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

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

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

415 as given in the text.

416 Rayleigh-Bénard thermal convection experiments use a different Rayleigh number, de-
417 fined as

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

418 where α and κ are thermal expansion and thermal diffusivity, respectively, and ΔT is the
419 temperature difference across the fluid. Heat transfer in Rayleigh-Bénard convection is
420 usually measured in terms of the Nusselt number, defined as

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

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

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Table 1: Table 1: Rain-Driven Convection Experiment Properties

Property	Notation	Units	Experiment Type 1	Experiment Type 2
upper fluid density	ρ_U	kg m^{-3}	0.82×10^3	0.82×10^3
lower fluid density	ρ_L	kg m^{-3}	1.0×10^3	1.0×10^3
rain-lower fluid density difference	$\Delta\rho$	kg m^{-3}	0.3-25	-10
rain viscosity	ν_r	$\text{m}^2 \text{s}^{-1}$	1×10^{-6}	1.6×10^{-6}
upper fluid viscosity	ν_U	$\text{m}^2 \text{s}^{-1}$	1.2×10^{-6}	1.2×10^{-6}
lower fluid viscosity	ν_L	$\text{m}^2 \text{s}^{-1}$	1×10^{-6}	1×10^{-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	σ_r	N m s^{-1}	$2.2 - 2.4 \times 10^{-2}$	2.6×10^{-2}
rain rate	R	m s^{-1}	$3.4 - 4.6 \times 10^{-4}$	4.5×10^{-4}
rain buoyancy flux	F	$\text{m}^2 \text{s}^{-3}$	$1.35 - 110 \times 10^{-6}$	-4.5×10^{-5}
rain drop radius	r_r	m	$\sim 10^{-3}$	$\sim 10^{-3}$
plume velocity	w	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 (Nominal)
Density Ratio, ϵ	$\Delta\rho/\rho_L$	$0.3\text{-}25 \times 10^{-3}$	-0.01	~ 4.6
Rayleigh Number, Ra_F	Fd_L^4/ν_L^3	$1.35\text{-}113 \times 10^9$	-4.5×10^{10}	$\sim 10^{29}$
Plume Reynolds Number, Re	wd_L/ν_L	$6\text{-}25 \times 10^2$	$< 2 \times 10^1$	$\sim 2 \times 10^7$
Magnetic Reynolds Number, Rm	ud_L/η	na	na	$\sim 2 \times 10^3$

Table 2: Table 2: Core Precipitation Properties (Nominal Values)

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

^a PREM Dziewonski and Anderson (1981); ^b Dumberry and Rivoldini (2015); ^c Williams (2000); ^d Ward et al. (1976); ^e Hirose et al. (2013); ^f Labrosse (2015); ^g Vocadlo et al. (2003); ^h PREM CMB buoyancy contrast; ⁱ Perrialt et al. (2010); ^j Christensen (2015); ^k Deng et al. (2013); ^l Hauk et al. (2006); ^m Chen et al. (2008); ⁿ O'Rourke and Stevenson, 2016.

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 1c 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-Bénard 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 Earth’s core (prior to inner core nucleation) driven by magnesium precipitation. Axes are core heat flux and precipitation rate. An adiabatic heat flux of 100 mW m^{-2} is assumed. Additional model 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_ℓ 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 mW m^{-2} are assumed. Additional model 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. Contours are calculated local Rossby number Ro_ℓ . Axes are heat flux at the top of the convecting region and precipitation rate. An adiabatic heat flux of 4 mW m^{-2} is assumed.

Figure 1

Figure 2

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8