

Methods

To numerically model core formation, we first identified Earth analogues from among the planets formed in the N -body simulations of Fischer and Ciesla (2014), and assigned each body in the simulations an initial composition based on its starting location. With each impact that the Earth analogue experienced, its pressure distribution was recalculated, and the incoming metal and silicate were allowed to equilibrate. First the major element compositions of the equilibrating metal and silicate were calculated, then their trace element compositions, then the resulting metal and silicate were added to the core and mantle. This process was repeated for every impact on the growing Earth analogue, whose oxygen fugacity evolves self-consistently. This supplemental text describes in greater detail how the input from simulations and experiments were applied, how the calculations were performed, and what regions of parameter space were explored.

Plausible mass evolution histories and provenances of the Earth were taken from a suite of 100 N -body simulations of terrestrial planet accretion from Fischer and Ciesla (2014), run using the MERCURY code (Chambers, 1999). Fifty simulations were run for each of two possible dynamical environments of the early Solar System, with Jupiter and Saturn on either their modern-day orbits (Eccentric Jupiter and Saturn, or EJS) or more circular orbits (Circular Jupiter and Saturn, or CJS), as predicted by the Nice Model (e.g., Tsiganis et al., 2005). The initial conditions were similar to those of previous studies (O'Brien et al., 2006; Raymond et al., 2009), but the number of simulations per dynamical environment was increased by a factor of 4–12. The simulations began with ~80 Moon- to Mars-mass planetary embryos and ~3000 smaller planetesimals, with mass initially divided evenly between the embryos and planetesimals. All collisions were treated as perfectly inelastic. This is unlikely to be the case for all collisions (e.g., Asphaug et al., 2006; Chambers, 2013; Stewart and Leinhardt, 2012). However, the modeling of

Dwyer et al. (2015) showed that including incomplete accretion may not significantly alter the average silicate:metal mass fraction in the resulting Earth analogues.

The large variations within each set of simulations were due entirely to small variations in the initial orbits of the embryos and planetesimals, as planetary accretion is an extremely stochastic process (e.g., Lissauer, 2007). 73 of 100 simulations produced an Earth analogue with a semimajor axis of 0.75–1.25 AU and a mass within a factor of 1.5 of an Earth mass (M_{\oplus}), so we used these simulations for core formation modeling. A lack of correlations between most Solar System observables indicates that any simulations that form a realistic Earth analogue can provide a plausible accretion history for the Earth, even if they do not match every other aspect of the Solar System (Fischer and Ciesla, 2014).

The liquid metal–liquid silicate partitioning behaviors of Ni, Co, V, Cr, Si, and O above 5 GPa were parameterized by Fischer et al. (2015) as functions of pressure, temperature, and metallic melt composition, incorporating data from 18 previous studies (Bouhifd and Jephcoat, 2003, 2011; Bouhifd et al., 2013; Chabot et al., 2005; Corgne et al., 2008; Geßmann and Rubie, 1998; Hillgren et al., 1996; Ito et al., 1995; Jana and Walker, 1997; Kegler et al., 2008; Mann et al., 2009; Ricolleau et al., 2011; Siebert et al., 2011, 2012, 2013; Thibault and Walter, 1995; Tsuno et al., 2013; Wade and Wood, 2005) as well as new experiments up to 100 GPa and 5700 K. Parameterizations of Ta and Nb partitioning were taken from Mann et al. (2009). The partitioning parameterizations that we used are listed in Table S2. Most of these parameterizations do not include compositional dependences and were fit to an equation of the form

$$\log_{10}(K_D^i) = a_i + \frac{b_i}{T} + \frac{c_i P}{T} \quad (\text{S1})$$

for each element i , where the temperature T is in Kelvin, pressure P is in GPa, and a_i , b_i , and c_i are fitting parameters. The exchange coefficient K_D is defined for a metal M with valence n as

$$K_D = \frac{D_M}{D_{Fe}^{n/2}} = \frac{X_M^{met}/X_{MO}^{sil}}{(X_{Fe}^{met}/X_{FeO}^{sil})^{n/2}} \quad (S2)$$

where D is a partition coefficient, defined in terms of mole fractions (e.g., X_M^{met} is the mole fraction of M in the metallic melt). In the case of oxygen, the exchange coefficient is defined as

$$K_D = \frac{X_{Fe}^{met} X_O^{met}}{X_{FeO}^{sil}} \quad (\text{Frost et al., 2010}).$$

The partitioning of V and Cr (elements i) are sensitive to metallic melt composition and were parameterized based on the epsilon model of Ma (2001):

$$\begin{aligned} \log_{10}(K_D^i) = & a_i + \frac{b_i}{T} + \frac{c_i P}{T} + \frac{1}{2.303} \sum_{\substack{k=2 \\ k \neq i}}^N \varepsilon_k^i X_k \left(1 + \frac{\ln(1-X_k)}{X_k} - \frac{1}{1-X_i} \right) \\ & - \frac{1}{2.303} \sum_{\substack{k=2 \\ k \neq i}}^N \varepsilon_k^i X_k^2 X_i \left(\frac{1}{1-X_i} + \frac{1}{1-X_k} + \frac{X_i}{2(1-X_i)^2} - 1 \right) \end{aligned} \quad (S3)$$

where X_i is the mole fraction of component i in the metallic liquid (here $i = 1$ for Fe), $\varepsilon_i^k = \varepsilon_k^i$ is the interaction parameter of elements i and k in the liquid, and the factor of 2.303 converts natural logarithms to base-10 logarithms (Fischer et al., 2015). The parameter ε_k^i at temperature T is described in the *Steelmaking Data Sourcebook* (Japan Society for the Promotion of Science and the Nineteenth Committee on Steelmaking, 1988) as $\varepsilon_k^i(T) = \frac{e_k^i M_i}{0.242} \frac{1873K}{T} - \frac{M_i}{55.85} + 1$, where e_k^i is an interaction parameter and M_i is the molar mass of element i (in g/mol). The partitioning parameterizations we used for V and Cr (Fischer et al., 2015) are also listed in Table S2.

The fits of Fischer et al. (2015) are based only on data from $P > 5$ GPa, due to a possible change in the silicate melt and/or metallic melt structure that may affect partitioning in this pressure range (e.g., Kegler et al., 2008; Palme et al., 2011; Sanloup et al., 2011). We also fit data from $P < 5$ GPa (Bouhifd and Jephcoat, 2003; Bouhifd et al., 2013; Chabot and Agee, 2003; Corgne et al., 2008; Kegler et al., 2008; Mann et al., 2009; Ricolleau et al., 2011; Richter et al.,

2010; Siebert et al., 2011; Thibault and Walter, 1995; Wade and Wood, 2005) to Eq. S1, forcing these low pressure fits to match those of Fischer et al. (2015) at 5 GPa. For V and Cr, data with carbon in the system were not included in the fitting, and these low pressure fits were forced to match the epsilon model fits of Fischer et al. (2015) extrapolated to zero light elements in the metal at 5 GPa. These fits (Table S2) were used in calculations performed below 5 GPa.

We calculated mantle and core compositions in terms of the partitioning of Ni, Co, Ta, Nb, V, Cr, Si, O, and Fe. Mg, Ca, and Al were assumed to be perfectly lithophile. Silicon and oxygen were the only light elements considered; carbon, sulfur, and other possibilities have not yet been incorporated into the modeling, due in part to a lack of experimental constraints on their partitioning behaviors and on their effects on the partitioning behaviors of other elements at $P > 25$ GPa.

First, the initial compositions of the bodies in the disk were calculated by equilibrating a CI chondrite enriched in refractory elements (Palme and O'Neill, 2003; Rubie et al., 2011) at a specified P , T , and fO_2 . All results presented here were based on initial equilibration at 0.1 GPa and 2000 K, with compositions calculated from the parameterizations in Table S2, and the oxygen content and metal–silicate ratio were allowed to vary to reach the specified fO_2 . We based our calculations on initial oxidation states of IW–3.5 and IW–1.5. An oxidation state of IW–3.5 is 1–2 log units more oxidized than an enstatite chondrite, while IW–1.5 more closely resembles some types of carbonaceous chondrites (e.g., Wadhwa, 2008; Wasson and Kallemeyn, 1988). Our starting compositions are listed in Table S1.

With each impact that the growing Earth analogue experienced, its full pressure, density, and gravity distribution was calculated, using the seismologically-determined density structure of the Earth's interior (Dziewonski and Anderson, 1981) as a pseudo-equation of state, to

accurately estimate the core–mantle boundary (CMB) pressure. Equilibration took place at a fixed fraction of the target’s CMB pressure at the time of impact and the mantle liquidus temperature at that equilibration pressure (Andrault et al., 2011). The methodology for the chemical model was inspired by that of Rubie et al. (2011). The appeal of this approach is that oxygen fugacity is allowed to evolve self-consistently as the Earth grows, rather than being imposed, which is a common but erroneous practice (e.g., Georg and Shahaar, 2015; Ricolleau et al., 2011; Siebert et al., 2011, 2012, 2013; Wade and Wood, 2005; Wade et al., 2012; Wood et al., 2009). The oxygen fugacity relative to the IW buffer (or equivalently, the FeO content of the mantle and Fe content of the core) is a result of the equilibration calculation, not an input. If an oxygen fugacity relative to IW were to be imposed, the partitioning behavior of either silicon or oxygen would be specified by that oxygen fugacity instead of constrained by experimental data. Otherwise, there would be a violation of mass balance, because any change in the iron distribution (caused by imposing an oxygen fugacity) must be compensated by a change in the silicon distribution to avoid creating or destroying oxygen atoms.

The major element compositions of the metal and silicate participating in the reaction were calculated first. Abundances of FeO, NiO, SiO₂, Fe, Ni, O, and Si were calculated using mass balance equations for Fe, Ni, O, and Si and equations describing the partitioning behaviors of Ni, O, and Si (Eq. S1; Table S2). This process approximately follows the method of Rubie et al. (2011), but uses new experimental data on oxygen partitioning (Table S2) rather than an activity model to determine the FeO content of the silicate melt.

After this procedure, trace element partitioning was calculated. K_D^V and K_D^{Cr} were calculated iteratively using the parameters in Table S2 in Eq. S3. With each iteration, the number

of moles of each trace element ($i = \text{Co, Ta, Nb, V, Cr}$) in the equilibrated metal (N_{metal}^i) was calculated as:

$$N_{metal}^i = \frac{D_i N_{total}^i N_{metal}}{N_{silicate} + N_{metal} D_i} \quad (\text{S4})$$

where D_i is a partition coefficient (Eq. S2), N_{total}^i is the total number of moles of element i participating in the equilibration reaction, N_{metal} is the total number of moles of metal participating in the equilibration (a sum over N_{metal}^i for all i and major elements), and $N_{silicate}$ is similarly the total number of moles of oxides participating in the equilibration. The moles of each trace element in the equilibrated silicate ($N_{silicate}^i$) can then simply be calculated from a mass balance: $N_{silicate}^i = N_{total}^i - N_{metal}^i$.

Once the equilibrated metal and silicate compositions were calculated in terms of both major and trace elements, the equilibrated metal was added to the growing Earth's core, the equilibrated silicate was added to the growing Earth's mantle, and each reservoir was assumed to have internally mixed before the next impact. This process was repeated for each body that the Earth accreted.

In some cases (Figures 5–7, Figure S3, Table S3), uncertainties in composition were calculated to reflect the variability that is possible due to accretion history alone for an Earth analogue planet that forms at 1 AU. We performed an unweighted linear fit to the desired aspect of planetary chemistry (e.g., the NiO content of a planet's mantle) as a function of final planetary mass, for Earth analogues that formed with final masses of 0.9–1.1 M_{\oplus} ($n = 23$). Two sigma uncertainties were calculated as twice the average deviation from this line for these planets (absolute value).

Trade-offs between different parameters: Extended discussion

The tradeoff between depth of equilibration and amount of metal and silicate equilibrating can be further visualized in Figure S7. This figure shows which combinations of these parameters produce the mantle's NiO composition, which can be described numerically as:

$$depth = \frac{0.0363}{k-0.0898} + \frac{0.1035}{silmass-0.1446} + \frac{0.0170}{(k-0.0898)(silmass-0.1446)} + 0.3636 \quad (S5)$$

where *depth* is the depth of equilibration (expressed as a fraction of the CMB pressure), *k* is the fraction of the impactor's metal that equilibrates, and *silmass* is the mass of silicate that equilibrates (expressed as a multiple of the impactor's silicate mass). If future work leads to constraints on one or two of these parameters, this relationship can be used to constrain the other(s). This relationship is defined for equilibration along the liquidus of Andraut et al. (2011) with initial oxidation states of materials described by a step at 2 AU from IW–3.5 to IW–1.5 (the reference set of parameters).

References

- Andraut, D., Bolfan-Casanova, N., Lo Nigro, G., Bouhifd, M.A., Garbarino, G., Mezouar, M., 2011. Solidus and liquidus profiles of chondritic mantle: Implication for the melting of the Earth across its history. *Earth Planet. Sci. Lett.* 304, 251–259, doi:10.1016/j.epsl.2011.02.006.
- Badro, J., Brodholt, J.P., Piet, H., Siebert, J., Ryerson, F.J., 2015. Core formation and core composition from coupled geochemical and geophysical constraints. *Proc. Natl. Acad. Sci. USA* 112, 12310–12314, doi:10.1073/pnas.1505672112.
- Bouhifd, M.A., Jephcoat, A.P., 2003. The effect of pressure on partitioning of Ni and Co between silicate and iron-rich metal liquids: A diamond-anvil cell study. *Earth Planet. Sci. Lett.* 209, 245–255, doi:10.1016/S0012-821X(03)00076-1.

- Bouhifd, M.A., Jephcoat, A.P., 2011. Convergence of Ni and Co metal–silicate partition coefficients in the deep magma-ocean and coupled silicon–oxygen solubility in iron melts at high pressures. *Earth Planet. Sci. Lett.* 307, 341–348, doi:10.1016/j.epsl.2011.05.006.
- Bouhifd, M.A., Andrault, D., Bolfan-Casanova, N., Hammouda, T., Devidal, J.L., 2013. Metal–silicate partitioning of Pb and U: Effects of metal composition and oxygen fugacity. *Geochim. Cosmochim. Acta* 114, 13–28, doi:10.1016/j.gca.2013.03.034.
- Chabot, N.L., Agee, C.B., 2003. Core formation in the Earth and Moon: New experimental constraints from V, Cr, and Mn. *Geochim. Cosmochim. Acta* 67, 2077–2091, doi:10.1016/S0016-7037(02)01272-3.
- Chabot, N.L., Draper, D.S., Agee, C.B., 2005. Conditions of core formation in the Earth: Constraints from nickel and cobalt partitioning. *Geochim. Cosmochim. Acta* 69, 2141–2151, doi:10.1016/j.gca.2004.10.019.
- Corgne, A., Keshav, S., Wood, B.J., McDonough, W.F., Fei, Y., 2008. Metal–silicate partitioning and constraints on core composition and oxygen fugacity during Earth accretion. *Geochim. Cosmochim. Acta* 72, 574–589, doi:10.1016/j.gca.2007.10.006.
- Chambers, J.E., 1999. A hybrid symplectic integrator that permits close encounters between massive bodies. *Mon. Not. R. Astron. Soc.* 304, 793–799, doi:10.1046/j.1365-8711.1999.02379.x.
- Corgne, A., Keshav, S., Wood, B.J., McDonough, W.F., Fei, Y., 2008. Metal–silicate partitioning and constraints on core composition and oxygen fugacity during Earth accretion. *Geochim. Cosmochim. Acta* 72, 574–589, doi:10.1016/j.gca.2007.10.006.

- Dwyer, C.A., Nimmo, F., Chambers, J.E., 2015. Bulk chemical and Hf–W isotopic consequences of incomplete accretion during planet formation. *Icarus* 245, 145–152, doi:10.1016/j.icarus.2014.09.010.
- Dziewonski, A.M., Anderson, D.L., 1981. Preliminary reference Earth model. *Phys. Earth Planet. Inter.* 4, 297–356, doi:10.1016/0031-9201(81)90046-7.
- Fischer, R.A., Ciesla, F.J., 2014. Dynamics of the terrestrial planets from a large number of *N*-body simulations. *Earth Planet. Sci. Lett.* 392, 28–38, doi:10.1016/j.epsl.2014.02.011.
- Fischer, R.A., Nakajima, Y., Campbell, A.J., Frost, D.J., Harries, D., Langenhorst, F., Miyajima, N., Pollok, K., Rubie, D.C., 2015. High pressure metal–silicate partitioning of Ni, Co, V, Cr, Si, and O. *Geochim. Cosmochim. Acta* 167, 177–194, doi:10.1016/j.gca.2015.06.026.
- Frost, D.J., Asahara, Y., Rubie, D.C., Miyajima, N., Dubrovinsky, L.S., Holzapfel, C., Ohtani, E., Miyahara, M., Sakai, T., 2010. Partitioning of oxygen between the Earth’s mantle and core. *J. Geophys. Res.* 115, B02202, doi:10.1029/2009JB006302.
- Geßmann, C.K., Rubie, D.C., 1998. The effect of temperature on the partitioning of nickel, cobalt, manganese, chromium, and vanadium at 9 GPa and constraints on formation of the Earth’s core. *Geochim. Cosmochim. Acta* 62, 867–882, doi:10.1016/S0016-7037(98)00023-4.
- Hillgren, V.J., Drake, M.J., Rubie, D.C., 1996. High pressure and high temperature metal–silicate partitioning of siderophile elements: The importance of silicate liquid composition. *Geochim. Cosmochim. Acta* 60, 2257–2263, doi:10.1016/0016-7037(96)00079-8.
- Ito, E., Morooka, K., Ujike, O., Katsure, T., 1995. Reactions between molten iron and silicate melts at high pressure: Implications for the chemical evolution of Earth’s core. *J. Geophys. Res.* 100, 5901–5910, doi:10.1029/94JB02645.

- Jana, D., Walker, D., 1997. The impact of carbon on element distribution during core formation. *Geochim. Cosmochim. Acta* 61, 2759–2763, doi:10.1016/S0016-7037(97)00091-4.
- Japan Society for the Promotion of Science and the Nineteenth Committee on Steelmaking, 1988. Part II: Recommended values of activity coefficients and interaction parameters of elements in iron alloys. In: *Steelmaking Data Sourcebook*, Gordon and Breach Science Publishers, New York, pp. 273–297.
- Kegler, Ph., Holzheid, A., Frost, D.J., Rubie, D.C., Dohmen, R., Palme, H., 2008. New Ni and Co metal–silicate partitioning data and their relevance for an early terrestrial magma ocean. *Earth Planet. Sci. Lett.* 268, 28–40, doi:10.1016/j.epsl.2007.12.020.
- Lissauer, J.J., 2007. Planets formed in habitable zones of M dwarf stars probably are deficient in volatiles. *Astrophys. J.* 660, L149–L152, doi:10.1086/518121.
- Ma, Z., 2001. Thermodynamic description for concentrated metallic solutions using interaction parameters. *Metall. Mater. Trans. B* 32B, 87–103, doi:10.1007/s11663-001-0011-0.
- Mann, U., Frost, D.J., Rubie, D.C., 2009. Evidence for high-pressure core–mantle differentiation from the metal–silicate partitioning of lithophile and weakly-siderophile elements. *Geochim. Cosmochim. Acta* 73, 7360–7386, doi:10.1016/j.gca.2009.08.006.
- McDonough, W.F., Sun, S.-s., 1995. The composition of the Earth. *Chem. Geol.* 120, 223–253, doi:10.1016/0009-2541(94)00140-4.
- O’Brien, D.P., Morbidelli, A., Levison, H.F., 2006. Terrestrial planet formation with strong dynamical friction. *Icarus* 184, 39–58, doi:10.1016/j.icarus.2006.04.005.
- Palme, H., O’Neill, H.St.C., 2003. Cosmochemical estimates of mantle composition. In: Carlson, R.W. (Ed.), *Treatise on Geochemistry*, Vol. 2, Elsevier, pp. 1–38.

- Palme, H., Kegler, Ph., Holzheid, A., Frost, D.J., Rubie, D.C., 2011. Comment on “Prediction of metal–silicate partition coefficients for siderophile elements: An update and assessment of PT conditions for metal–silicate equilibrium during accretion of the Earth” by K. Righter, EPSL 304 (2011) 158–167, 2011. *Earth Planet. Sci. Lett.* 312, 516–618, doi:10.1016/j.epsl.2011.10.018.
- Raymond, S.N., O’Brien, D.P., Morbidelli, A., Kaib, N.A., 2009. Building the terrestrial planets: Constrained accretion in the inner Solar System. *Icarus* 203, 644–662, doi:10.1016/j.icarus.2009.05.016.
- Ricolleau, A., Fei, Y., Corgne, A., Siebert, J., Badro, J., 2011. Oxygen and silicon contents of Earth’s core from high pressure metal–silicate partitioning experiments. *Earth Planet. Sci. Lett.* 310, 409–421, doi:10.1016/j.epsl.2011.08.004.
- Righter, K., Pando, K.M., Danielson, L., Lee, C.-T., 2010. Partitioning of Mo, P and other siderophile elements (Cu, Ga, Sn, Ni, Co, Cr, Mn, V, and W) between metal and silicate melt as a function of temperature and silicate melt composition. *Earth Planet. Sci. Lett.* 291, 1–9, doi:10.1016/j.epsl.2009.12.018.
- Rubie, D.C., Frost, D.J., Mann, U., Asahara, Y., Nimmo, F., Tsuno, K., Kegler, Ph., Holzheid, A., Palme, H., 2011. Heterogeneous accretion, composition and core–mantle differentiation of the Earth. *Earth Planet. Sci. Lett.* 301, 31–42, doi:10.1016/j.epsl.2010.11.030.
- Sanloup, C., van Westrenen, W., Dasgupta, R., Maynard-Casely, H., Perrilat, J.-P., 2011. Compressibility change in iron-rich melt and implications for core formation models. *Earth Planet. Sci. Lett.* 306, 118–112, doi:10.1016/j.epsl.2011.03.039.

- Siebert, J., Corgne, A., Ryerson, F.J., 2011. Systematics of metal–silicate partitioning for many siderophile elements applied to Earth’s core formation. *Geochim. Cosmochim. Acta* 75, 1451–1489, doi:10.1016/j.gca.2010.12.013.
- Siebert, J., Badro, J., Antonangeli, D., Ryerson, F.J., 2012. Metal–silicate partitioning of Ni and Co in a deep magma ocean. *Earth Planet. Sci. Lett.* 321–322, 189–197, doi:10.1016/j.epsl.2012.01.013.
- Siebert, J., Badro, J., Antonangeli, D., Ryerson, F.J., 2013. Terrestrial accretion under oxidizing conditions. *Science* 339, 1194–1197, doi:10.1126/science.1227923.
- Thibault, Y., Walter, M.J., 1995. The influence of pressure and temperature on the metal–silicate partition coefficients of nickel and cobalt in a model C1 chondrite and implications for metal segregation in a deep magma ocean. *Geochim. Cosmochim. Acta* 59, 991–1002, doi:10.1016/0016-7037(95)00017-8.
- Tsiganis, K., Gomes, R., Morbidelli, A., Levison, H.F., 2005. Origin of the orbital architecture of the giant planets of the Solar System. *Nature* 435, 459–461, doi:10.1038/nature03539.
- Tsuno, K., Frost, D.J., Rubie, D.C., 2013. Simultaneous partitioning of silicon and oxygen into the Earth’s core during early Earth differentiation. *Geophys. Res. Lett.* 40, 66–71, doi:10.1029/2012GL054116.
- Wade, J., Wood, B.J., 2005. Core formation and the oxidation state of the Earth. *Earth Planet. Sci. Lett.* 236, 78–95, doi:10.1016/j.epsl.2005.05.017.
- Wade, J., Wood, B.J., Tuff, J., 2012. Metal–silicate partitioning of Mo and W at high pressures and temperatures: Evidence for late accretion of sulphur to the Earth. *Geochim. Cosmochim. Acta* 85, 58–74, doi:10.1016/j.gca.2012.01.010.

Wadhwa, M., 2008. Redox conditions on small bodies, the Moon and Mars. *Rev. Mineral.*

Geochem. 68, 493–510, doi:10.2138/rmg.2008.68.17.

Wasson, J.T., Kallemeyn, G.W., 1988. Compositions of chondrites. *Phil. Trans. R. Soc. Lond. A*

325, 535–544, doi:10.1098/rsta.1988.0066.

Wood, B.J., Wade, J., Kilburn, M.R., 2009. Core formation and the oxidation state of the Earth:

Additional constraints from Nb, V and Cr partitioning. *Geochim. Cosmochim. Acta* 72,

1415–1426, doi:10.1016/j.gca.2007.11.036.

Supplemental tables and table captions

Table S1: Starting compositions of planetary embryos and planetesimals. Compositions were determined by equilibrating a CI chondrite enriched in refractory elements (Palme and O'Neill, 2003; Rubie et al., 2011) at 0.1 GPa, 2000 K, and either IW–1.5 or IW–3.5. All compositions are reported in wt%.

	(wt%)	IW–1.5	IW–3.5
Silicate	MgO	29.7	37.6
	CaO	2.97	3.77
	AlO _{1.5}	3.67	4.64
	TaO _{2.5}	3.94×10^{-6}	4.95×10^{-6}
	VO _{1.5}	0.0164	0.0142
	SiO ₂	42.6	51.2
	NbO _{2.5}	8.00×10^{-5}	7.04×10^{-5}
	CrO	0.632	0.567
	FeO	20.4	2.22
	CoO	0.00547	3.50×10^{-4}
	NiO	0.0154	9.45×10^{-4}
Metal	Ta	2.06×10^{-10}	7.68×10^{-8}
	V	0.386×10^{-4}	0.00993
	Si	0.025	2.84
	Nb	1.80×10^{-7}	4.70×10^{-5}
	Cr	0.0492	0.415
	Fe	88.8	90.8
	Co	0.430	0.259
	Ni	9.56	5.53

O	1.17	0.117
Metal mass fraction	0.173	0.315

Table S2: Partitioning parameterizations used in core formation modeling (Eqs. S1 and S3), for elements i (with valences n). Asterisks (*) indicate parameterizations for use in calculations below 5 GPa only. See supplemental text for studies on which each parameterization is based.

i	n	a_i	b_i	c_i	e_O^i	e_{Si}^i	P range of data	Source of parameterization
Ni*	2	0.46	3400	-200			0–4 GPa	This study
Ni	2	0.46	2700	-61			5–100 GPa	Fischer et al. (2015)
Co*	2	0.36	1800	-92			0–3.5 GPa	This study
Co	2	0.36	1500	-33			5–100 GPa	Fischer et al. (2015)
Ta	5	0.84	-13806	-101			2–24 GPa	Mann et al. (2009)
Nb	5	2.66	-14032	-199			2–24 GPa	Mann et al. (2009)
V*	3	-1.5	-2200	-17			0.5–3 GPa	This study
V	3	-1.5	-2300	9	-0.077	0.039	5–100 GPa	Fischer et al. (2015)
Cr*	2	-0.3	-2900	130			0.5–3 GPa	This study
Cr	2	-0.3	-2200	0	-0.037	0	5–74 GPa	Fischer et al. (2015)
Si*	4	1.3	-11400	-430			1.5–3.6 GPa	This study
Si	4	1.3	-13500	0			6–100 GPa	Fischer et al. (2015)
O*		0.6	-2500	-240			2–3.6 GPa	This study
O		0.6	-3800	22			5–100 GPa	Fischer et al. (2015)

Table S3: Summary of results from all model runs, listing NiO, CoO, and FeO contents of the mantle; Si and O contents of the core; and the core mass fraction, as a function of model parameters used. Each row of data represents the average results from all 73 Earth analogues for the set of model parameters described in that row, with uncertainties reported as 2σ variability due to accretion history for planets with masses of $0.9\text{--}1.1 M_{\oplus}$. “Step”: location of step in initial oxygen fugacity distribution from IW-3.5 to IW-1.5. “Depth”: depth of equilibration, expressed as a fraction of the core–mantle boundary pressure. “k”: fraction of incoming metal that

equilibrates. “mantle k”: fraction of impactor or target+impactor silicate that equilibrates.

“tar:imp mantle”: ratio of target to impactor silicate that equilibrates. Relevant portion of target silicate is added to the impactor silicate according to this ratio, then a portion of this mixture is equilibrated according to the mantle k value. “deltaT”: deviation from the mantle liquidus of Andraut et al. (2011) in K. “CMF”: core mass fraction, expressed as a percentage.

Table S4: Average core and mantle compositions produced in several models, compared to observed values. Analogous to Table 1 of the main text, but showing separate average compositions for different styles of accretion. Values and uncertainties are averages and standard deviations of 35 EJS simulations or 38 CJS simulations. All compositions are reported in weight percent. Earth’s mantle composition is from McDonough and Sun (1995). CMF, core mass fraction. EJS, Eccentric Jupiter and Saturn model. CJS, Circular Jupiter and Saturn model. fO_2 is calculated from Eq. 3 of the main text. Reduced case: all bodies initially at IW–3.5. Oxidized case: all bodies initially at IW–1.5. Reference parameter set includes a step function in initial oxidation state at 2 AU. All model parameters are held fixed at their reference values except for the initial distribution of oxidation states.

Supplemental figure captions

Figure S1: Evolution of the FeO content of the mantle during accretion, for all 73 Earth analogues run through the core formation model with the reference case parameters (cf. Figure 1 of Badro et al., 2015). Shaded region is Earth value from McDonough and Sun (1995). Color is a guide for the eye only. Circles indicate final values.

Figure S2: Change in bulk oxidation state (Eq. 3 of the main text) of the modeled Earth analogue for different oxidation states of accreted material. Results are shown as a histogram of the change in oxidation state for EJS (filled bars) and CJS (open bars) simulations forming an Earth analogue, summarizing the aggregate results from all Earth analogues. Purple: all accreted material is initially reduced (IW−3.5). Teal: reference parameters, in which material originating inside of 2 AU is reduced and material originating outside of 2 AU is oxidized (IW−1.5). Green: all material is initially oxidized. Other model parameters are held constant to their reference values. Values and uncertainties are averages and standard deviations from all 73 Earth analogues.

Figure S3: Si content of the core (left axis, red diamonds) and NiO content of the mantle (right axis, orange circles) as a function of the depth of metal–silicate equilibration, for 20% of the incoming metal equilibrating (k). All core formation model parameters except the depth and degree of metal equilibration are fixed at their reference values. Each data point represents the average of all 73 Earth analogues, with error bars representing the 2σ variability due to accretion history for planets with masses of 0.9–1.1 M_{\oplus} . Shaded region is Earth value from McDonough and Sun (1995). This figure is analogous to Figure 5a and b of the main text, with $k = 0.2$ instead of $k = 1$.

Figure S4: Effects of depth of equilibration on composition. All core formation model parameters are held fixed to their reference values except for the depth of equilibration. (a) Mantle composition. Left axis: CoO (blue triangles) and $\text{VO}_{1.5}$ (peach circles). Right axis: SiO_2 (teal squares) and FeO (light green diamonds). (b) Core composition. Left axis: oxygen (navy

blue squares), silicon (red diamonds), and nickel (brown triangles). Right axis: niobium (blue circles). Shaded regions indicate mantle composition from McDonough and Sun (1995). Each data point represents the average of all 73 Earth analogues.

Figure S5: Final Si and O contents of planetary cores from 73 Earth analogues formed in *N*-body simulations, run through a core formation model with the reference case parameters but variable temperature offsets from the liquidus. Each colored line represents one Earth analogue, illustrating how its core composition changes with the temperature of equilibration (cf. Figure 2 of Badro et al., 2015). Color is a guide for the eye only. The three points on each line correspond to equilibration at 300 K below the mantle liquidus, the mantle liquidus of Andraut et al. (2011), and 600 K above this liquidus.

Figure S6: Effects of temperature of equilibration on mantle (a) and core (b) compositions. Temperatures are expressed as an offset from the mantle liquidus of Andraut et al. (2011). All core formation model parameters are held fixed to their reference values except for this temperature offset. Symbols and colors are as in Figure S4. Shaded regions indicate mantle composition from McDonough and Sun (1995). Each data point represents the average of all 73 Earth analogues.

Figure S7: Depth of equilibration required to match the mantle's NiO content (on average, considering the aggregate of all 73 Earth analogues) for different combinations of the fraction of incoming metal that equilibrates and the mass of silicate that equilibrates. All other model parameters were held fixed to their reference values. The tradeoff between these three parameters

is described numerically by Eq. S5. Black region is unphysical parameter space. The impactor silicate always equilibrates; variable amounts of target silicate are added to vary the equilibrating silicate mass (in some impacts, this mass was necessarily lower; see Section 2 of the main text for details).