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Empirical constraints on the salinity of the europan ocean and implications for a thin ice shell

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Abstract

Induced electrical currents within Europa inferred from *Galileo* spacecraft magnetometer instrument data have been interpreted as due to a salty europan ocean. Published compositional models for Europa's ocean, based on aqueous leaching of carbonaceous chondrites, range over five orders of magnitude in predicted magnesium sulfate concentrations. We combine the *Galileo* spacecraft magnetometer-derived oceanic conductivities and radio Doppler data-derived interior models with laboratory conductivity vs concentration data for both magnesium sulfate solutions and terrestrial seawater to determine empirically the range of salt concentrations permitted for Europa's ocean. Solutions for both a three-layer spherical model, and a five-layer half-space model, that satisfy current preferred best fits to magnetometer data imply high, near-saturation salt concentrations and require a europan ice shell of less than 15 km thick, with a best fit at 4 km ice thickness. Adding a conductive core and mantle has a negligible effect on the amplitude when ocean conductivities are greater than a few Siemens per meter. Similarly, we find that including a realistic ionosphere has a negligible effect. We examine the implications of these results for the subsurface habitability of Europa.

Keywords: Europa; Astrobiology; Satellites, composition; Jupiter, magnetosphere

1. Introduction

Both the salt concentration (defined as g dissolved salts per kg of H_2O ($kg^{-1}_{H_2O}$)) of Europa's ocean, and the overlying ice shell thickness, have been only poorly constrained. Experimental (Fanale et al., 2001) and theoretical (Kargel et al., 2000; Zolotov and Shock, 2001; McKinnon and Zolensky, 2003) compositional studies that assume that Europa's salts derive from the leaching or aqueous alteration of carbonaceous chondrites suggest that Europa's ocean should have magnesium (Mg^{2+}) as the dominant cation and sulfate (SO_4^{2-}) as the dominant anion. These results are broadly consistent with *Galileo* spacecraft observations of the near-infrared ice absorption bands (1.0, 1.25, 1.5, and 2.0 μm) of Europa's surface (McCord et al., 1999), although there are alternative models for the identification of these bands (Carlson et al., 1999; Dalton et al., 2003). Total extraction models can yield salt concentrations as high as ∼560 g $MgSO_4\,kg^{-1}_{H_2O}$ (Kargel et al., 2000) or even higher, 50–50 mixtures of 1000 g $MgSO_4\,kg^{-1}_{H_2O}$ (Zolotov and Shock, 2001). Zolotov and Shock's (2001) preferred partial extraction model gives 7.6 g $MgSO_4\,kg^{-1}_{H_2O}$, with other models ranging as low as 0.018 g $MgSO_4\,kg^{-1}_{H_2O}$, corresponding to 0.0036 g $Mg^{2+}\,kg^{-1}_{H_2O}$. McKinnon and Zolensky (2003), arguing against a sulfate-rich model, conclude that $MgSO_4$ concentrations should lie below ∼100 g $MgSO_4\,kg^{-1}_{H_2O}$. That is, compositional models currently in the literature allow an almost five order of magnitude range in the possible $MgSO_4$ concentration in Europa's ocean.

Constraints on the ice shell thickness are currently limited by surface observations and, broadly, by the gravity data (Anderson et al., 1998). Analyses of europan surface features (Greeley et al., 2004; Greenberg et al., 2002), crater morphologies (Turtle and Pierazzo, 2001), ice convection models (Pappalardo et al., 1998), and thermal models (Ojakangas and

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Stevenson, 1989) point to an ice shell thickness in the range of ~3 to >30 km. Comparison of *Voyager* and *Galileo* image data provide no evidence for contemporary, active resurfacing (Phillips et al., 2000). Differences in the interpretation of linaea, cycloids, ridges, and chaos features are hard to resolve without additional data and without knowledge of the age of the features. Some workers have made a case that while some of the observed features fit a thin-shell model, such features could be indicative of a thin-shell epoch occurring tens of millions of years ago (Pappalardo et al., 1998). Contemporary Europa could have a thick ice shell (>30 km) with thin-shell features literally frozen in time on the surface. Thermal-orbital evolution models support this possibility (Hussmann and Spohn, 2004; Hussmann et al., 2002).

Here we address empirical constraints on the ice shell thickness by examining the relationship between the induced magnetic field signature of Europa and the conductivity of the putative subsurface ocean. We are able to set limits on the inductive response of the ocean, thereby necessitating a thin ice shell in order to explain the observed induced magnetic field.

2. Conductivity vs concentration

The conductivity of Europa's ocean, inferred from fitting *Galileo* spacecraft magnetometer measurements to a near-surface conducting layer, must exceed 58 $\mathrm{mS\,m^{-1}}$ in order to explain the observed induced magnetic field (Zimmer et al., 2000). However, coupling the magnetometer measurements to the gravity data, thereby constraining the hydrosphere to be <200 km thick (Anderson et al., 1998), requires the conductivity of the ocean to exceed 72 $\mathrm{mS\,m^{-1}}$ (Zimmer et al., 2000). More recent modeling work by Schilling and Neubauer (2005) argues for a minimum conductivity of 250 $\mathrm{mS\,m^{-1}}$. If Europa's ocean had the composition of seawater (dominated by the salt NaCl), empirical fits for standard seawater conductivity as a function of salinity and temperature could be used to determine salinity (defined as g salt per kg of solution) (Poisson, 1980; Denny, 1993). For the case $T = 0\,^\circ\mathrm{C}$ (a reasonable first estimate for europan sea temperatures, though see Melosh et al., 2004 for more detail), salinity s and conductivity σ are related via

$$s = 35\big[A_1\big(R_0^{5/2} - R_0\big) + A_2\big(R_0^2 - R_0\big) + A_3\big(R_0^{3/2} - R_0\big) + R_0\big] - 15R_0(R_0 - 1)\big(B_0 + R_0^{1/2}B_1R_0\big), \tag{1}$$

where $R_0 = 0.3443\sigma$ for σ in $\mathrm{S\,m^{-1}}$, $A_1 = 0.08887$, $A_2 = -0.2368$, $A_3 = 0.4450$, $B_0 = 0.03579$, and $B_1 = -0.001529$. For a solution of terrestrial seawater ions at $0\,^\circ\mathrm{C}$, $\sigma = 72$ $\mathrm{mS\,m^{-1}}$ corresponds to $s = 0.68\ \mathrm{g\,kg^{-1}}$, or about fifty times below the salinity of Earth's ocean. At a conductivity of $\sigma = 250\ \mathrm{mS\,m^{-1}}$ the salinity rises to 2.5 $\mathrm{g\,kg^{-1}}$, still over twelve times less than that of the terrestrial ocean.

However, most models (Kargel et al., 2000; McKinnon and Zolensky, 2003; Zolotov and Shock, 2001) as well as aqueous leaching experiments (Fanale et al., 2001) suggest $MgSO_4$, not NaCl, as the dominant salt in Europa's ocean. Therefore, we have searched the literature to obtain concentration–conductivity data for $MgSO_4$. In Fig. 1 we plot the available conductivity–concentration data for $MgSO_4$ along with, for comparison, Eq. (1) for terrestrial ocean conductivity–salinity. Fig. 1 shows magnetometer constraints for conductivity as well as some published compositional model constraints on salinity. Conductivity data at $0\,^\circ\mathrm{C}$ (Washburn and Klemenc, 1936; Bremner et al., 1939) have been combined with data taken at $25\,^\circ\mathrm{C}$ (Calvert et al., 1958; Pethybridge and Taba, 1977; Dunsmore and James, 1951; Harkins and Paine, 1919; Fisher, 1962; Fisher and Fox, 1979) that we have scaled to $0\,^\circ\mathrm{C}$ by multiplying the $25\,^\circ\mathrm{C}$ conductivity data by a scaling factor of 0.525. This scaling is possible because Washburn and Klemenc (1936) report conductivity–concentration data for both 0 and $25\,^\circ\mathrm{C}$ across the concentration range from 0.6 to 60 g $MgSO_4\ \mathrm{kg}_{\mathrm{H_2O}}^{-1}$. The scaling factor was calculated by finding the ratio of conductivities for $T = 0$ and $T = 25\,^\circ\mathrm{C}$ for a given salt concentration. The average of the nine ratios available from the Washburn and Klemenc (1936) data was used as our final scaling factor. The fact that the temperature-scaled data fall along the curve sketched by the more sparse Washburn and Klemenc's (1936) $0\,^\circ\mathrm{C}$ and the Bremner et al.'s (1939) $0\,^\circ\mathrm{C}$ data suggest that this temperature scaling has been done correctly. We fit a third-degree polynomial to the data,

$$C = c_3\sigma^3 + c_2\sigma^2 + c_1\sigma + c_0, \tag{2}$$

finding the coefficients, $c_3 = -1.7268$, $c_2 = 12.0161$, $c_1 = 15.2108$, $c_0 = -0.0129$, where C is concentration and σ is conductivity. Errors for values calculated using the polynomial fit are on the order of a few percent across the range of conductivities considered here.

From Eq. (2) or Fig. 1, we see that the lower conductivity limit of 58 $\mathrm{mS\,m^{-1}}$ corresponds to a minimum $MgSO_4$ concentration of 0.91 g $MgSO_4\ \mathrm{kg}_{\mathrm{H_2O}}^{-1}$. With the added constraint of a hydrosphere <200 km thick ($\sigma > 72\ \mathrm{mS\,m^{-1}}$), we find a minimum salt concentration of 1.1 g $MgSO_4\ \mathrm{kg}_{\mathrm{H_2O}}^{-1}$. This value is a factor of 50 greater than the lower limit partial extraction model of Zolotov and Shock (2001) so that the lowest extraction models are excluded by the data. The lower limit on σ proposed by Schilling and Neubauer (2005), based on longer period wave propagation, requires 4.5 g $MgSO_4\ \mathrm{kg}_{\mathrm{H_2O}}^{-1}$. Zimmer et al. (2000) do not explicitly provide an upper limit to the conductivity of the ocean. However, they determine that an ocean of terrestrial salinity ($s = 35$, or a concentration of 34 g sea salt per kg of water ($\mathrm{kg}_{\mathrm{H_2O}}^{-1}$) at $0\,^\circ\mathrm{C}$, corresponding to 2.75 $\mathrm{S\,m^{-1}}$) would imply a near-surface ocean of only 3.5 km thickness. (An insulating ice shell is ignored in their model; they therefore treat the uppermost 3.5 km of Europa as conducting.) A shell of conductivity 2.75 $\mathrm{S\,m^{-1}}$ would require 96.8 g $MgSO_4\ \mathrm{kg}_{\mathrm{H_2O}}^{-1}$.

In summary, the range of empirically supported $MgSO_4$ salt concentrations permitted by Fig. 1 and previously published conductivity constraints falls between 1.1 and 96.8 g $MgSO_4\ \mathrm{kg}_{\mathrm{H_2O}}^{-1}$. Note that the high-end value is not well constrained by values in the literature. Therefore in subsequent sections we work to provide a stronger constraint on this value and show that empirical constraints from *Galileo*, combined with physical limitations on the conductivity resulting from salinity, necessitate a thin ice shell.

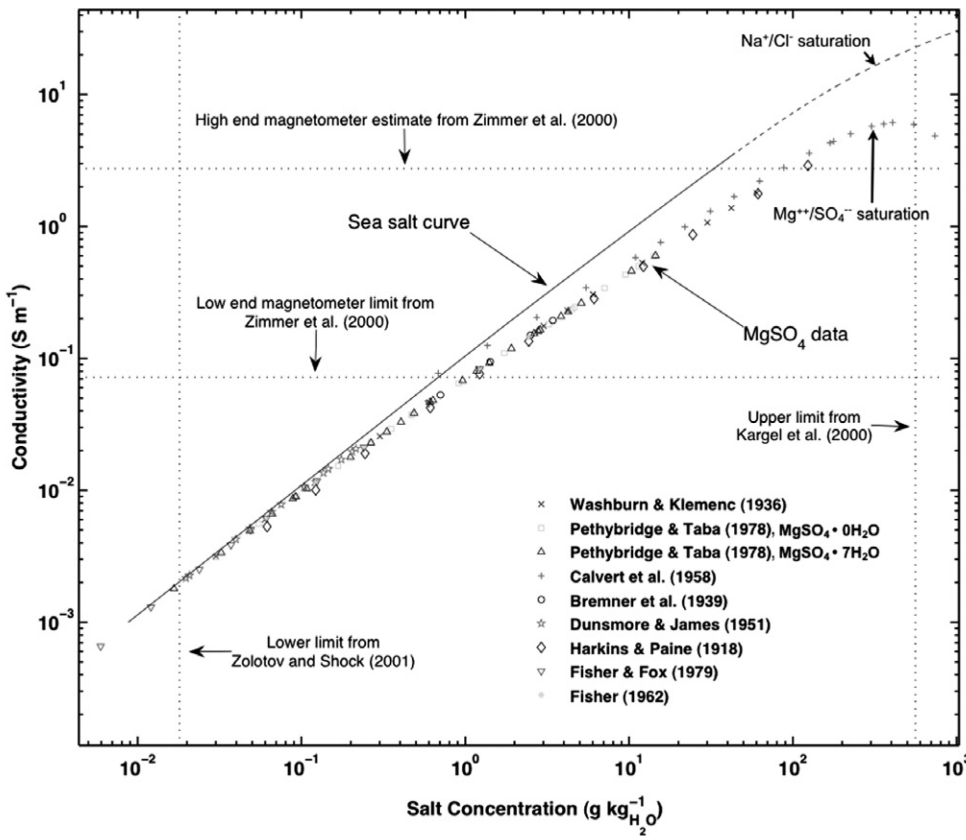


Fig. 1. Specific conductivity of an aqueous solution as a function of sea salt or $MgSO_4$ concentration. The sea salt curve was calculated using the polynomial of Poisson (1980) (see text). The region of the curve beyond 42 g salt $kg^{-1}_{H_2O}$ is shown in dashes in order to emphasize that this region extends beyond the data of Poisson (1980). Data from Washburn and Klemenc (1936) and Bremner et al. (1939) represent actual measurements of a magnesium sulfate solution at 0 °C. All other data represent measurements made at 25 °C and then scaled to at 0 °C by using a scaling factor derived from the 0 and 25 °C data of Washburn and Klemenc (1936). Limits for the dissolution of NaCl (304 g $kg^{-1}_{H_2O}$) and $MgSO_4$ (282 g $kg^{-1}_{H_2O}$) are shown.

3. Amplitude constraints

We derive a first set of limits for conductivity by combining information from Zimmer et al. (2000) with the gravity measurement constraints of Anderson et al. (1998), viz. that the total thickness of Europa's ice plus liquid water shell must lie in the range 80 to 170 km.

We solve for the induced magnetic field amplitude for a range of ocean thicknesses and ice shell thickness. The total magnetic field, as observed by the *Galileo* spacecraft, consists of the primary field of Jupiter plus the secondary, induced field of Europa: $\mathbf{B} = \mathbf{B}_{\text{primary}} + \mathbf{B}_{\text{secondary}}$. The induced magnetic field is a function of the primary field, and in the three-shell model (ice, ocean, and mantle) of Zimmer et al. (2000) this field is a dipole field (Srivastava, 1966; Parkinson, 1983) and can be expressed as

$$\mathbf{B}_{\text{secondary}} = -Ae^{-i(\omega t-\phi)} B_{\text{primary}}\big[3(\mathbf{r}\cdot\mathbf{e}_0)r - r^2\mathbf{e}_0\big]r_m^2/(2r^5) \tag{3}$$

The amplitude A and phase ϕ of the induced magnetic field are found by combining Maxwell's equations and Ohm's law, and solving the resulting diffusion equation:

$$\nabla^2\mathbf{B} = \mu_0\sigma\,\partial\mathbf{B}/\partial t, \tag{4}$$

where μ_0 is the vacuum permeability and σ is the conductivity. For nested spheres, Eq. (4) may be solved by separation of variables; the radial component of the solution is given by Bessel's equation and continuity of $\mathbf{B}$ across the spherical boundaries yields recursion formulae (Srivastava, 1966; Parkinson, 1983). For a three-shell spherical model where the conducting ocean is nested between an insulating interior and an insulating ice shell, A and ϕ are given by (Zimmer et al., 2000)

$$Ae^{i\phi} = \left(\frac{r_0}{r_m}\right)^3 \frac{RJ_{5/2}(r_0k) - J_{-5/2}(r_0k)}{RJ_{1/2}(r_0k) - J_{-1/2}(r_0k)}, \tag{5}$$

where

$$R = \frac{r_1 k J_{-5/2}(r_1k)}{3J_{3/2}(r_1k) - r_1kJ_{1/2}(r_1k)}. \tag{6}$$

Here the J_n denote Bessel functions of the first kind and order n. The radius of Europa, the radius to the ocean surface, and the radius to the ocean floor are symbolized by r_m, r_0, and r_1, respectively. The complex parameter k is a measure of the extent to which the time-varying field penetrates the conductor, here taken to be the ocean. The field oscillation frequency ω and conductivity σ determine k:

$$k = (1-i)\sqrt{\mu_0\sigma\omega/2}. \tag{7}$$

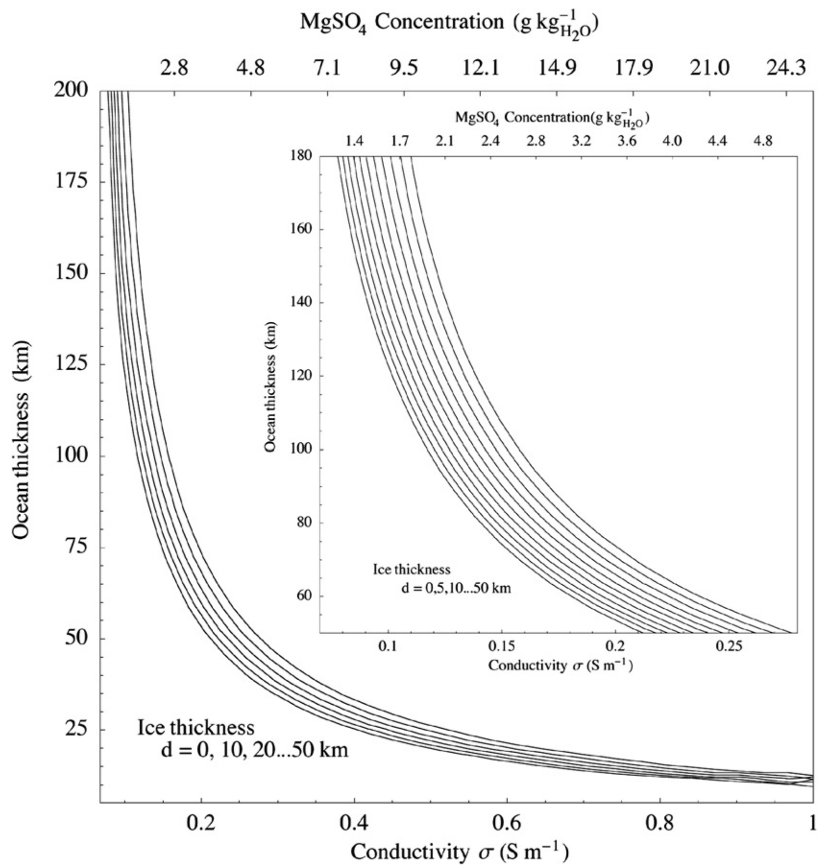


Fig. 2. Ocean depth–salt concentration relationship for an induced magnetic field amplitude of $A = 0.7$. Magnesium sulfate concentrations (g per kg water) for varying ocean thicknesses were calculated (see text) for a variety of ice shell thicknesses. The plot shows lines of constant ice thickness. Starting from the left, lines are given for no ice shell (0 km) on up to a 50 km ice shell, in 10 km increments. The inset plot details the 50–180 km ocean thickness region and shows lines of ice shell thickness ranging from 0 to 50 km in 5 km increments. All of the solutions shown in the inset plot would be considered freshwater, or nearly freshwater (i.e. oligohaline or mildly brackish), by Earth standards.

We calculate A for a range of ocean thicknesses, ice thicknesses, and ocean conductivities. In this first model, the ice shell and mantle are assumed to have zero conductivity; we solve for a conductive mantle later in this work. For small σ ($\leqslant 1\ \mathrm{S\,m^{-1}}$) we solve for A exactly. As σ increases, however, the terms in the numerator and denominator of Eq. (5) rapidly approach zero, creating a region of numerical instability. For large σ we use an approximation to Eq. (5) wherein the highest-order (smallest) terms are discarded. The full approximation is detailed in Appendix A; we demonstrate there that it is a good approximation for $\sigma \geqslant 0.1\ \mathrm{S\,m^{-1}}$.

Zimmer et al. (2000) concluded that induced field amplitudes of $A < 0.7$ are hard to reconcile with the *Galileo* observations; so in all cases we have solved for conductivity variations with $A \geqslant 0.7$. Furthermore, amplitudes for $A > 1.0$ imply significant conductivity outside the moon (e.g. ionosphere, cloud of pick-up ions) and Zimmer et al. (2000) argue that the observed magnetic signature does not fit such a model. We nevertheless address the possible contribution of an ionosphere in a subsequent section. More recent work using a least-squares fit to the magnetometer data (Schilling et al., 2004) concludes that $A = 0.97 \pm 0.02$. Here we examine how variations in the amplitude A, coupled with variations in the ice shell and ocean thicknesses, constrain the salt concentration in the putative ocean.

3.1. Amplitude response $A \leqslant 0.9$

First, we examine the low-end case of $A = 0.7$. Results are shown in Fig. 2. The range of ocean thicknesses shown in Fig. 2 covers the range allowed by the gravity data (Anderson et al., 1998). Lines of constant ice shell thickness are shown for shells ranging from 0 to 50 km thickness. For the case of a 100 km thick ocean with ice shells of different thicknesses, the resulting salt concentrations all fall below 2.7 g $MgSO_4\ \mathrm{kg}^{-1}_{H_2O}$. Note that even for the case of a 50 km ice shell overlying a 50 km ocean, the required salt concentration when $A = 0.7$ still falls below 5 g $MgSO_4\ \mathrm{kg}^{-1}_{H_2O}$.

Considering some of the cases permitted by the two extremes of the gravity data, we have calculated that a 160 km liquid water ocean beneath a 10 km ice shell would require a concentration of 1.3 g $MgSO_4\ \mathrm{kg}^{-1}_{H_2O}$ ($\sigma \sim 88\ \mathrm{mS\,m^{-1}}$). Similarly, for the lower-limit case of an 80 km hydrosphere, we find that a 60 km thick ice shell with a 20 km liquid water conducting layer at the bottom would require 15.6 g $MgSO_4\ \mathrm{kg}^{-1}_{H_2O}$ ($\sigma \sim 0.69\ \mathrm{S\,m^{-1}}$) whereas an ice shell of 10 km with a 70 km ocean requires only 2.8 g $MgSO_4\ \mathrm{kg}^{-1}_{H_2O}$ ($\sigma \sim 165\ \mathrm{mS\,m^{-1}}$). That is, for the case of $A = 0.7$, even the extreme choices permitted by the gravity data give low $MgSO_4$ concentrations, in the range 1 to 16 $\mathrm{g\,kg}^{-1}_{H_2O}$.

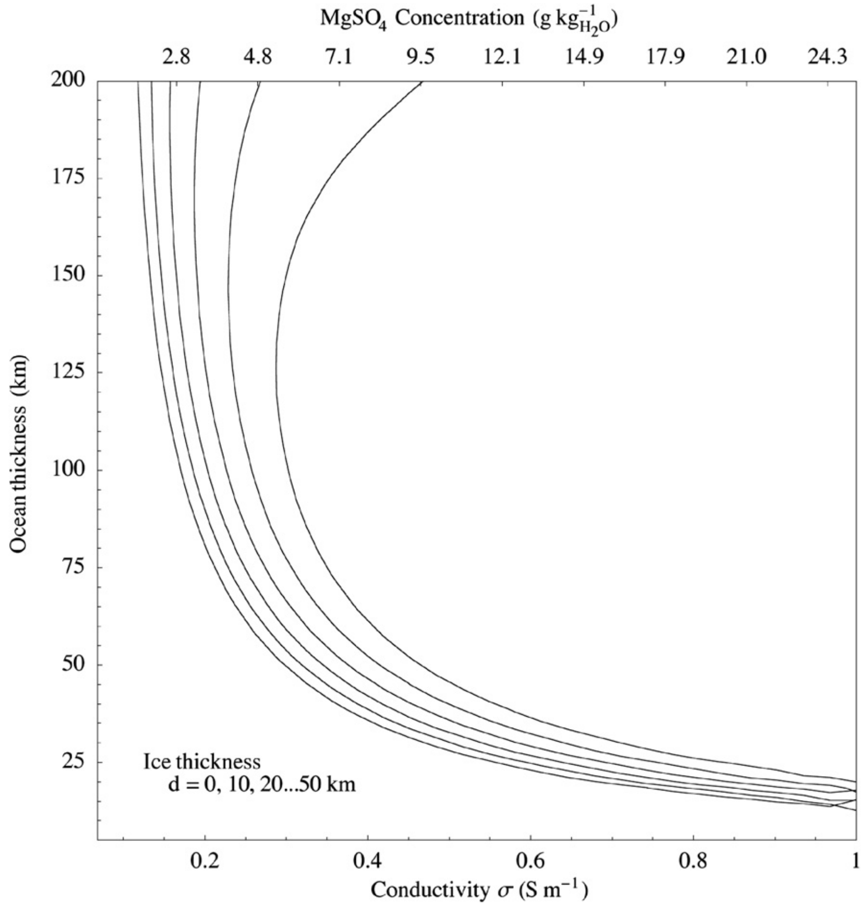


Fig. 3. Ocean depth–salinity relationship for an induced magnetic field amplitude of $A = 0.8$. Magnesium sulfate concentrations (g per kg water) for varying ocean thicknesses were calculated (see text) for a variety of ice shell thicknesses. The plot shows lines of constant ice thickness. Starting from the left, lines are given for no ice shell (0 km) on up to a 50 km ice shell, at 10 km increments.

As we increase the amplitude response, A, the conductive layer approaches a perfect conductor ($A = 1.0$). Solutions for $A = 0.8$ and $A = 0.9$ are shown in Figs. 3 and 4, respectively. Examining the case of a 10 km ice shell overlying a 100 km thick ocean, we find that for $A = 0.8$ the required salt concentration is 3.2 g $MgSO_4\,kg^{-1}_{H_2O}$ while for $A = 0.9$, the requirement increases to 10 g $MgSO_4\,kg^{-1}_{H_2O}$. Both values are still significantly lower than the sea salt concentration of the terrestrial ocean. Increasing the ice shell to 50 km thickness pushes the salt concentration up to 5.6 g $MgSO_4\,kg^{-1}_{H_2O}$ for the case of $A = 0.8$. Under the same circumstances, raising A to 0.9 requires increasing the conductivity to $\sim$175 $S\,m^{-1}$ (see Fig. 5). This value is far beyond the conductive capabilities of salt ions in water, so these parameter combinations are not permitted. We explore these issues in depth in subsequent sections.

3.2. *Amplitude response $A = 0.97 \pm 0.02$*

As A increases toward 1.0, salt concentrations must concomitantly increase. While Zimmer et al. (2000) argued that A must lie in the range $0.7 \leqslant A \leqslant 1.0$, more recent work by Schilling et al. (2004) suggests that $A = 0.97 \pm 0.02$. In Figs. 5 and 6 we extend the range of conductivities to include large values for σ and we show how the combination of ocean thickness, ice shell thickness, and conductivity determine the amplitude of the induced magnetic field. We explore the range of physically plausible conditions that could allow such a high value of A. Ultimately we show that because there is an upper limit on the conductivity of the ocean placed by the salt ion saturation, values of $A = 0.97 \pm 0.02$ are only possible if the ice shell is thin.

In Fig. 5, three models for the ocean thickness are considered: 80, 100, and 120 km, and in each case we examine the solutions for ice shell thicknesses of 0, 10, 30, and 50 km. At conductivities below 1.2 $S\,m^{-1}$ the ocean thickness plays an important role in determining the amplitude. However, beyond 1.2 $S\,m^{-1}$ the ice shell thickness dominates the amplitude response. As shown in Fig. 5, when $\sigma > 2\,S\,m^{-1}$ ice shells of greater thickness require higher oceanic conductivities in order to achieve the same amplitude response. As conductivity increases, the amplitude slowly approaches the idealized perfect conductor case of $A = 1.0$. Here we solve for values of σ up to 300 $S\,m^{-1}$ ($\sim$120 times the conductivity of Earth's ocean), at which point the maximum achievable amplitude response is $A = 0.99$. This is for no ice shell ($d = 0$), obviously not physically plausible. If we instead consider the case of $d = 10$ km, then at $\sigma = 300\,S\,m^{-1}$ we find $A = 0.97$. Similarly, for $d = 30$ and $d = 50$ km we find $A = 0.94$ and $A = 0.90$, respectively.

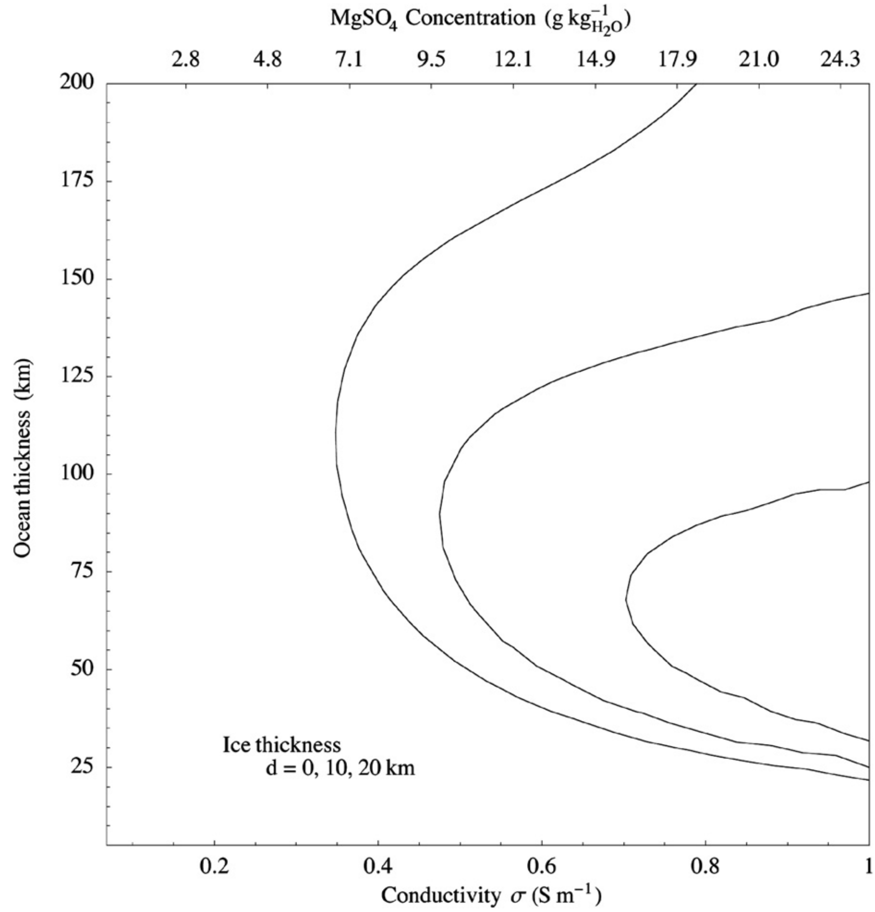
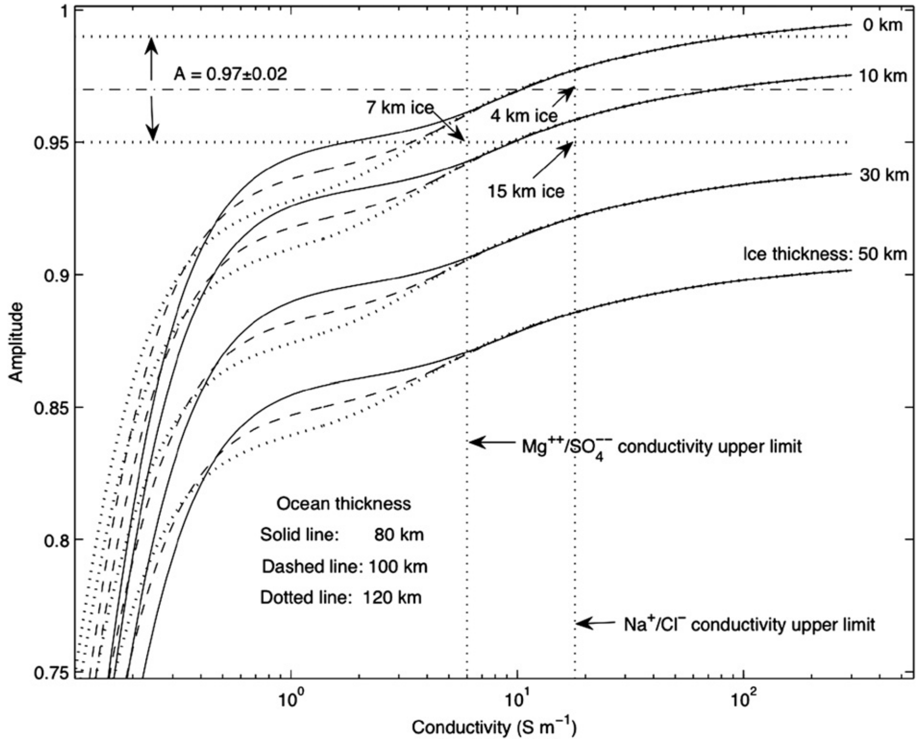


Fig. 4. Ocean depth–salt concentration relationship for an induced magnetic field amplitude of $A = 0.9$. Magnesium sulfate concentrations (g per kg water) for varying ocean thicknesses were calculated (see text) for a variety of ice shell thicknesses. The plot shows lines of constant ice thickness. Starting from the left, lines are given for no ice shell (0 km), a 10 km shell, and a 20 km shell. The double-valued shape of the curves is due to skin depth effects in amplitude and phase (see text).

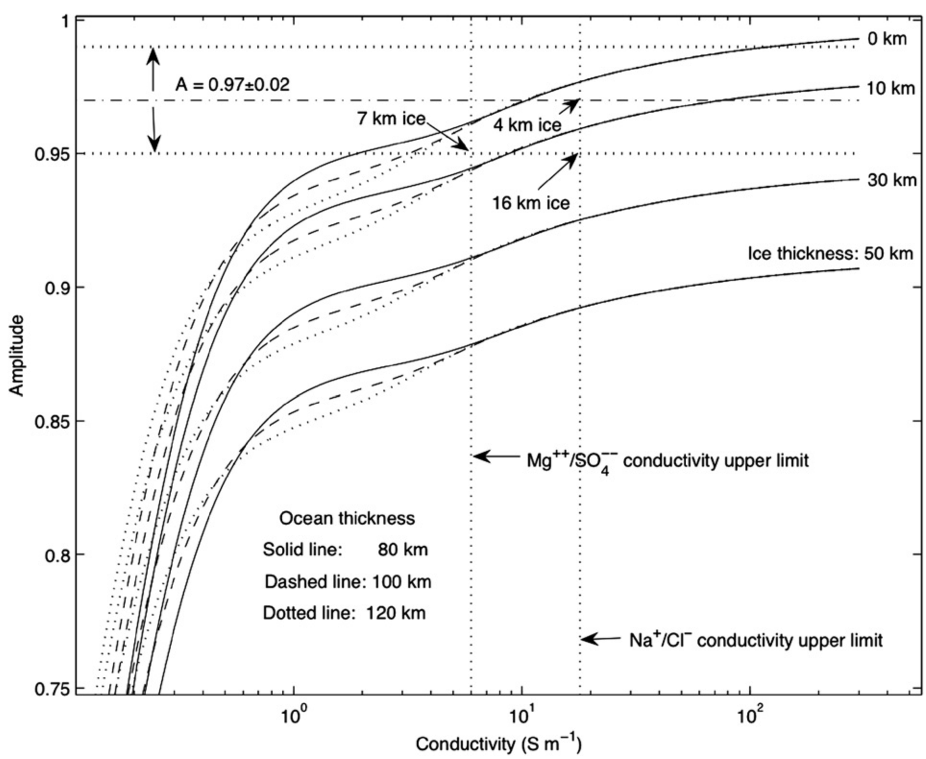
In practice there are physical limits on how much salt can be added to water and thus there are limits to how high the conductivity can be driven by the concentration of salt ions. Saturation for the dissolved ionic form of $MgSO_4$ occurs at $\sim$282 g $MgSO_4\,kg^{-1}_{H_2O}$ (Hogenboom et al., 1995; Kargel, 1991) and for NaCl saturation occurs at $\sim$304 g $NaCl\,kg^{-1}_{H_2O}$ at 25 °C (Oren, 2001). Hydrated forms of $MgSO_4$ may exist in suspension in the water (e.g. $MgSO_4{\cdot}7H_2O$), but these forms do not contribute to the conductivity of the ocean water. In addition, as the salt concentration increases the viscosity of the solution also increases, reducing the mobility of the ions and limiting the conductivity (Anderko and Lencka, 1997). At low temperatures, increasing pressure over the range 0–400 MPa has little effect on conductivity (Quist and Marshall, 1968). The $\sim$5–200 MPa range within Europa's ocean should thus cause little variation in conductivity. Hypersaline environments on Earth are rarely above a few Siemens per meter. For example, the brine of Mono Lake in Northern California is reported to be 8.57 $S\,m^{-1}$ (at 25 °C) (Jellison et al., 1999), and the brine saturated sediments of Lake Magadi in Southwest Kenya are measured to be $\sim$11 $S\,m^{-1}$ (34 °C) (Jones et al., 1998).

The saturation limits for salt concentration impose limits on the conductivity ($\sim$6 $S\,m^{-1}$ for $MgSO_4$ and $\sim$18 $S\,m^{-1}$ for sea salt) that are hard to reconcile with the high amplitude requirement of $A \sim 0.97$. In Fig. 5 we show the upper limit for conductivity of an $MgSO_4$ or NaCl solution. As is shown, the intersection of these lines with $A = 0.97$ yields an ice shell of thickness equal to, or less than, 4 km. Taking the lower end of the error bar, $A = 0.95$, we find that at most the ice shell is 7 km ($MgSO_4$) to 15 km (NaCl) thick. Even if a solution of conductivity 30 $S\,m^{-1}$ were physically plausible, we would still have upper limits on the ice shell thickness of $\sim$6 and $\sim$17 km for $A = 0.97$ and $A = 0.95$, respectively (for NaCl). Additionally, it should be noted that salt concentrations of $\geqslant$300 g per kg of H_2O would push the density of the solution beyond those permissible by the three-shell internal modeling results of Anderson et al. (1998). This density constraint is not terribly strict—alternative interior models fitting for the bulk density and moment of inertia could allow for higher density outer shells—but it does provide another piece of information pertaining to the bulk properties of putative ocean.

Does the above analysis demonstrate that Europa's ice shell must be $\leqslant$15 km? While Schilling et al. (2004) conclude that $A = 0.97 \pm 0.02$, this value is derived by comparing two different models for the external and internal magnetic fields in and around Europa. In the first case, Europa is allowed to have an

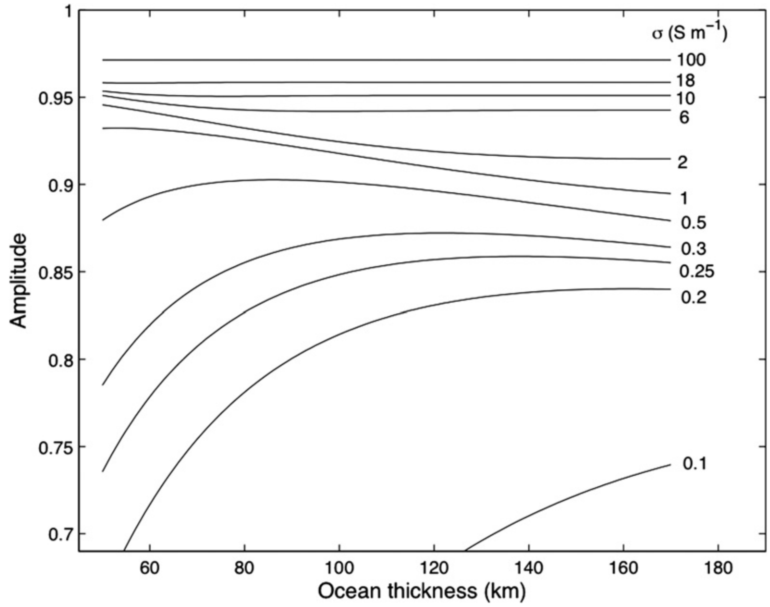


(a)

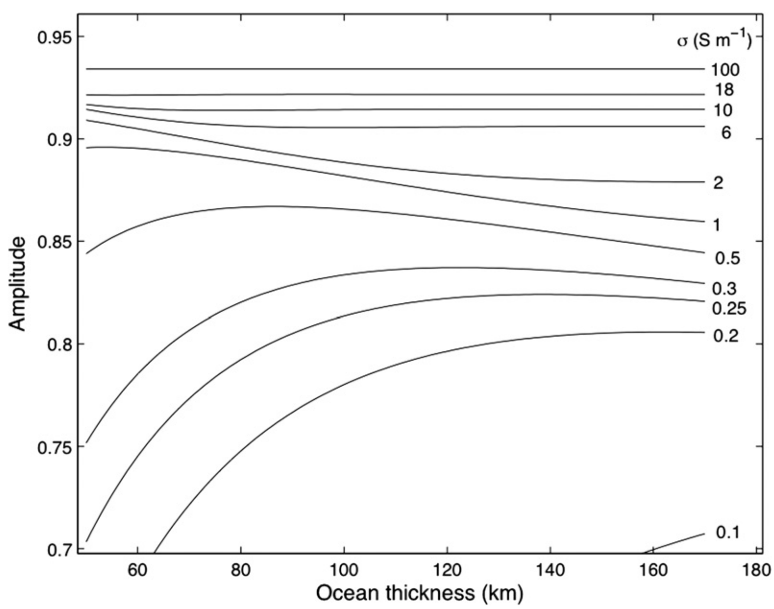


(b)

Fig. 5. Amplitude, A, as a function of conductivity, ocean thickness, and ice shell thickness. At low conductivities the ocean thickness dominates the amplitude response, however, as the conductivity of the solution increases, ocean thickness becomes less important and the thickness of the non-conducting ice shell dominates the inductive amplitude response. (a) Three-layer spherical model with non-conducting ice shell and mantle. The only solutions consistent with the magnetometer data [Schilling et al. (2004) require $A = 0.97 \pm 0.02$] and the gravity data (Anderson et al., 1998) are those in which the ice shell is thin (between 0 and 15 km thick). The optimal fit for an ocean dominated by NaCl is 4 km thickness. The maximum allowable ice shell thickness for an $MgSO_4$ dominated ocean is 7 km. (b) Five-layer half-space model. Results are very similar to the three-layer spherical model. Values for the conductivity of individual layers are: $\sigma_{\text{Ionosphere}} = 2 \times 10^{-4}\ \text{S m}^{-1}$; $\sigma_{\text{Ice}} = 1 \times 10^{-10}\ \text{S m}^{-1}$; $\sigma_{\text{Mantle}} = 0.01\ \text{S m}^{-1}$; $\sigma_{\text{Core}} = 3.3 \times 10^{5}\ \text{S m}^{-1}$. Again the optimal fit to the magnetometer data is achieved with a 4 km thick ice shell. The maximum thickness allowed by the error bars is 16 km, slightly more than calculated with the three-layer spherical model. This difference is primarily due to conductivity limits for the half-space model (see text).



(a)



(b)

Fig. 6. Contours of conductivity as a function of amplitude and ocean thickness. In (a), the ice thickness in all cases is assumed to be 10 km. In (b), the ice thickness in all cases is assumed to be 30 km.

internal permanent dipole moment with a surface magnitude of 23 nT and a tilt of 59° from the Europa's spin axis. From this fit to the magnetometer data, they find that $A = 0.96 \pm 0.03$. In other words, if we allow Europa to have an internal magnetic field, we can push the amplitude down to 0.93. Even for this low of a value for A, we find that the ice shell must be equal to, or less than 26 km thick. In their model without an internal dipole, Schilling et al. (2004) find that $A = 0.98 \pm 0.01$. Here the lower limit of $A = 0.97$ again leads us to an ice shell of $\leqslant 4$ km for the case of NaCl. At $A = 0.98$, the ice shell can be only meters thick. Ultimately, the work of Schilling et al. (2004) is data limited—only four *Galileo* passes of Europa were close enough to satisfy their criterion. However, in their rigorous analysis and fitting of models so as to minimize root-mean square fits to the data, they conclude that $A = 0.97 \pm 0.02$, taking all models into account. In this case, the ice must indeed be $\leqslant 15$ km.

In Fig. 6 we show how varying the ocean thickness affects the amplitude response. Contours for the saturation limits on conductivity, along with several other conductivity contours of interest are shown. In Fig. 6a the ice shell thickness is 10 km, whereas in Fig. 6b the ice shell is 30 km thick. These contours reveal that for conductivities of a few Siemens per meter the inductive response actually decreases as we increase the thickness of the conducting layer (i.e. the ocean thickness). This is likely a skin depth effect, and can be understood semi-quantitatively in the following way. In the case of a plane conductor, the solution to Eq. (4) is simply (e.g. Parkinson, 1983):

$$B = B_0 e^{-z/\delta} e^{-i(\omega t - z/\delta)}, \tag{8}$$

where $z = 0$ at the boundary of the conductor and increases positively into the conductor, and

$$\delta = (\mu_0 \sigma \omega / 2)^{-1/2} \tag{9}$$

is the skin depth. For every increment δ of depth z into the conductor, the amplitude decreases by a factor of e and the phase changes by one radian. Therefore, increasingly deep layers of the conductor both contribute less to the amplitude, and contribute increasingly destructively (out of phase). For $\sigma = 1$ and $10\ \mathrm{S\,m^{-1}}$, $\delta = 101$ and 32 km, respectively, consistent with the behavior seen in Figs. 3–6.

Of course, a nested spherical shell model for Europa requires that Eq. (4) be solved in spherical coordinates, and the solution is Eq. (5), not Eq. (8). However, Eqs. (8) and (9) will provide a good local approximation for the solution into the spherical conductor provided that (Srivastava, 1966)

$$\sigma \gg 2/(\mu_0 \omega r^2). \tag{10}$$

In this case, the real part of $kr \gg 1$, allowing us to drop the first derivative term in the Bessel equation solution to the radial part of Eq. (4), and the system has solutions given by Eqs. (8) and (9) (Srivastava, 1966). For Europa, $r = 1560$ km and $\omega = 2\pi/P = 1.6 \times 10^{-4}$, where $P = 4.03 \times 10^4\ \mathrm{s} = 11.2$ h is Jupiter's synodic rotation period. Equation (10) will therefore hold when $\sigma \gg 10^{-3}\ \mathrm{S\,m^{-1}}$, which will be the case for the conductivities of interest here. Therefore Eqs. (8) and (9) provide good insight into the decay and phase change of the magnetic field as it extends down into Europa's ocean.

For certain combinations of σ and ocean thickness, the induced response at depth will cancel part of the induced response generated near the surface; this behavior explains, e.g., the shape of the curves in Fig. 6. For large conductivity the skin depth goes to zero—all of the induced response is generated near the surface—and cancellation effects are avoided. A curious consequence of this behavior is that if the salinity of the putative europan ocean is comparable to that of the Earth's ocean (with σ of a few $\mathrm{S\,m^{-1}}$), then thinner oceans provide a better fit to the high amplitude requirement of Schilling et al. (2004).

Table 1 summarizes the relationship between layers on our 3-layer model, providing values of A for a variety of plausible combinations of ice thickness, ocean thickness, and conductivity. Again, A can only exceed 0.95 for thin ice shells and near-saturation salt concentrations, regardless of which scenario is considered.

4. Adding a conducting core, mantle, and ionosphere

We have shown that the three-layer model can only fit the magnetometer amplitude constraints if the conductivity is high and the ice shell thin. Here we explore the influence of a conducting ionosphere, mantle, and core on the induced field response. We use a plane stratified half-space (Parkinson, 1983; Srivastava, 1965) instead of a spherical model. As previously shown, this model is a good approximation for Europa for conductivities greater than approximately $10^{-3}\ \mathrm{S\,m^{-1}}$. This leads to a planar approximation with m shells and a recursive expression for the ratio of the induced to external fields given by

$$\frac{B_i}{B_e} = \frac{R_1(\alpha_1 - \nu) - \alpha_1 - \nu}{R_1(\alpha_1 + \nu) - \alpha_1 + \nu}, \tag{11}$$

where

$$R_j = e^{2\alpha_j z_j} \frac{R_{j+1}(\alpha_j + \alpha_{j+1}) + (\alpha_j - \alpha_{j+1}) e^{2\alpha_{j+1} z_j}}{R_{j+1}(\alpha_j - \alpha_{j+1}) + (\alpha_j + \alpha_{j+1}) e^{2\alpha_{j+1} z_j}}, \tag{12}$$

and the recursion is initiated by

$$R_{m-1} = e^{2\alpha_{m-1} z_{m-1}} \frac{(\alpha_{m-1} + \alpha_m)}{(\alpha_{m-1} - \alpha_m)}. \tag{13}$$

Here $\alpha_m = \sqrt{\nu^2 + k_m^2}$, where k_m is given by (7) and ν is a constant resulting from separation of variables when solving the half-space model (Price, 1962). Physically, the value $2\pi/\nu$ is a measure of the horizontal scale and uniformity of the source field (Srivastava, 1965; Price, 1962), with $\nu = 0$ representing no spatial variation in the field (Cagniard, 1953). Price (1962) however showed that even very small values of ν are important for accurate modeling. For the case of the Earth, Price finds the smallest permissible value of ν to be found by setting $2\pi/\nu$ equal to the circumference of the Earth. The largest value is found by setting $2\pi/\nu$ equal to a few times the height of the terrestrial ionosphere. Following Srivastava (1965) we take $\nu = [n(n+1)]^{1/2}/r_{\mathrm{Europa}}$, where n is the order of the Bessel function, so $n = 1$ for the external dipole field. This leads to a good match with the 3-layer spherical model, though for thick ice shells (>30 km) the results yield slightly higher values for the amplitude. This is in part due to the fact that adding the very low-conductivity ice layer means adding a layer that does not satisfy the conductivity criteria for the half-space model. The net effect is that our 5-layer model over-estimates the amplitude response and consequently over-estimates the ice shell thickness. Even so, the 5-layer model still predicts an ice shell of <16 km, with a best-fit at 4 km. Fig. 5b shows these results.

In Fig. 7 we show amplitude profiles for several different combinations of ionosphere, mantle, and core conductivity. For ocean conductivities greater than $\sim 3\ \mathrm{S\,m^{-1}}$, the contribution from these additional layers is negligible. Below this value, the core and mantle are seen to have a strong influence on amplitude. The effect of induction in the core and mantle is primarily deconstructive interference with induction in the ocean. The result is that for a given ocean conductivity, a conductive core (and/or mantle) lowers the amplitude response of the total induced field. Values for both layer conductivity and layer thickness were varied based on published estimates (Zimmer et al., 2000; Saur et al., 1998; Parkinson, 1983; Petrenko and Whitworth, 1999; Anderson et al., 1998; Stacey, 1992). Though we have shown results for mantle conductivity of $200\ \mathrm{S\,m^{-1}}$ (dash-dotted line) this is a very high-end value that would likely only persist for the upper few kilometers of oceanic crust.

Table 1
Induced amplitude response

Ocean thickness (km)	Ice thickness (km)	Amplitude of induced magnetic field (A)					
		Salt concentration (g salt $kg^{-1}_{H_2O}$) [conductivity, $S\,m^{-1}$]					
		1.14 [0.072]	5.0 [0.25]	100 [3.0]	282 [6.0]	304 [18.0]	350 [23.0]
60	0	0.38	0.80	0.96	0.97	0.98	0.98
	10	0.37	0.78	0.94	0.95	0.96	0.96
	30	0.35	0.75	0.91	0.91	0.92	0.92
	50	0.33	0.71	0.87	0.88	0.89	0.89
80	0	0.46	0.84	0.95	0.96	0.98	0.98
	10	0.45	0.83	0.94	0.94	0.96	0.96
	30	0.43	0.79	0.90	0.91	0.92	0.92
	50	0.41	0.76	0.86	0.87	0.87	0.89
100	0	0.53	0.87	0.95	0.96	0.98	0.98
	10	0.52	0.85	0.93	0.94	0.96	0.96
	30	0.49	0.81	0.89	0.91	0.92	0.92
	50	0.47	0.78	0.86	0.87	0.89	0.89
120	0	0.59	0.88	0.95	0.96	0.98	0.98
	10	0.57	0.86	0.93	0.94	0.96	0.96
	30	0.54	0.82	0.89	0.91	0.92	0.92
	50	0.52	0.79	0.86	0.87	0.89	0.89
150	0	0.64	0.88	0.95	0.96	0.98	0.98
	10	0.63	0.86	0.93	0.94	0.96	0.96
	30	0.60	0.82	0.89	0.91	0.92	0.92
	50	0.57	0.79	0.86	0.87	0.89	0.89

Note. Induced amplitude response, A, is provided for a variety of $MgSO_4$ concentrations, ocean thicknesses, and ice shell thicknesses. The salt concentrations shown here correspond to: (1) the low-end conductivity limit of Zimmer et al. (2000); (2) the low-end conductivity limit of Schilling and Neubauer (2005); (3) the high-end estimate for $MgSO_4$ concentration of McKinnon and Zolensky (2003); (4) the ionic saturation concentration for $MgSO_4$; (5) the saturation concentration of NaCl; and (6) the density limit for the water layer on Europa (Anderson et al., 1998). Errors in the salt concentration, primarily resulting from the polynomial fit, are on the order of ± 0.1 g $MgSO_4$ $kg^{-1}_{H_2O}$.

Measurements of terrestrial mantle conductivity suggest that a ∼600 km thick mantle would likely have a bulk conductivity in the range of 0.01 $S\,m^{-1}$ (solid line) (Parkinson, 1983; Stacey, 1992). Comparisons with electrical properties of chondrites indicate that such conductivities are also compatible with a mantle of chondritic origin (Campbell and Ulrichs, 1969). Thus, as is shown in Fig. 7, we expect that on Europa the amplitude response at low ocean conductivities will be dominated by the core. In all cases, we find that adding conductive mantle and core layers to our model does not significantly influence the amplitude response in the region where $A = 0.97 \pm 0.02$.

The case of a conducting ionosphere is treated separately. First we note that in order to properly compare the amplitude response with and without an ionosphere, we must scale the models with an ionosphere by a factor of $\sim(r_0/r_m)^3$. Here r_0 is the total thickness and r_m is the thickness without the ionosphere. This scaling normalizes A for thickness variations between the models, and in so doing reveals amplitude differences resulting from ionospheric conductivity. The scaling factor allows us to compare with published values of A, all of which are calculated for B_{ind}/B_{ext} at the surface of Europa. The cube of the ratio accounts for the $1/r^3$ change in the external and induced fields as one moves from the ice surface to the top of the ionosphere. The scaling factor is chosen such that at zero ionospheric conductivity, the ratio of the amplitudes for models with and without an ionosphere is one.

Estimates for ionospheric conductivity range from $<5 \times 10^{-5}$ to 2×10^{-4} $S\,m^{-1}$ (Khurana et al., 2002; Saur et al., 1998; Zimmer et al., 2000). We take 300 km as our nominal ionosphere thickness (Zimmer et al., 2000), but we find that this parameter can be changed without impacting the results, provided the conductivity is constrained to the range above.

Fig. 8 shows that even with the high-end value of 2×10^{-4} $S\,m^{-1}$ for ionospheric conductivity, there is negligible effect on the amplitude response. Only when the conductivity is approximately two orders of magnitude higher do we see significant changes in the amplitude response. In Fig. 9 we show this relationship by plotting the conductivity of the ionosphere versus the ratio of the amplitudes for a model with an ionosphere to that without a conducting ionosphere. At high conductivities for the ionosphere, the amplitude ratio becomes greater than one, indicating that the ionosphere could drive the amplitude response to very high values if such conductivities were physically plausible. Only when the ocean conductivity is very low (e.g. 0.1 $S\,m^{-1}$) do we begin to see some influence of the ionosphere at realistic values for ionospheric conductivity. Even then, the net effect does not help us resolve the issue of the observed high values for the amplitude response. At very low ocean conductivities a conducting ionosphere can influence the amplitude response, but since the ocean conductivity is so low, the amplitude response is limited to values below $A \sim 0.7$. In other words, adding a conductive ionosphere could alter our in-

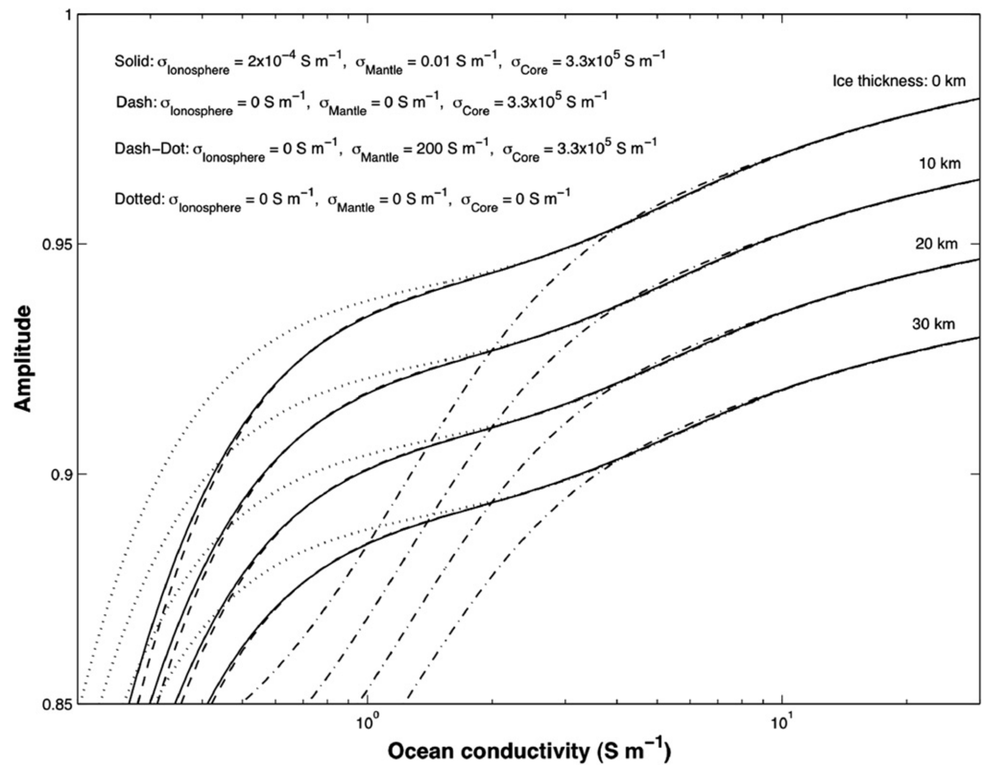


Fig. 7. Adding a conducting core and mantle influences the total amplitude response when the ocean conductivity is small. Even for models with a very conducting core and mantle, the amplitude in the region $A \geqslant 0.95$ is dominated by the conductivity of the ocean and the ice shell thickness.

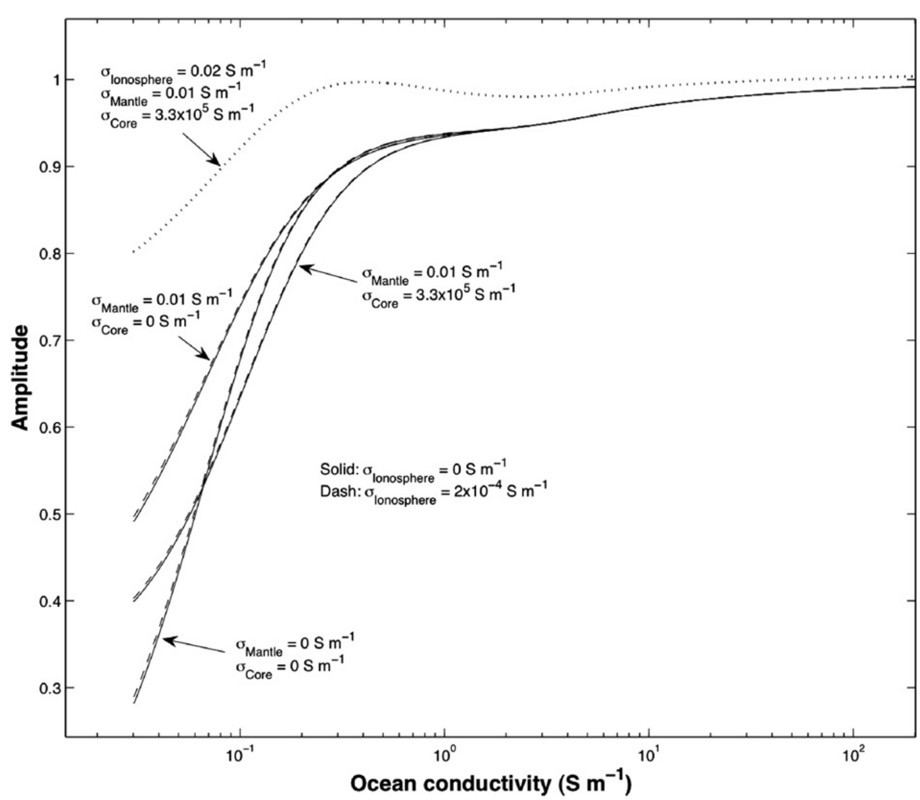


Fig. 8. Adding an ionosphere to our model results in little change to the amplitude profile for all reasonable values of ionospheric conductivity ($\leqslant 2 \times 10^{-4}\ \mathrm{S\,m^{-1}}$). The dashed lines show results for models with an ionosphere of conductivity $2 \times 10^{-4}\ \mathrm{S\,m^{-1}}$ and the solid lines show results for an ionosphere of zero conductivity. As shown by the dotted line, it is only when the ionosphere reaches physically implausible levels of conductivity that an influence on the amplitude is observed. In all cases the ionosphere is 300 km thick, the ice has zero thickness, and the ocean is 100 km thick. The amplitude has been normalized for comparison to published values of A (see text).

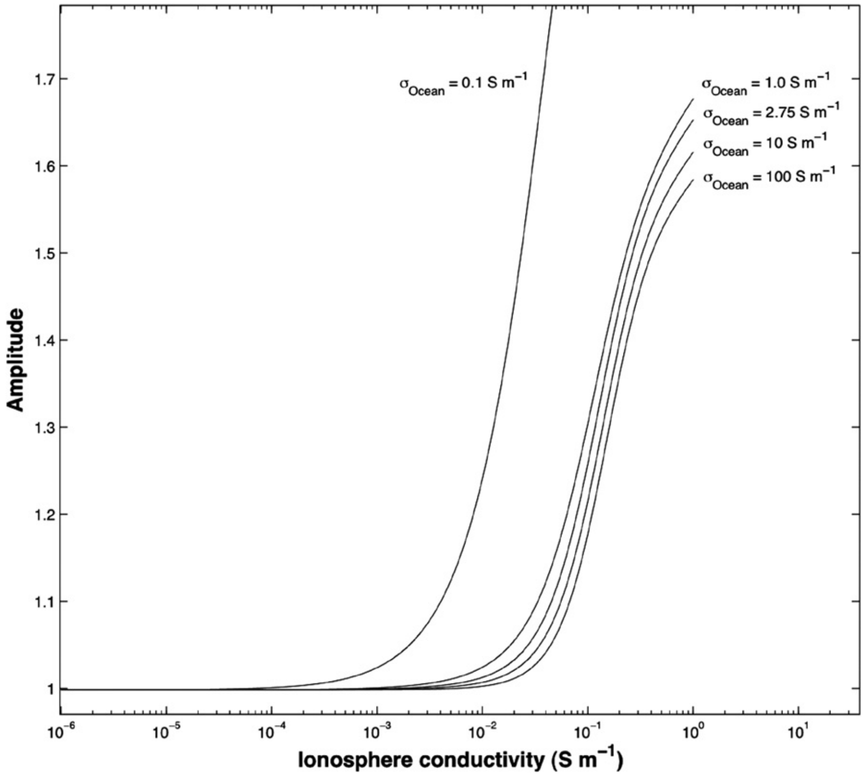


Fig. 9. Comparison of amplitudes, A, for models with and without conducting ionosphere. Here we define the ratio of the amplitudes as A_{IC}/A_{I0}, where A_{IC} is the amplitude for a model with a conducting ionosphere and A_{I0} is the amplitude for a model without a conducting ionosphere. We then plot this ratio as a function of ionospheric conductivity for several different choices of ocean conductivity. In all cases the ratio is very close to unity for reasonable choices of ionospheric conductivity ($\leqslant 2 \times 10^{-4}\ \mathrm{S\,m^{-1}}$). The ionosphere shows a stronger effect for small values of ocean conductivity; however, such low values for the ocean conductivity are not consistent with the empirical constraints on the observed amplitude (Zimmer et al., 2000; Schilling et al., 2004).

terpretation of the broad range of allowed amplitudes, $0.7 < A < 1.0$ (Zimmer et al., 2000), but it has no influence when considering the much tighter constraints, $A = 0.97 \pm 0.02$, of Schilling et al. (2004).

In summary, even when we add a conducting core, mantle, and ionosphere, we find that the critical factors for achieving high values of A are the ocean conductivity and ice shell thickness.

5. Implications for ice shell thickness

We have shown that for the lower end limits of the Zimmer et al. (2000) model ($0.7 < A < 0.9$) many of the most plausible solutions yield oceanic salt concentrations of less than ∼15 g $MgSO_4\ kg^{-1}_{H_2O}$, and solutions of <3 $MgSO_4\ kg^{-1}_{H_2O}$ are entirely possible. In other words, by terrestrial standards, the putative ocean of Europa could be much less saline than our ocean and for some solutions it could even qualify as a freshwater ocean.

However, given the interplay of the conductivity, ocean thickness, and ice shell thickness in this model, we find that in order to satisfy the *Galileo*-constrained amplitude response of $A = 0.97 \pm 0.02$ (Schilling et al., 2004), the ice shell must be $\leqslant$15 km thick, irrespective of ocean thickness. For an ice shell of near-zero conductivity ($\sigma_{Ice} = 1 \times 10^{-10}\ \mathrm{S\,m^{-1}}$), we find that an ice shell thickness of ∼4 km—comparable to that of the Antarctic ice sheet—best satisfies the $A = 0.97$ requirement. Even if we account for changes in the ice shell conductivity resulting from variations in temperature with depth, allowing the conductivity to reach $\sim 10^{-4}\ \mathrm{S\,m^{-1}}$ (Addison, 1969; Moore et al., 1994), we find no change in our results.

For a three-shell gravity model, the combined ocean plus ice thickness must be greater than 80 km (Anderson et al., 1998), thus implying an ocean at least 70 km thick. Trading liquid water for ice (e.g. 50 km water and 30 km ice) results in a poor amplitude response, one only satisfied by the lower end fits to the *Galileo* magnetometer data (Zimmer et al., 2000). If one considers the $MgSO_4$ constraints suggested by McKinnon and Zolensky (2003) of <100 g $MgSO_4\ kg^{-1}_{H_2O}$, it is then impossible to satisfy the $A \sim 0.97$ constraint of Schilling et al. (2004) with the model of Zimmer et al. (2000). At 100 g $MgSO_4\ kg^{-1}_{H_2O}$ ($\sim 3.0\ \mathrm{S\,m^{-1}}$) the largest amplitude response is $A = 0.94$ when the ice shell thickness is set to zero.

An upper limit on salinity is set by the saturation limits of the various salt ions. With Mg^{2+} and SO_4^{2-} as the primary cation and anion, saturation is reached at approximately 282 g $MgSO_4\ kg^{-1}_{H_2O}$ in Europa's ocean. If more $MgSO_4$ is added it forms a hydrate and does not dissolve. Models of plume currents in the europan ocean indicate that such material is unlikely to remain in suspension and will most likely precipitate to the ocean floor (Goodman et al., 2004). Consequently, adding more than 282 g $MgSO_4\ kg^{-1}_{H_2O}$ does not help explain the induced magnetic field response of Europa. The conductivity can be pushed higher by considering sea salt (e.g. Na^+ and Cl^-), but even then the conductivity of the solution will not exceed tens of Siemens per meter. As a result, invoking a salty ocean to explain the observed induced magnetic field fails unless the ice shell is thin or the amplitude is less than that determined by Schilling et al. (2004). In order to have a thick ice shell ($\geqslant$30 km), the induced amplitude must be less than 0.92, contradictory to the results of Schilling et al. (2004).

6. Implications for habitability

If the ice and liquid water layers on Europa fall within the limits of Fig. 2 ($A = 0.7$) then, by standard definitions of "freshwater" environments on Earth [broadly meaning <3 g salt $kg^{-1}_{H_2O}$ (Barlow, 2003)], Europa's ocean would be a freshwater ocean, though admittedly more salty than most terrestrial lakes. Indeed, in this case, the putative global ocean of Europa could be more like the mildly saline environment of Pyramid Lake, Nevada than like the Earth's ocean. While the drinking water regulations of the U.S. Environmental Protection Agency recommend no more than 0.25 g of sulfate per kilogram of water, adult humans can acclimatize to drinking water with nearly 2 g $MgSO_4$ $kg^{-1}_{H_2O}$ without much discomfort (EPA, 2004; CDC-EPA, 1999). Animal toxicity (the lethal dose for 50% of the population) is in the range of 6 g $MgSO_4$ $kg^{-1}_{H_2O}$ (CDC-EPA, 1999), but most livestock are satisfied provided the total salt concentration is less than 5 grams per kilogram of water (ESB-NAS, 1972). If we assume the low amplitude regime for our solution ($A < 0.8$) then it is possible that human or beast could drink the water of Europa.

What are the implications for habitability for $A = 0.97$? Terrestrial halophilic microorganisms are capable of surviving at NaCl saturation (Oren, 1994, 2002). Among these are microbes from each domain of life (*Archaea*, *Bacteria*, and *Eucarya*). Metabolic pathways for these microbes include oxygenic and anoxygenic photosynthesis (*Dunaliella salina, Halorhodospira halphila*), aerobic respiration (*Halobacterium salinarum*), and fermentation (*Halobacterium salinarum*). Obviously, photosynthesis is an unlikely metabolic pathway if the ice is more than a few tens of meters thick, but respiration and fermentation could be possible, especially if radiolytically produced oxidants are delivered to the sub-surface (Chyba, 2000; Chyba and Hand, 2001). Methanogens—sometimes considered a plausible model for europan life (McCollom, 1999)—are capable of surviving in solutions near NaCl saturation if methanol or methylated amines are available (Oren, 2001). Data for $MgSO_4$ tolerance is limited, but microbes such as *Halobacterium sodomense* are known to survive in solutions above 2 M Mg^{2+} (equivalent to $\sim$260 g $MgSO_4$ $kg^{-1}_{H_2O}$) (Oren, 1994). Thus on Europa, where the high amplitude constraint necessitates a high salt concentration, the habitability of the ocean may be limited but the conditions would not appear to exclude life as we know it.

But habitability for life is not the same as suitability for the origin of life. Experimental investigations of the influence of ionic inorganic solutes on self-assembly of monocarboxylic acid vesicles, and the nonenzymatic, nontemplated polymerization of activated RNA monomers may support the contention that life originated in a freshwater solution (Monnard et al., 2002). In these prebiotic simulation experiments, sodium chloride or sea salt concentrations as low as 25 mM NaCl (1.5 g per kg of H_2O) were found to "substantially" reduce oligomerization, with higher concentrations having worse effects. But these experiments have not yet been performed with $MgSO_4$. With regard to vesicle formation, if the ratio of the cation to amphiphile falls below $\sim$1, then it becomes possible for the excess amphiphile to form membranes; above $\sim$2 and the amphiphile precipitates and no membrane formation occurs. The Monnard et al.'s (2002) results therefore suggest that the upper range of salinities implied by the *Galileo* magnetometer experiments could pose a serious challenge to the origin of life in Europa's bulk contemporary ocean, were abiotic RNA oligomerization or amphiphile membrane formation on the critical path to the origin of life, and if the experimental results for NaCl carry over to $MgSO_4$.

7. Conclusion

Aside from surface imagery, the magnetometer results are the only dataset that provides information about the relationship between the ice shell and putative subsurface ocean. In the analyses presented here we have provided empirical constraints on both the salinity of the europan ocean and the overlying ice shell thickness. By the low-end, poorly constrained analysis where $0.7 \leqslant A \leqslant 1.0$ (Zimmer et al., 2000) we find that a freshwater ocean is possible on Europa. By the tighter Schilling et al. (2004) constraints of $A = 0.97 \pm 0.02$, our results show that an ice shell of thickness $\leqslant$4 km, overlying a very salty ocean, is the best fit to the data. Thicknesses ranging from 0 to 15 km are permissible if one allows for the ± 0.02 uncertainties in the magnetic field signature. These results apply to present day Europa and are independent of any geological interpretation of surface features.

While our work provides new insight into the nature of the ice shell and ocean, we note two limiting factors in our models, the first of which is the focus of subsequent studies. First, we have assumed that the bulk ocean temperature is $\sim$273 K. For a given salt concentration, higher temperatures would mean higher conductivity. If the ocean is stratified and contains layers of significant thickness with temperatures slightly higher than 273 K, then such layers would have slightly higher conductivity. Using this relationship, a magnetometer on a future spacecraft mission could potentially help determine ocean structure and temperature profile.

The second limiting factor is simply the constraint on the amplitude. With additional analysis of the *Galileo* data, some greater resolution on this issue may be achieved. However, final answers to the questions of Europa's ice shell thickness, ocean chemistry, and habitability, await a future spacecraft mission.

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Appendix A

The approximation used to calculate the amplitude, A, for large values of conductivity, σ, was derived by eliminating terms of order x^2 and higher in the denominator of Eq. (5), where $x = r_0 k$ or $r_1 k$. With this approximation, Eq. (5) simplifies to

$$Ae^{i\phi} \approx \left(\frac{r_0}{r_m}\right)^3 \left(\frac{r_1}{\left(r_0 - \left(\frac{r_0 r_1 k}{3}\right)\tan(k(r_0 - r_1))\right)} - 1\right) + O\left(\frac{1}{x^2}\right). \tag{A.1}$$

All variables are as defined in the text. Using Eq. (7) and an 11.2-h synodic period for the primary field oscillation, we find

$$|x| = r_0|k| \approx r_1|k| \approx 10\sigma^{1/2}, \tag{A.2}$$

with σ in $\mathrm{S\,m^{-1}}$. Therefore $O(1/x^2) \approx 10^{-2}\sigma^{-1}$ (σ in $\mathrm{S\,m^{-1}}$), and Eq. (A.1) is a good approximation to Eq. (5) provided $\sigma > 0.1\ \mathrm{S\,m^{-1}}$. Explicit numerical ratioing of A from Eq. (A.1) to A from Eq. (5) verifies this analytical prediction. The solution for A is found by expanding Eq. (A.1) into real and imaginary components, and then eliminating ϕ in order to solve for A. Figs. 2–4 were plotted using the exact solution for A, whereas in Fig. 5 we use the above approximation.

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