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Research Paper

Searching for Liquid Water in Europa by Using Surface Observatories

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ABSTRACT

Liquid water, as far as we know, is an indispensable ingredient of life. Therefore, locating reservoirs of liquid water in extraterrestrial bodies is a necessary prerequisite to searching for life. Recent geological and geophysical observations from the Galileo spacecraft, though not unambiguous, hint at the possibility of a subsurface ocean in the Jovian moon Europa. After summarizing present evidence for liquid water in Europa, we show that electromagnetic and seismic observations made from as few as two surface observatories comprising a magnetometer and a seismometer offer the best hope of unambiguous characterization of the three-dimensional structure of the ocean and the deeper interior of this icy moon. The observatories would also help us infer the composition of the icy crust and the ocean water. Key Words: Europa—Ocean—Galileo—Liquid water—Search for life—Electromagnetic induction—Seismic exploration. Astrobiology 2, 93–103.

INTRODUCTION

GRAVITY MEASUREMENTS using Doppler data from Galileo (Anderson *et al.*, 1998) show that Europa's axial moment of inertia normalized to MR_E^2 (M is the mass of Europa and R_E = 1,565 km is its radius) is 0.346 ± 0.005, a value substantially less than the 0.4 expected of a uniform sphere. The low moment of inertia implies that Europa's interior regions are denser than its outer layers. Anderson *et al.* (1998) show that the most plausible models of Europa's interior have an H_2O layer of thickness 80–170 km overlying a rocky mantle. The models favor a metallic core, but its size is uncertain because the composition of the core material is unknown.

The physical state of the H_2O layer is also uncertain. Ever since Cassen *et al.* (1979, 1980) showed that tidal stressing is a significant internal heat source for Europa, the possibility of a liquid ocean in its H_2O crust has been hotly debated. Images from Voyager spacecraft (Smith *et al.*, 1979) showed that Europa is very sparsely cratered, indicating that its surface is fairly young, possibly in the range of a few tens of millions of years (Shoemaker and Wolfe, 1982; Zahnle *et al.*, 1998; Levison *et al.*, 2000). Many models put forward to explain the rejuvenation of Europa's surface by extrusions of long lineaments (Greeley *et al.*, 1998), chaotic terrain (Carr *et al.*, 1998), and lenticulae (Pappalardo *et al.*, 1998) involve a liquid ocean or warm ductile ice to provide low viscosity at the base of the icy crust. By studying the age relationships of linea-

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ments, Geissler *et al.* (1998) show that the orientations of the lineaments have changed in a manner consistent with the nonsynchronous rotation of an ice shell decoupled from the mantle by the action of a liquid ocean or a warm ductile ice shell.

In a careful reevaluation of all of the geological and geophysical evidence presented so far, Pappalardo *et al.* (1999) state "An internal ocean would be a simple and comprehensive explanation for a broad range of observations; however, we cannot rule out the possibility that all of the surface morphologies could be due to processes in warm, soft ice with only localized or partial melting." Thus, at this time, further observations are needed to provide direct verification of a subsurface global ocean. To date, four different approaches have been put forward to verify directly the existence of a subsurface liquid ocean.

The most direct approach is an ice-penetrating radar mounted on an Europa orbiting spacecraft (Chyba *et al.*, 1998). Detailed modeling studies (Chyba *et al.*, 1998; Moore and Schubert, 2000) indicate that a 20-W radar operating at 50 MHz could provide sounding to a depth of ~10 km in cold clean ice at the ambient surface temperature of Europa (100–150K) but may be able to sound through only 2 km if the ice is warm (~250K) or if a significant amount of salt-like impurities exist (>10% by volume). The radar technique would not be able to determine the thickness of the ocean because a 50-MHz electromagnetic wave attenuates by an e-folding in <1 m in oceanic water with a conductivity of 1 S/m.

Another promising technique would use the variation of the Europan surface tides to characterize the interior. Modeling results (Moore and Schubert, 2000) show that if a global subsurface ocean exists, the surface of Europa would flex with an amplitude of ~30 m. However, if Europa has frozen solid the amplitude of the temporal tide would be only ~1 m. A laser altimeter on the proposed Europa Orbiter could make such an experiment possible.

The next proposed technique also uses electromagnetic waves but with periods as long or substantially longer than the rotation period of Jupiter (~10 h). The passive induction technique uses the rotating and changing field of Jupiter's magnetosphere as a primary signal. The induction measurements can be made either from an orbiting spacecraft or from at least two observatories located on the surface of Europa. Multifrequency induction sounding of Europa as discussed below could provide independent estimates of conductivity and ocean thickness.

Finally, Kovach and Chyba (2001) have proposed that one or more seismometers placed on the surface of Europa could passively sound the interior of Europa by using ambient seismic noise as signal and detecting frequency dispersion in surface waves. If long-period seismograms are available, the body waves detected at multiple stations could be used to characterize the ocean (i.e., determine its thickness, composition, and temperature) and provide information on the size and density of the Europan core.

In this paper we show that seismic and magnetic measurements made from as few as two surface observatories would provide a platform not only for directly detecting the ocean but also for providing constraints on the ocean's thickness and the deeper interior of Europa. In the following sections, we first provide some basic information on the principles behind these measurements. Next, we provide the evidence from Galileo of the presence of a strong electromagnetic induction produced by a subsurface conductor, presumably a liquid salty ocean. Finally, we discuss the logistics of building, landing, and operating small sounding observatories on the surface of Europa.

THE PRINCIPLE BEHIND ELECTROMAGNETIC INDUCTION SOUNDING

When a conductor is placed in a time-varying primary magnetic field, eddy currents flow on the surface of the conductor in response to the accompanying varying electric field (see Fig. 1, top). These eddy currents generate a secondary or induced field, which tends to reduce the primary field inside the conductor. If the primary field is uniform (i.e., order 1 in external spherical harmonics) and the conductivity distribution has spherical symmetry, the induced field outside the conductor would also be of order 1 in the internal spherical harmonics (i.e., a dipole field). The secondary field may have a phase lag with respect to the primary signal, but its frequency would be identical to that of the primary signal. The sum of primary and secondary fields creates a total field, which avoids the conductor (Fig. 1, bottom). If the period of the primary signal is

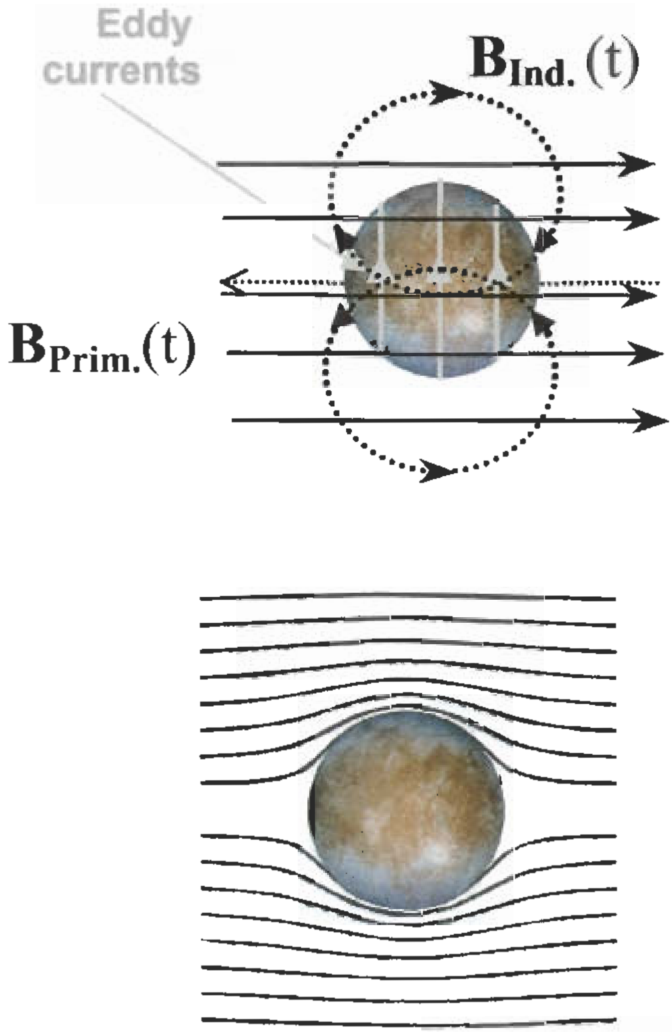

FIG. 1. **Top:** Time-varying primary field (black solid lines) and the induced field (black dotted lines) generated by the eddy currents (gray arrows) that flow on the surface of a conductor like Europa's ocean to prevent the primary field from penetrating the conductor. **Bottom:** The sum of primary and induced fields shows that the lines of force of the varying magnetic field avoid the conducting obstacle.

many seconds or longer, the fields can be measured with a DC magnetometer.

EVIDENCE OF ELECTROMAGNETIC INDUCTION FROM EUROPA

Europa is located in the inner magnetosphere of Jupiter where it interacts with Jupiter's magnetic field and plasma (Kivelson *et al.*, 1999). Because the magnetic and rotation axes of Jupiter are not aligned, Europa in its own frame experiences a changing field at the synodic rotation period (11.1 h) of Jupiter. Figure 2 shows the changing magnetic field experienced by Europa during a rotation period of Jupiter. The field was calculated from an empirical model of Jupiter's magnetic field (Khurana, 1997). The coordinate system used has its x-axis parallel to the direction of motion of Europa, and its y-axis points towards Jupiter. Europa also experiences a strong but constant magnetic field (~500 nT) directed along its spin axis. Because this component does not change in magnitude, no induction field is generated in response to this component. Thus, the changing magnetic field experienced by Europa as it orbits Jupiter lies in the equatorial plane of Jupiter and is elliptically polarized.

The spacecraft Galileo has made nine close flybys of Europa. The first two flybys in which the conditions were optimum for the detection of an induction field took place in the fourth and the

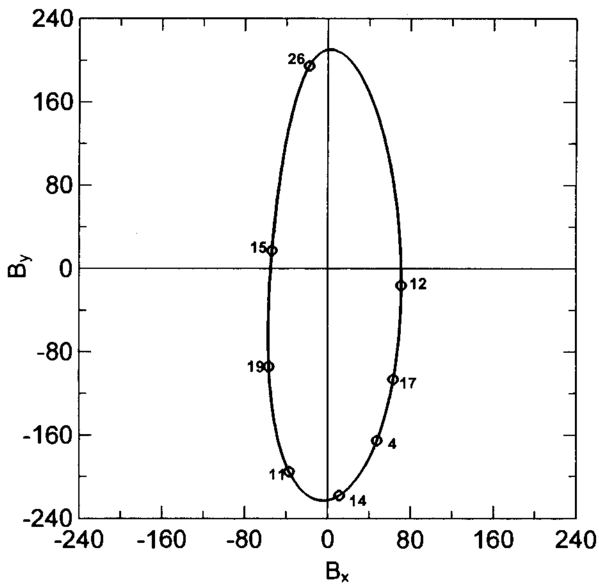

FIG. 2. The varying magnetic field sampled by Europa during an 11.1-h synodic rotation period of Jupiter. Open circles correspond to the field values observed by Galileo during flybys of the moon. The numbers next to the circle correspond to the orbit numbers of Galileo during which the flybys occurred.

14th orbits of Jupiter. Figure 3 shows the detected and modeled magnetic field from these flybys plotted along the spacecraft trajectories in a vectorial representation. The model assumes that Europa is a perfect conductor and that the magnetic field of Jupiter is uniform at the scale size of Europa. For such a situation, it can be shown (Parkinson, 1983) that the surface value of the induction dipole at its pole would be equal to but opposite in direction to that of the Jovian equatorial field. Thus, the figure makes a strong case that a global conductor is located very close to the surface of Europa. If the conductor were buried under a thick layer of ice with thickness d, then the induced field would have been reduced by a factor of $(1 - d/R_M)^{-3}$.

An idea of the effectiveness of a material in generating an induction response is given by the skin depth $S = (2/\mu\sigma\omega)^{1/2}$ of a signal; here, σ is the conductivity of the material, and ω is the circular frequency of the signal. For high conductivity, the skin depth is small (i.e., the propagating signal attenuates by an e-folding in a short distance S). On the other hand, in poor conductors, the skin depths of the signals are large, and they can easily penetrate the material. A consideration of the conductivities of common geophysical materials (Table 1) shows that the skin depths of pure liquid water, ionosphere, rocks, and common mineral assemblages for a wave with a period on the order of 10 h are very large and could not have generated the observed induction dipole signatures detected near Europa. Global layers of pure metals with thicknesses in the range of tens of meters are capable of generating the observed signal but are unlikely to occur in a light differentiated material like ice. The only plausible material that could account for the observed induction signature is a globe-wide liquid salt-water ocean located at a depth of a few to a few tens of kilometers from the surface (Khurana *et al.*, 1998; Kivelson *et al.*, 1999; Zimmer *et al.*, 2000).

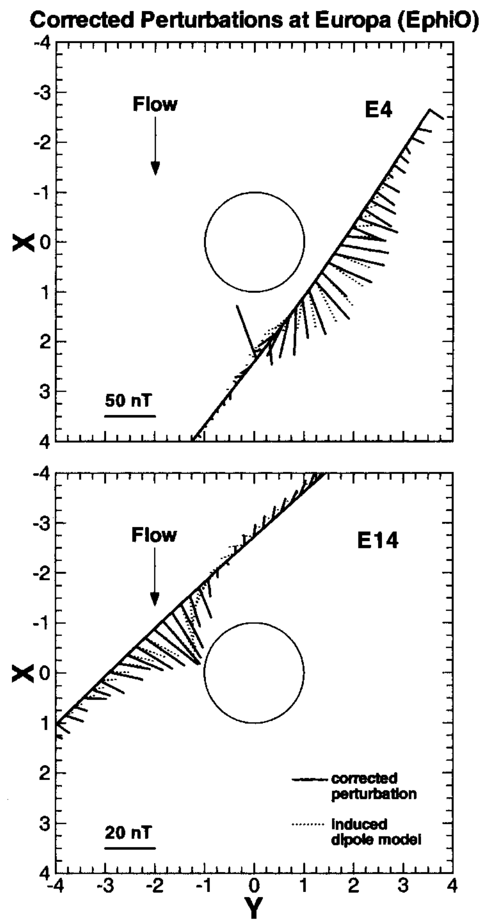

FIG. 3. Perturbation field observed by Galileo during the E4 and E14 flybys of Europa (solid lines) plotted as vectors along the spacecraft trajectory. The modeled values of the induction signal are plotted as dotted vectors. The observed perturbation vectors have been corrected for the plasma effects caused by the interaction of Jovian plasma with Europa. The coordinate system is explained in the text.

TABLE 1. CONDUCTIVITIES OF COMMON MATERIALS AND THEIR SKIN DEPTH FOR A 10-H WAVE

Material	*Conductivity (at 0°C)* S/m	*Conductivity (at 0°C)* Reference	*Skin depth for a 10-h wave (km)*
Water (pure)	10^{-8}	Holzapfel (1969)	10^{6}
Ocean water	2.75	Montgomery (1963)	60
Ice	10^{-8}	Hobbs (1974)	10^{6}
Ionosphere (E layer)	2×10^{-4}	Johnson (1961)	7×10^{3}
Granite	10^{-12}–10^{-10}	Olhoeft (1989)	10^{8}–10^{7}
Basalt	10^{-12}–10^{-9}	Olhoeft (1989)	10^{8}–3×10^{6}
Magnetite	10^{4}	Olhoeft (1989)	1
Hematite	10^{-2}	Olhoeft (1989)	10^{3}
Graphite	7×10^{4}	Olhoeft (1989)	0.4
Cu	5.9×10^{7}	Olhoeft (1989)	0.01
Fe	1×10^{7}	Olhoeft (1989)	0.03

MULTIFREQUENCY ELECTROMAGNETIC SOUNDING

The induction observations from Galileo have a major limitation. The spacecraft spends a very short time near Europa during a typical flyby. Therefore, the short interval of observations is interpreted with the assumption that the inducing signal consists of a single frequency at the synodic rotation period of Jupiter (11.1 h). Observations at a single frequency cannot be inverted to determine independently both the conductivity and the thickness of the conducting layer. In reality, Europa is bathed in a variety of low-frequency electromagnetic waves that could be used to obtain information from different depths in the conducting layer. Some of these waves arise from the dynamics of Io's plasma torus (Strobel, 1989). An important long-period frequency arises from the eccentricity of Europa's orbit. As Europa orbits Jupiter, it perceives a difference of ~15 nT in Jupiter's field strength between its perijove and apojove, thus sampling a wave at the 85.2-h orbital period of Europa (3.3 μHz). Figure 4 shows the amplitude spectrum of the primary field observed by Europa calculated from an empirical model of Jupiter's magnetic field (Khurana, 1997).

In order to understand induction at multiple frequencies, we modeled the response of Europa at any frequency ω by using a three-shell model. We assumed that the outermost shell of Europa consists of solid ice and has an outer radius r_m equal to that of Europa. This shell is assumed to have zero conductivity. The next shell containing the Europan ocean has an outer radius r_0 and conductivity σ. Finally, the innermost shell consisting of silicates is again assumed to have negligible conductivity and radius r_1. As previously discussed, since the primary field is uniform and the conductivity distribution has spherical symmetry, the induced field outside the conductor ($r > r_0$) would be dipolar (Parkinson, 1983):

$$\mathbf{B}_{\text{sec}} = \frac{\mu_0}{4\pi}\left[3(\mathbf{r}\cdot\mathbf{M})\mathbf{r} - r^2\mathbf{M}\right]/r^5 \tag{1}$$

The induction moment would be opposite in direction to the primary field. The moment **M** oscillating at the same frequency ω can therefore be written

$$\mathbf{M} = -\frac{4\pi}{\mu_0}Ae^{i\phi}\mathbf{B}_{\text{prim}}{r_m}^3/2 \tag{2}$$

and

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

The parameters A (relative amplitude) and ϕ (phase lag) are real numbers, which after Parkinson (1983) can be written as complex equations:

$$Ae^{i\phi} = \left(\frac{r_0}{r_m}\right)^3 \frac{RJ_{5/2}(r_0k) - J_{-5/2}(r_0k)}{RJ_{1/2}(r_0k) - J_{-1/2}(r_0k)} \tag{4}$$

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

where $k = (1 - i)\sqrt{\mu_0\sigma\omega/2}$ has the dimension of a (complex) wave vector and J_m is the Bessel function of first kind and order m.

Figure 5 shows the polar surface induction field created by the interaction of Europa with Jupiter's varying field at the two main frequencies experienced by Europa. We assumed that the

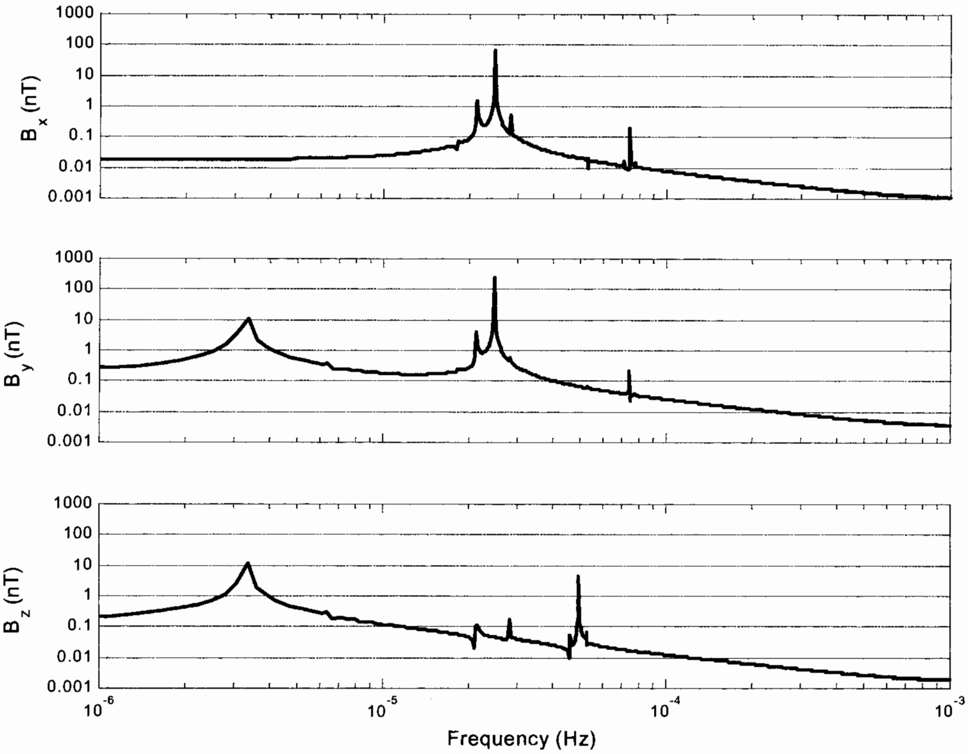

FIG. 4. The amplitude spectrum of the primary field that would be observed by a magnetometer located on the surface of Europa.

wave amplitude at the synodic rotation period of Europa is 250 nT and the wave amplitude at the orbital period is 14 nT. It can be seen that for ocean thicknesses of less than a few tens of kilometers and for ocean conductivities <0.2 S/m, the amplitude curves of the two frequencies are essentially parallel to each other. However, when the ocean thickness exceeds 100 km and the conductivity is >0.2 S/m, the curves begin to intersect each other. In a portion of this parameter regimen ($\sigma > 0.2$ S/m and $r_0 - r_1 > 100$ km), if the responses of Europa to 11.1-h and 85.2-h period waves are known, one could uniquely determine the ocean thickness and its conductivity. Yet even in this parameter range it may only be possible to obtain lower limits on the conductivity and the ocean thickness (i.e., when the ocean conductivity or the thickness exceed a certain value).

In Fig. 6 we plot the amplitude response of an Europan ocean with a conductivity similar to that of the Earth's ocean (2.75 S/m) as a function of the thickness of the ocean. It is clear that the amplitude response of the 11.1-h wave saturates at a thickness of 10 km. Thus, this wave by itself cannot distinguish between an ocean of 10 km thickness and an ocean of 60 km thickness. On the other hand, the 85.2-h wave has no problem in distinguishing between these two types of oceans.

THE PRINCIPLE BEHIND SEISMIC SOUNDING

There are four basic seismic wave modes. P waves (also called compressional or longitudinal waves) cause a ground motion in the same direction as the transmission direction of the wave and are the first to arrive at an observatory because of their fast speeds. The next to arrive are the S waves (also known as shear or transverse) waves, whose associated ground motion is transverse to the direction of the propagation of the waves. Shear waves cannot propagate through liquids because they require rigidity as a restoring force for propagation. The third wave modes to arrive are the surface waves known as Love waves, which produce transverse motion in the horizontal direction. The last wave modes to arrive are the Raleigh surface waves, which have a rolling type of motion associated with them.

Seismic sounding exploits the fact that seismic

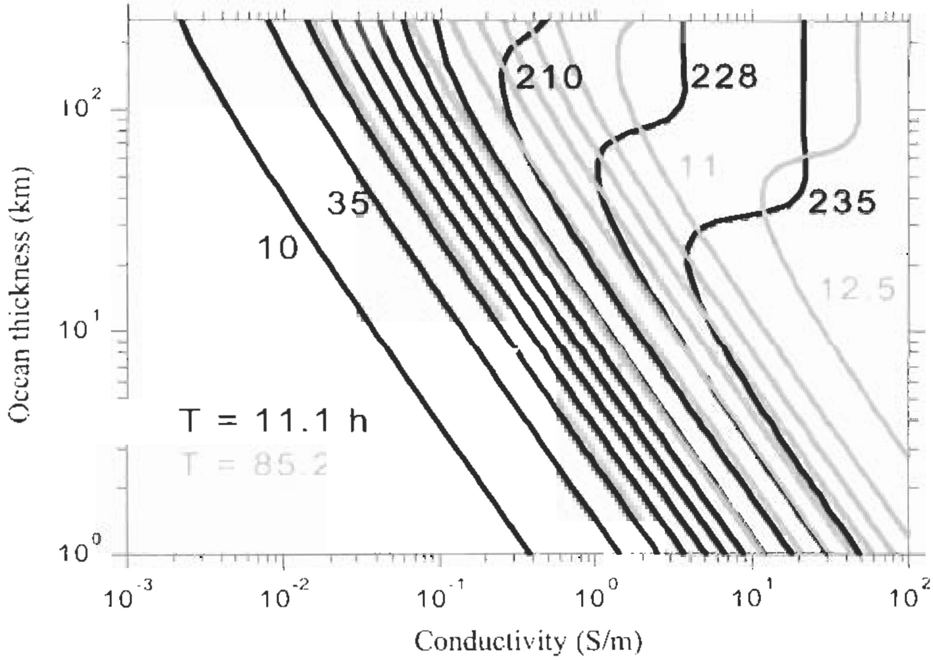

FIG. 5. **Contours of induced field (in nT) generated at the surface pole in response to the 11.1–h wave (black lines) at the synodic rotation period of Jupiter and the 85.2–h wave (gray lines) at the orbital period of Europa.** The x-axis covers a range of ocean water conductivities, and the y-axis covers a range of ocean thickness values.

waves travel at different velocities in media that have different Young modulii (compressibility), shear modulii (rigidity), or densities. At the interfaces of different materials, body waves (P and S waves) are reflected, refracted, diffracted, or converted into each other. By using an array of seismometers and identifying different wave types in the recorded wave train, one can locate the source and the reflecting interfaces. Using this type of analysis on the Earth, seismic observations have located and characterized the liquid outer core and the solid inner core. In addition, normal modes of various shells (the icy crust, liquid ocean, rocky mantle, and the metallic core) also reveal the gross radial structure of a planetary body. Figure 7 shows various wave modes that provide information on the location and properties of the inner and outer cores of the Earth. Of great significance for Europa is the shadow zone observed in terrestrial earthquakes at angles of >105° from the epicenter. Beyond this distance, no directly transmitted S waves are observed because of the fluid nature of the outer core. If a region devoid of S waves could be identified in the H_2O layer of Europa, this would constitute direct proof of the presence of a liquid ocean.

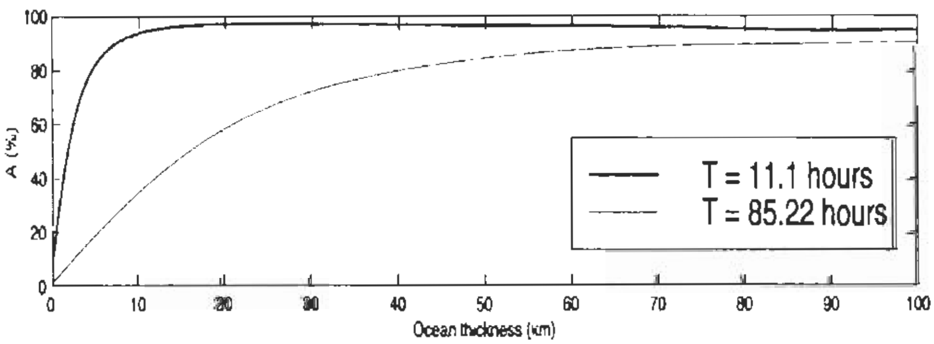

FIG. 6. **The normalized amplitude responses of the 11.1–h wave (black curve) and the 85.2–h wave for an ocean with a conductivity of 2.75 S/m.**

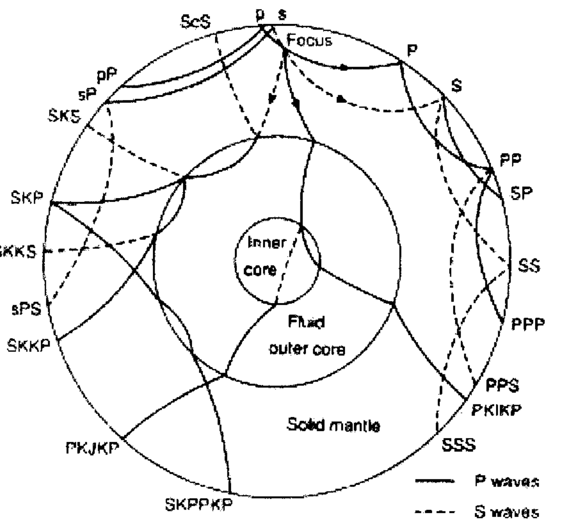

FIG. 7. Possible wave paths for P waves (solid segments) and S waves (dashed segments) as they travel through the interior of the Earth. Reproduced with permission from Doyle (1995).

Satellite-quakes, interesting in their own right, are likely to provide signals for the seismic measurements on Europa. As discussed by Greenberg *et al.* (1998) and by Hoppa *et al.* (1999), diurnal variations of the Europan tides and the nonsynchronous rotation of Europa's interior create significant stresses, which can be released periodically in localized regions. Buildup and release of such stresses may cause fracturing of ice and release seismic energy. Asteroid and comet impacts would also contribute to the background signals, although their impact frequency at subkilometer sizes is poorly constrained (Zahnle *et al.*, 1998).

The wave velocities of P and S waves at a temperature of 270K in solid ice are ~3.58 and 1.66 km/s, respectively (Hobbs, 1974). Hobbs (1974) and Kovach and Chyba (2001) discuss the effect of temperature on the seismic wave velocities in ice. Kovach and Chyba (2001) estimate that at 110K, the average surface temperature of Europa, the P and S wave velocities would be 4.0 and 2.0 km/s, respectively. On the other hand, the P wave velocity in liquid water at 270K is ~1.5 km/s. The P wave velocity through rocks is ~6 km/s. The seismic technique is therefore highly suited for locating liquid water because of the large velocity mismatch at the liquid/solid interface, which would be a strong reflector of seismic waves. Recently, Kovach and Chyba (2001) have discussed the use of Crary waves (an unusual type of vertically polarized shear wave, which are reflected at the critical angle), Love waves, and Raleigh waves to determine the thickness of the ice cover on the liquid ocean. In particular, the Love waves are dispersive (i.e., different frequencies travel at different speeds) in thin shells, a property that can be exploited to determine the thickness of the ice layer (Kovach and Chyba, 2001).

THE OBSERVATORY

To probe the Galilean satellites interiors, we introduce the concept of SOUNDERs, Surface Observatories for UNDErground Remote-sensing. Extremely lightweight (<5 kg total mass) and autonomous, these observatories could be deployed as penetrators from a passing spacecraft (Khurana *et al.*, 2001). A typical SOUNDER would comprise a fluxgate vector magnetometer (see Parkinson, 1983, to understand the principle behind this instrument), a three-axis seismometer (see Doyle, 1995, for principle of such an instrument), a data processing unit, and a transmission unit. Because modern miniaturization techniques allow the observatories to be extremely lightweight, several of them could be carried on a single spacecraft and deployed to multiple sites on multiple satellites. The SOUNDERs could also be used to supplement a payload on a future Europa lander mission to study prebiotic chemistry.

By design, the penetrator-borne observatories would be buried under tens of centimeters of ice, so that the high radiation in the environment can be overcome. Paranicas *et al.* (submitted for publication) and Cooper *et al.* (2001) have discussed the electron and ion irradiation of Europa's surface by the Jovian plasma. Figure 8 of Paranicas *et al.* (submitted for publication) shows that a 50 (100) cm shield composed of water ice reduces the combined dose rates of electrons and ions by a factor of ~10^5 (10^6) from the ambient value. Thus, an unshielded subsurface observatory buried at a depth of 50 (100) cm in water ice would be expected to receive a radiation dosage of ~10 (1) rad/day, an easily manageable rate. By comparison, the Galileo spacecraft was designed to survive 150 Krad of radiation, and the Europa Orbiter would be designed to withstand >2 Mrad of radiation.

It would be beneficial to operate several SOUNDERs simultaneously on the surface of Europa. Electromagnetic sounding requires measurements from a minimum of two observatories to separate uniquely the uniform external field harmonics (the inducing signal) from the dipolar internal field harmonics (the induction response). Similarly, a seismic source can be uniquely located from the measurements of a single wave mode if observations from a minimum of three locations are available. However, by using multiple wave modes or by exploiting the dispersion properties of certain specialized wave modes, as discussed by Kovach and Chyba (2001), even a single surface observatory can characterize the thickness of the ice layer.

As discussed above, electromagnetic sounding at multiple frequencies from fixed locations has the potential of providing unique estimates of the conductivity and thickness of the ocean. If periods >2 weeks can be measured in the induction signal, invaluable information on the locations and conductivities of the oceans and deeper layers can be obtained. The seismicity of Europa is expected to have a Europan-diurnal cycle [85.2 h (Hoppa *et al.*, 1999)]. It would be prudent to collect seismic data over several diurnal cycles to ensure that a sufficient number of useful events are recorded. Because both sets of sensors (magnetic and seismic) provide measurements of long vector time series, the data acquisition, processing, storage, and transmission functions can be combined in a single unit to provide major mass and power savings.

Another benefit of integrating the two instruments is risk reduction. If, for example, the ambient seismic sources are found to be less than optimum for subsurface sounding, then the induction sounding would still provide a fallback option because the electromagnetic sounding signal is always present. On the other hand, if the induction sounding is unable to characterize both the thickness and the conductivity of the ocean layer, then the seismic data could be used to deduce the thickness of the ocean, and induction data could be used to estimate its conductivity and, thus, help place limits on the composition of the ocean water. For all of these reasons, we recommend that a minimum of two surface observatories be operated on Europa over a time period of at least several weeks to explore the structure and the properties of the subsurface ocean.

The power requirements of such small remotely operated geophysical observatories have not been fully analyzed. However, preliminary studies show that the power consumption can be kept to a minimum (2–3 W). Communication with passing or orbiting spacecraft would be made in burst mode; any excess power not needed to operate the instruments and the observatory would first be stored in a battery so that when an orbiting spacecraft was overhead the transmitter could use the power from the battery to transfer data to the spacecraft.

The only known source of power that can operate the observatory for the long duration required for a minimum mission (1 month) is a radioactive thermal generator. Additional radioactive heating units may be required to keep the observatory at an optimal temperature in the cold environment of Europa's surface (100–150K).

DISCUSSION

Magnetic sounding technique has a long history of use in deciphering the Earth's interior (Parkinson, 1983). Magnetic sounding has also been used to sound the interior of the Moon (Hood *et al.*, 1982). The technique is especially well suited for studying the interiors of the Galilean satellites because Jupiter provides a strong natural primary signal. From continuous observations from just two magnetic observatories over a few weeks, both the internal (the response of the Jupiter moon) and the external (Jupiter's primary signal) harmonics can be uniquely determined over a range of frequencies. As discussed above, electromagnetic sounding at multiple frequencies from fixed locations has the potential of providing unique estimates of both the conductivity and the thickness of the ocean. If periods >2 weeks can be measured in the induction signal, invaluable information on the locations and conductivities of the deeper concentric structural layers of Europa can also be obtained. One advantage of having magnetic observatories located on the surface rather than on an orbiting spacecraft is that the surface measurements can be inverted using simple linear least-squares techniques to obtain information at all of the frequencies contained in a spectrum. For an instrument on an orbiter it is possible to invert only a few prime frequencies (for example, those corresponding to 11.1 and 85.2 h) because the time-dependent factor r in equation 3 renders the equations nonlinear and requires the use of special inversion techniques.

There have been two previous efforts to obtain seismic information from extraterrestrial bodies. Between 1969 and 1977 a four-station seismic network left on the Moon by the Apollo astronauts operated successfully and helped unravel its interior structure. The Viking 1 and 2 Mars Landers also carried seismometers that remained onboard the spacecraft after the spacecraft landed on Mars. The Viking 1 seismometer could not be uncaged, and therefore the experiment was lost. However, the Viking 2 seismometer deployed successfully and sent useful data over a period of 19 months. The results and recommendations from the operation of the Apollo and Viking seismometers are summarized in a panel report by Solomon *et al.* (1991).

Seismic data from the Moon showed that even very weak signals can be detected on its surface because of its small size and low background noise. The seismic signals decayed more slowly than at the Earth (tens of minutes to 2 h) because of the very high Q value (a quality factor that measures the lightness of wave damping) of the crust and upper mantle (4,000–15,000 for the Moon compared with $\sim$3,000 for the Earth). The Apollo seismic data allowed construction of a model for the internal structure of the Moon that postulates an inner core with a radius of $\sim$0.2 R_M, a mantle divided into three distinct layers with the middle mantle responsible for the deep earthquakes, and a thin crust where the shallow earthquakes occur. This model was fully corroborated by magnetic induction studies carried out on the data from the Apollo surface magnetometers (Hood *et al.*, 1982). Similar discoveries could be forthcoming if small autonomous observatories could be successfully landed and operated on the surface of Europa.

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ABBREVIATION

SOUNDER, Surface Observatory for UNDEr-ground Remote-sensing.

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