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Note

The temperature of Europa's subsurface water ocean

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Abstract

A 100 km deep liquid water ocean probably underlies the icy exterior of Jupiter's satellite Europa. The long-term persistence of a liquid ocean beneath an ice shell presents a thermal conundrum: Is the temperature of the ocean equal to the freezing point of water at the bottom of the ice shell, or is it equal to the somewhat warmer temperature at which water attains its maximum density? We argue that most of the ocean is at the temperature of maximum density and that the bulk of the vigorously convecting ocean is separated from the bottom of the ice shell by a thin "stratosphere" of stably stratified water which is at the freezing point, and therefore buoyant. If Europa's subsurface water ocean is warm, it could explain the widespread geologic evidence for apparent melt-through events observed on its surface and may constrain the overall age of its surface.

Keywords: Europa; Interiors; Thermal history

Data from NASA's Galileo mission indicate that a liquid water ocean more than 100 km deep probably underlies the water-ice surface of Jupiter's large moon Europa (Anderson et al., 1998; Khurana et al., 1998; Showman and Malhotra, 1999). Because of the peculiar thermophysical properties of water, however, the temperature of the ocean beneath the ice shell may not simply equal the melting temperature of ice. Instead, most of the ocean may be close to the temperature at which water attains its maximum density.

Estimates of Europa's surface heat flow are around 50 $\mathrm{mW/m^2}$, but most of this energy is probably created by tidal flexing of the ice shell itself (Ojakangas and Stevenson, 1989). Nevertheless, Europa's overall density (Anderson et al., 1998) of 3.02 $\mathrm{gm/cm^3}$ indicates that it is mainly composed of silicates, not ice, and thus contains radioactive heat-producing elements. Assuming that Europa's rocky interior produces heat at the same rate as a carbonaceous chondrite, the heat flow Q_b at the base of the water ocean is about 8 $\mathrm{mW/m^2}$ (see Fig. 1). This may be two times larger if some of Europa's tidal heat is dissipated in its deep interior (Squyres et al., 1983). This is enough to keep water beneath the ice shell stirred by vigorous thermal convection.

This minimum heat flux implies a Nusselt number (a dimensionless ratio between the total heat transfer and the conducted heat flux) $Nu = Q_b/(k\Delta T/D)$, where k is the thermal conductivity of water (0.55 W/m K) and D is the depth of the convecting ocean (about 100 km). The temperature difference ΔT between the top and bottom of the ocean is derived from the relation between Nu and the Rayleigh number Ra (a dimensionless measure of convective vigor) $Nu = (Ra/Ra_{cr})^{\beta}$, where Ra_{cr} is the critical Rayleigh number, typically about 1000, and the exponent $\beta \approx 0.309$ at high Ra (Niemela et al., 2000). The Rayleigh number is defined as $Ra = \alpha g \Delta T D^3/\kappa\nu$, where g is Europa's acceleration of gravity (1.33 $\mathrm{m/sec^2}$), α is the thermal expansion coefficient (typically about 2×10^{-5}, but see the discussion below), ν is kinematic viscosity (1.5×10^{-6} $\mathrm{m^2/sec}$), and κ is thermal diffusivity (1.33×10^{-7} $\mathrm{m^2/sec}$). Eliminating ΔT between these equations yields large values for $Ra \sim 10^{20}$ and $Nu \sim 10^6$.

The theory of Rayleigh–Bénard thermal convection (Turcotte and Schubert, 1982) predicts that the thicknesses d of both hot and cold conductive boundary layers at the bottom and top of the convecting ocean, respectively, are $d = D/Nu \approx 10$ cm. The total super-adiabatic temperature difference ΔT between the top and bottom of the ocean is a few milliKelvins. Moreover, because the convective overturn times are long compared to Europa's rotational period of 3.5 days, the convective regime should be rotationally dominated (geostrophic) (Boubnov and Golitsyn, 1990; Fernando et al., 1991; Jones and Marshall, 1993; Vorobieff and Ecke, 2002). Experiments and numerical simulations (Boubnov and Golitsyn, 1990; Fernando et al., 1991; Jones and Marshall, 1993) with Rayleigh numbers up to 10^{11} suggest that in this regime the vertical velocities w are of order $w \sim (\alpha g Q_b/\rho c_p \Omega)^{1/2} \sim 0.1$ mm/sec, where Ω is Europa's rotational angular frequency, ρ is the density, and c_p is the specific heat of oceanic water. Nevertheless, debate exists over how the scaling should depend on Rayleigh number (Klinger and Marshall, 1995; Solomatov, 2000), and the high Rayleigh numbers applicable for Europa's ocean may allow more vigorous velocities up to 1 mm/sec. The subsurface water ocean of Europa should be well-stirred and rather homogeneous in temperature, with root-mean-square thermal fluctuations (Niemela et al., 2000) of $5 \times 10^{-4}\Delta T$.

If the fluid in Europa's ocean were almost anything else but water, this would be the end of the story. The top and bottom would differ only slightly in temperature and the internal heat would be transported rather efficiently through the deep, low-viscosity ocean. However, because of water's peculiar properties, the story is more complex. First, the less dense solid-ice phase floats on the liquid phase, so there can be a cold, solid shell overlying the warmer liquid below. This is familiar from frozen lakes and ponds on Earth. Furthermore, the density of water is maximum at 3.98 °C, not at its freezing point at 0 °C (at 1 bar pressure). Although this maximum is not strong (the difference in density is only 1.32×10^{-4} $\mathrm{gm/cm^3}$), it does mean that the thermal expansion coefficient α for liquid water is negative

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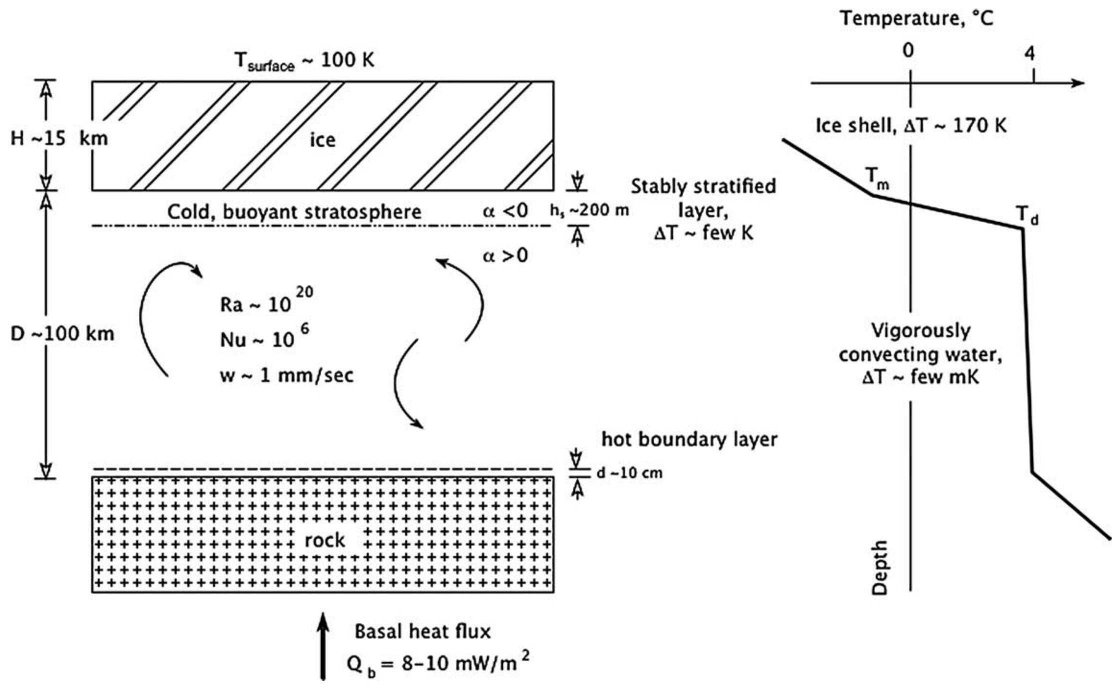

Fig. 1. The thermal structure of a deep, vigorously convecting liquid water ocean on Europa that is capped by a "stratosphere" of cool, buoyant water. The left panel illustrates the various oceanic regions discussed in the text and the right panel schematically shows the temperature as a function of depth. Note that the temperature in the ice shell is purposely left vague—it may be influenced by convection and tidal heat deposition, but that is not important for the thermal structure of the water ocean.

between 0 and 3.98 °C and so Rayleigh–Bénard convection cannot occur. But if convection shuts off in the water below the ice shell, the ocean must be warmed from below until α becomes positive above 3.98 °C (a possibility first suggested for the galilean satellites by Showman et al. (1997)). Convection then begins again, and if the convection were vigorous enough, it would bring this relatively warm water into contact with bottom of the ice, leading to melting of ice. This is the thermal conundrum of a stable ice shell. It does not arise in terrestrial lakes and ponds because seasonal freezing and thawing never permits a long-term thermal equilibrium. The closest terrestrial analogue may be Lake Vostok beneath the Antarctic ice sheet, although, even in this case, exchange of material with the overriding ice flow prevents the lake from being a closed system (Siegert et al., 2001).

Careful examination of the properties of terrestrial seawater (Feistel and Hagen, 1995) shows that there may be an easy way around this conundrum. Figure 2 shows the freezing temperature (solid lines) and temperature of maximum density (dashed lines) as a function of pressure and salinity. As the pressure increases, even for fresh water, the maximum density occurs at the freezing point for pressures greater than about 27 MPa (depth 22.5 km on Europa). Similarly, if the oceans are more saline than about 30 per mil, then the maximum density occurs at the melting point for all pressures. Thus, if the ice shell is more than about 23 km thick, or the ocean as salty as the Earth's oceans (34.4 per mil on average), then convection proceeds as described above and no thermal contradiction occurs. The precise thickness of the Europan ice shell is presently a contentious issue, with estimates ranging from 1 to 30 km or more (Turtle and Pierazzo, 2001). Although spectral reflectance suggests that the subsurface ocean contains salts of some kind (McCord et al., 1998), the only real data on the salinity come from the detection of an induced magnetic moment. Analysis of these data (Zimmer et al., 2000) put only a weak lower limit on the conductivity of about 116 mS/m for a 100 km deep ocean, which corresponds to a salinity of 1 per mil (Weast, 1972), assuming that the salt composition is similar to terrestrial seawater. The conductivity and thermal reflectance spectra could also be due to other components, such as sulphuric acid (Carlson et al., 1999) or the sulfate salts of sodium and magnesium (Zolotov and Shock, 2001), but we presently lack the thermochemical data to evaluate this quantitatively. Hogenboom et al. (1995) measured thermochemical properties of very concentrated solutions of magnesium sulfate, for which no maximum in density occurs. Data by Chen et al. (1980), while limited to temperatures above 0 °C, indicate that, at 1 atm pressure, the salinities needed to decrease the temperature of maximum density from 3.98 to 0 °C are 18, 26, 19, and 34 per mil for the salts NaCl, $MgCl_2$, Na_2SO_4, and $MgSO_4$, respectively, suggesting that the results quoted above for the better-studied seawater solutions are reasonably typical. More thermochemical work on such solutions as a function of pressure is urgently needed. Present data on

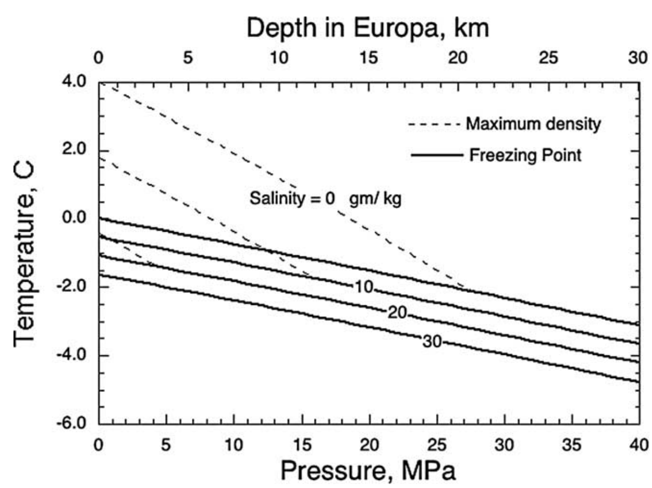

Fig. 2. Melting temperature T_m and maximum density temperature T_d of water as a function of pressure and salinity (Feistel and Hagen, 1995). The dashed lines indicate the pressure and temperature of the maximum density, and the solid curves show the melting temperature. The multiple curves correspond to four different values of the salinity, which are cited in units of gm/kg.

Europa's ocean thus does not guarantee that the thermal conundrum posed by the peculiarities of water can be simply evaded.

A second possibility is that, even though the thermal expansion coefficient may be negative some distance below the ice-water interface, convection in the ocean below may be so vigorous that rising currents carry through the convectively stable zone and bathe the bottom of the ice shell with water from the deep ocean. This process is known as penetrative convection (Zhang and Schubert, 2000; Straughan, 1993) and, if it occurs, ensures that the temperature of the convecting ocean must be within a few milliKelvins of the freezing temperature of water at the top of the ocean.

The most interesting case occurs when the cool buoyant water beneath the ice forms a stable layer overlying a convecting ocean that may be several K warmer than the base of the ice shell (Showman et al., 1997). See Fig. 1. The temperature at the base of the ice shell is fixed by the melting point of ice at the appropriate pressure (thickness of ice) and ocean salinity. The temperature then rises linearly through the buoyant "stratosphere" until it reaches the temperature at which water achieves its maximum density, where α becomes positive. The temperature difference ΔT_s across this layer is the difference between the melting temperature T_m and the temperature at maximum density T_d, both of which are functions of pressure and salinity. Heat is transferred by conduction in this layer, which may be only a few hundred meters thick. The layer thickness h_s is determined by the requirement that conduction carries the heat flux Q_b transferred from depth, $h_s = k\Delta T_s/Q_b$. Note that this layer is considerably thicker than the Rayleigh–Bénard boundary layer of thickness d. At greater depths, the temperature remains close to the adiabat. The adiabatic gradient is quite small, only about 0.011 K/MPa, which is much less than the average slope of the melting curve, −0.085 K/MPa, so the adiabat appears nearly vertical in Fig. 1 (in fact, the adiabatic gradient is slightly negative in the region where α is negative, but the slope is too small to represent in the plot). The ocean is then nearly isothermal, because convection itself causes only small departures from the adiabat. Although the temperature of most of the ocean is well above the temperature of maximum density of water and thus convects normally with a positive thermal expansion coefficient, the temperature at its top is fixed at T_d.

The principal question is thus whether such a stratosphere is stable in the face of vigorous convection below. Just as thunderstorms top out against the Earth's stratosphere, the cool buoyant water layer on Europa may stop rising convection currents. A crude way to analyze the stability is to compute the distance δ that a plume rising at velocity w may penetrate into water with a density difference $\Delta\rho$. If we assume that the density increases linearly across the layer, then from the maximum density to the density at the melting point, at the maximum height of penetration, $\Delta\rho = \delta(\rho_{\text{max}} - \rho_{\text{melt}})/h_s$. Equating the kinetic and gravitational energy of such a plume we find $\delta = w^2/(2g\Delta\rho/\rho)$, where ρ is the mean density of the water. Inserting the above equation for $\Delta\rho$ and solving, we find $\delta = w\{\rho h_s/(2g(\rho_{\text{max}} - \rho_{\text{melt}}))\}^{1/2}$. Using this equation, δ ranges between 0.05 and 0.5 meters for $w = 0.1$–1 mm/sec, implying that convective plumes entering the stable layer can ascend only a fraction of a meter before descending back into the oceanic interior. At first glance, this suggests that the stratosphere can be maintained against the penetrative convection. However, the overshooting plumes mix hot, dense water into the stable layer, increasing its temperature and reducing its buoyancy. Whether the stable layer can be maintained depends on a competition between the rate at which overshooting plumes erode the layer and the rate at which conduction cools the layer from above, restoring its buoyancy. The timescale for "eating through" the entire stable layer is given by the mean potential energy of the stable layer (relative to an adiabat) divided by the rate at which impinging plumes deposit potential energy into the layer through the mixing they induce. To order of magnitude it is given by $(\rho_{\text{max}} - \rho_{\text{melt}})gh_s^2/(e\rho w^3)$, where e is the fraction of the mechanical energy flux that does work mixing cold water into the stable layer. We estimate that e is about 0.1. The resulting timescale is uncertain, but is probably 3000–300,000 yr, depending on the vigor of the convective plumes and the fraction of the kinetic energy flux that causes mixing. In contrast, the thermal diffusion time for the stable layer, given by h_s^2/κ, is ~ 3000 yr. A comparison of these timescales suggests that, most likely, the stable layer can be maintained against the mixing caused by plumes penetrating from below.

Nevertheless, dynamics in the ocean may affect the stable layer and in local regions allow the hot interior water to impinge directly on the ice. To first order, the stratosphere tends to stabilize the ice shell thickness by removing any topography that exists at the bottom of the ice shell. Where the ice shell is thin, buoyant water in the stable layer pools beneath the thin ice, insulating it and decreasing the heat flux into the ice from below, so that the ice grows thicker. Where the ice is thick, the buoyant layer flows away and allows the warm ocean water to contact the ice and thus reduce its thickness. (The pressure-dependence of the melting temperature and lateral flow of ice away from thick regions, toward thin regions, also tend to maintain a flat ice bottom.)

On the other hand, there may be situations where oceanic fluid dynamics increases the topography at the ice-water interface and perhaps even allows massive melt-through of the ice layer. Several processes, such as warmer vents in the ocean floor (Thomson and Delaney, 2001), or nonlinear vortex-ice topographic interactions, could cause persistent horizontal pressure gradients that induce long-lived flow patterns (such as vortices) in the water. Depending on the pressure field, the cold, buoyant water may be driven away from thin regions of ice, which would bring the underlying hot water in contact with the ice in the region where it is already thinner—hence magnifying the thinning process. Rapid melting would continue so long as the supply of warm water is maintained. This may be the explanation for the widely-observed Chaos regions, which seem to be most simply explained by melt-though events (Greenberg et al., 1999). Many theories for Chaos formation have been proposed (Collins et al., 2000). A natural explanation is that the Chaos regions represent areas where deep, warm ocean waters have come into contact with the overlying ice. The fact that blocks in Conamara Chaos are displaced in a clockwise pattern (Spaun et al., 1998) also supports a vertical, geostrophic flow beneath the region at the time that melting occurred.

Complete melt through is made difficult by the facts that the conductive flux through the ice shell depends inversely on the ice shell thickness (so that extremely large heat flows are required for complete melting) and that lateral flow of ice tends to fill in the "hole" in thousands of years unless the ice shell is thinner than ~ 10 km on average (O'Brien et al., 2002; Stevenson, 2000). Nevertheless, *partial* melt-through events, which should be more common, may also allow disruption of the surface ice and formation of Chaos. For example, if continuous cycling of warm ocean water past the bottom of the shell causes local thinning by an amount Δh (relative to surrounding regions), then the differential stress within the ice associated with this hole is $\sim g\Delta\rho\Delta h \sim 1$ bar $(\Delta h/1$ km), where g is gravity and $\Delta\rho$ is the density difference between ice and liquid water. Thinning of the ice shell by even a few km—potentially only a fraction of the total thickness—would therefore produce internal stresses of a few bars, which may lead to surface disruption as the underlying soft ice flows in response to these stresses. This possibility is supported by evidence that Europa's lithosphere is extremely weak, with a failure stress of less than 1 bar (Hoppa et al., 1999).

More speculatively, Europa's global tectonics may operate in the mode that has been suggested for Venus (Turcotte, 1993): After a paroxysmal overturn that resurfaces the satellite, the ice shell may gradually cool and thicken for a time as the ocean warms beneath. When the ocean becomes sufficiently warm to breach the stable layer once again, the ice shell is consumed and another cycle begins. In this respect Europa's ice-ocean system may exhibit time variability, as is common in many physical systems such as geysers and relaxation oscillators.

If the endpoint scenario described above is valid, it implies that the entire heat flow supplied to the ocean bottom eventually is used (after being temporarily stored as oceanic thermal energy) to melt the ice shell. This scenario, while extreme, allows a simple estimate of the age of the visible surface from the time required to disrupt the entire surface by the melt-through events. We make this estimate by comparing the heat necessary to melt the ice shell, per unit area, to the rate at which heat is supplied from the deep ocean, Q_b. The heat necessary to melt through an ice shell of thickness H is approximately ρLH, where L is the latent heat of melting of ice, 3×10^5 J/kg. (We neglect the smaller amount of heat necessary to raise the temperature to the melting point.) The timescale necessary to melt through this shell is thus $\tau = \rho L H / Q_b$, which, for a basal oceanic heat flow of 10 mW/m^2, corresponds to $\tau = 10^6$ H yr/km. Thus, for a 10 km thick shell the resurfacing time scale is about 10 Myr, which is, perhaps not coincidentally, the same order as the age of the surface of Europa, estimated between 10 Myr (Zahnle et al., 1998) and 50 Myr (Levison et al., 2000). Note that the ocean only contains about $1/3$ of the thermal energy to melt the shell at one time, so an additional increment must come from convective transfer of geothermal energy. These estimates are independent of the exact size of the melt-through events. Nevertheless, as discussed previously, formation of Chaos may not require complete melting of the ice, and furthermore not all of the heat flux will necessarily be used for melting. These caveats tend to decrease and increase the age estimate, respectively.

Substantial storage of thermal energy in the ocean water, and episodic release of this energy, may be possible even if the mean oceanic salinity exceeds 30 per mil. Although such a salty ocean would have positive thermal expansivity and hence no "stratosphere," Europa's ice-sheet thickness may vary in time, and any epoch of global ice-sheet thinning would introduce relatively pure water at the top of the ocean (Kargel et al., 2000). This downward-increasing gradient of salinity inhibits convection and allows build-up of deep ocean temperature; the final state might consist of multiple stacked convecting layers, with underlying layers being denser (more salty) and warmer; minimal mixing between the layers would occur (e.g., Turner (1979); this phenomenon is called double-diffusive convection). Furthermore if the water immediately underlying the ice has salinity less than 30 per mil, then a stratosphere can exist as long as mixing with the underlying salty water is weak. Such a state is not necessarily stable in the long-term, however, and catastrophic release of the thermal energy could occur, leading to surface disruption.

Water is a peculiar substance. The possibility that a subsurface liquid water ocean might exist at all is due to the peculiar fact that solid water is less dense than liquid water. It may be that another peculiarity, the fact that water is most dense at a temperature higher than its melting point, is responsible for not only the tectonic style of Europa's surface, but the overall age of the surface.

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