THE TECTONIC STATE OF EUROPA'S SEAFLOOR LIMITS POTENTIAL WATER–ROCK INTERACTIONS THERE. Henry G. Dawson[1], Paul K. Byrne[1], Christian Klimczak[2], Paul V. Regensburger[3], Steven D. Vance[4], Mohit Melwani Daswani[4], and Douglas J. Hemingway[5], [1]Department of Earth and Planetary Sciences, Washington University in St. Louis, St. Louis, MO 63130, USA (dawson.h@wustl.edu), [2]Department of Geology, University of Georgia, Athens, GA 30602, USA, [3]Department of Earth Sciences, University of Oregon, Eugene, OR 97403, USA, [4]Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109, USA, [5]University of Texas Institute for Geophysics, University of Texas at Austin, Austin, TX, 78758, USA.

Introduction: Numerous moons in the outer Solar System are suspected to have subsurface oceans [1–3], with at least one satellite, Enceladus, showing chemical evidence of water–rock interactions [4]. Although much has been done to study the surfaces of these icy bodies, the tectonic states of the seafloors of ocean worlds remain unknown. To better understand these environments, the rock strength of a given satellite's seafloor and the stresses acting upon that seafloor must be modeled. When those stresses are sufficiently high so as to match the strength of the lithosphere, thus fracturing it and exposing fresh rock to the ocean, serpentinization and other reactions may drive the production of redox couples and dissolved metals into the water column, components thought necessary for the development and sustainment of life [5,6].

Approach: This study is a continuation of previous work [7], which modeled the seafloor strengths of Europa, Enceladus, Titan, and Ganymede on the basis of recent interior models [8]. Overburden pressure was calculated as $\partial P/\partial r = -\rho_r g$, where r is the radius of the body, ρ_r is the density at a given depth, and g is the gravitational acceleration at that depth. Gravity, in turn, is calculated by $\partial g/\partial r = 4\pi G\rho_r - 2(g/r)$, where G is the gravitational constant [9].

Once pressure and gravity are known, the fault strength for normal and thrust faults can be calculated as $\sigma_1/\sigma_3 = (S_v - P_p)/(S_h - P_p) = (\{\mu^2 + 1\}^{0.5} + \mu)^2$ and $\sigma_1/\sigma_3 = (S_H - P_p)/(S_v - P_p) = (\{\mu^2 + 1\}^{0.5} + \mu)^2$, respectively, where S_v is the vertical stress, S_H and S_h are the maximum and minimum horizontal stresses, P_p is the pore fluid pressure, and μ is the coefficient of friction [10]. These equations hold under the assumption that principal stress components are parallel to the vertical and horizontal stresses. Coefficients of friction for both basalt and serpentinite were considered, to account for the possible previous alteration of some seafloor rock.

The ductile strength of the lower lithosphere was calculated as $\dot{\varepsilon} = C\sigma^n\exp(-E/RT)$, where $\dot{\varepsilon}$ is strain rate, C is a constant, σ is deviatoric stress, n is the stress exponent, E is activation energy, R is the gas constant, and T is temperature [11]. Strain rates between 10^{-14} and 10^{-18} s^{-1} were considered, with thermal gradients between 2 K/km and 20 K/km.

Although this modeling approach can be applied to any icy moon for which an interior model exists and an estimate of heat flux can be made, here we focus on Europa. Titan and Ganymede are believed to have a layer of high-pressure ice directly in contact with, or close above, the seafloor [12]; this layer may prevent substantial hydrous reactions and trap any resultant products below the ice, limiting their being widely circulated into the ocean. Enceladus has shown chemical evidence of water–rock reaction products entering the ocean [13], but the high porosity and permeability of the core may result in different water–rock reaction dynamics [14].

Rock Strength: With a choice of parameters to best favor a weak lithosphere (and thus fracturing), the minimum differential stress required to produce normal faults at 1 km depth is 1.5 MPa, and the differential stress to produce thrust faults at 1 km depth is 2.6 MPa. The brittle–ductile transition (BDT) depth was found by modeling where the brittle and ductile strength profiles intersect. For a thermal gradient of 20 K/km and a strain rate of 10^{-18} s^{-1}, we calculated this depth to be ~30 km. For a thermal gradient of 2 K/km for that same strain rate, the BDT lies at a depth of ~160 km (**Figure 1**).

Stresses: Potential stresses acting on the Europan seafloor include those from the diurnal tide and from radial contraction arising from secular interior cooling. Tidal tensile stresses develop at two nodes at equatorial latitudes [15], leading to the possibility of producing extensional structures there—optimal for exposing fresh rock to the ocean. The maximum tidal stress from a tidal bulge is given by $3eEH/R_s$, where e is the eccentricity of the satellite's orbit, E is Young's modulus (here, for basalt), H is the height of the tidal bulge, and R_s is the radius of the satellite. The height of the tidal bulge is $H = h_2R_s(M_p/M_s)(R_s/a_s)^3$, where h_2 is the solid body tidal Love number, M_p and M_s are the masses of Jupiter and Europa, respectively, and a_s is the semi-major axis of the satellite [16].

For Europa, under the assumption of an altered, serpentinite seafloor, an eccentricity of e = 0.249 is required to produce a differential stress of 1.5 MPa (that stress needed to drive normal faulting at equatorial latitudes) [17]—an eccentricity far higher than the present value of 0.0041 [18]. Thus, we conclude that stresses from the diurnal tide do not induce fracturing of the Europan seafloor.

The stress arising from global contraction would lead to horizontal shortening of the surface, possibly forming thrust faults. Although this structure type would not be as efficient in exposing fresh rock to the ocean as joints or normal faults, it could still plausibly allow the conduction of seawater if such faults were critically stressed. The stress from contraction has a value of $2S(\{1+v\}/\{1-v\})(\Delta R/R_s)$, where S is shear modulus, v is Poisson's ratio, and ΔR is the change in radius of the satellite [19]. Considering thermal contraction alone, $\Delta R = \alpha_l R_s \Delta T$, where α_l is the coefficient of linear thermal expansion and ΔT is the change in temperature [20].

For a 980 km-thick silicate mantle and a 475 km-thick metallic core for Europa [8], and thus a value of α_l of 7×10^{-5} [18,21], the required change in radius of the solid interior to match the absolute thrust fault strength of 119 MPa at a depth of 1 km beneath the seafloor is 1 km. Such a decrease in radius of the rocky interior is considerable, and thus would be unlikely to be a means of continually exposing fresh, unaltered rock to Europa's ocean over geological time.

Future Development: Additional sources of stress that have plausibly driven fracturing of the Europan seafloor include seafloor volcanism [22,23], mantle convection-induced tractions at the base of the brittle lithosphere [24], and even the volume increase associated with the reaction of olivine to serpentine [25]. Serpentinization may also play an important role in the chemistry of the ocean, as the process could release H_2 into the water column, acting to help maintain a redox disequilibrium that would be necessary for life [5]. However, depending on the reaction rates of serpentinization under the chemical and physical conditions present at the Europan seafloor, the weathering of rock might be intermittent, resulting in brief pulses of H_2 and metals into the ocean that are unable to sustain meaningful concentrations of life continually over geological time.

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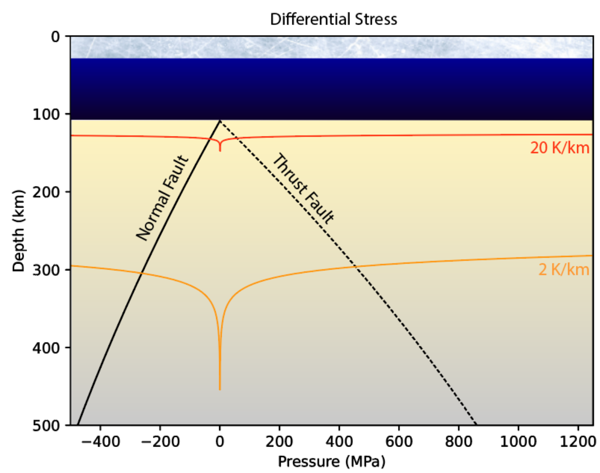

Figure 1. *Lithospheric strength profiles for the Europan seafloor showing the differential stress required to drive normal and thrust faulting. Ductile strength profiles are also shown for heat fluxes of 2 K/km and 20 K/km (orange and red profiles, respectively), for a strain rate of 10^{-18} s^{-1}. The upper icy surface is at the top of the figure, below which is the liquid water ocean. Internal structure are Vance et al. (2017).*