THE INTERNAL STRUCTURE OF ERIS INFERRED FROM ITS SPIN AND ORBIT EVOLUTION. F. Nimmo¹ and M.E. Brown². ¹Dept. Earth and Planetary Sciences, U.C. Santa Cruz, fnimmo@ucsc.edu, ²Division of Geological and Planetary Sciences, Caltech, mbrown@caltech.edu.

Introduction: The large Kuiper Belt object (KBO) Eris [1] is nearly as big as Pluto [2] and has a small moon, Dysnomia [3]. Beyond this, rather little is understood about Eris. For instance, although its bulk density is known [2], its internal structure (e.g. whether it is a homogeneous ice-rock mixture) is not. Below we use recently-obtained constraints on the system spin and orbit characteristics to argue that Eris must be quite dissipative, inconsistent with a simple ice-rock mixture [4]. Differentiation was likely driven by long-lived radiogenic heat production. Unlike Pluto [5], however, we find no evidence for a subsurface ocean at Eris.

Observations: The first observation of importance is that the Eris-Dysnomia system, like Pluto-Charon, is doubly synchronous [6,7]. This in turn indicates that Eris must be quite dissipative for it to have spun down over 4.5 Gyr. The second observation is an upper bound on the mass of Dysnomia, obtained using ALMA astrometry [8]. The small size of Dysnomia (upper bound mass ratio of 0.0084 at 1-σ) places a lower bound on how dissipative Eris must be. This lower bound turns out to be unexpectedly informative [4].

Spin-orbit evolution: Driven by tidal torques, Dysnomia evolved away from Eris over geological time, and as it did so Eris spun down to eventually reach a synchronous end state [6,7]. To analyze this process we use the formulation of [9], assuming that eccentricities are zero throughout and neglecting higher-order spin-orbit resonances. Numerical integration of the equations in [9] show that Eris just becomes synchronous at the present day (i.e. after 4.5 Gyr) if \( Q/k_2 \) is 3200, consistent with the results of [7]. Here \( k_2 \) is the tidal Love number and \( Q \) is the dissipation factor. A low \( Q/k_2 \) indicates a dissipative body; the Moon’s present-day \( Q/k_2 \) is roughly 1500 [10], or twice as dissipative as Eris. Earlier synchronization of Eris is possible but would require a lower (more dissipative) \( Q/k_2 \).

Rather than assuming a constant \( Q/k_2 \), it is more appropriate to assume that \( Q \) varies approximately linearly with the forcing frequency \( \Omega - \omega \), where \( \Omega \) is the spin frequency of Eris and \( \omega \) is the mean motion. Such behaviour is characteristic of viscoelastic materials away from the dissipative peak [11].

Figure 1 shows the system evolution when \( Q \) varies linearly with forcing frequency. In this case Eris just becomes synchronous at the present day, and the starting \( Q \) (at a forcing period of 165 hours) is 6300. Eris becomes more dissipative (\( Q/k_2 \) decreases) as synchronous rotation is approached. The advantage of this frequency-dependent model is that it is easier to interpret in terms of internal structure (see below).

Because we only have an upper bound on the mass of Dysnomia [8], Eris could be more dissipative than our baseline calculations indicate, so our calculations below are all conservative.

Internal Structure: The bulk density of Eris of about 2500 kg m\(^{-3} \) [2] indicates that it is composed primarily (>85%) of rock. If Eris were a homogeneous mix of rock and ice, its material properties would therefore be dominated by the rock fraction. The resulting \( k_2 \) would be roughly 0.01, assuming a rock rigidity of 30 GPa, requiring a \( Q \) of <30 to match the constant-\( Q \) constraint.

The central temperature of Eris \( T_{cen} \), assuming uniform heat production and steady-state, is given by

\[
T_{cen} = T_s + \frac{\rho HR^2}{6k} \tag{1}
\]

where \( T_s \) is the surface temperature, \( \rho \) the density, \( H \) the heat production rate, \( R \) the radius and \( k \) the thermal conductivity. Taking \( H=4.5x10^{-12} \text{ Wkg}^{-1} \), \( k=3 \text{ Wm}^{-1}\text{K}^{-1} \) and \( R=1163 \text{ km} \), we find a central temperature of about 875 K. This is ~500 K below the melting temperature of rock, indicating that dissipation should be negligible (Q>>30).

However, equation (1) also indicates that temperatures inside Eris were high enough to melt ice and thus allow differentiation to proceed. A differentiated body is likely to be much more dissipative: ice is less rigid and more deformable than
rock, so that a near-surface ice shell will dissipate far more energy than a homogeneous rock-ice mixture.

**Differentiated Eris:** We calculate the tidal response of a differentiated Eris assuming Maxwell viscoelasticity and the method of [11]. Here we assume a rigid central rock core surrounded by an isoviscous ice shell 90 km thick, topped with an elastic ice lithosphere 30 km thick.

![Figure 2](image.png)

Figure 2. $Q/k_2$ for a three-layer Eris as a function of forcing frequency, calculated using the method of [11] and three different ice shell viscosities. The red crosses denote values used in Fig 1.

Figure 2 plots the calculated $Q/k_2$ for three different ice shell viscosities as a function of forcing period. The crosses are the $Q/k_2$ and forcing periods used in Figure 1, which demonstrate that an ice shell viscosity in the range $3 \times 10^{14}$ Pa s would provide the correct orbit evolution timescale. Since the viscosity of ice near its melting point is expected to be in the range $10^{13}$ to $10^{15}$ Pa s [12], we conclude that a warm ice shell can explain the dissipative nature of Eris.

The natural way to explain a warm ice shell beneath a rigid lid is if the ice is convecting. In this situation, the bulk of the shell will be of a roughly constant viscosity. The tendency of an ice shell to convect depends on the Rayleigh number of the shell compared to the critical Rayleigh number [13]. In the case of Eris, a 120 km thick ice shell with a viscosity $3 \times 10^{14}$ Pa s has a Rayleigh number 20 times the critical value, indicating that sluggish convection is expected. Conversely, a conductive Eris would be too cold to permit the required level of dissipation.

**Contrast with Pluto:** At Pluto, there is some evidence for a subsurface ocean and a conductive ice shell [5]. In contrast, a convecting ice shell at Eris would likely remove heat sufficiently fast that a subsurface ocean would not be expected to form. Why should Eris be so different from Pluto?

For Pluto’s ice shell to be conductive, it must be very cold, either due to the presence of clathrates in the ice shell or ammonia in the ocean [5]. Both these possibilities require the presence of volatile species (CH$_4$ or NH$_3$). If Eris lacks such volatiles, it would most likely also lack an ocean.

Eris’s rock-rich nature compared to Pluto is consistent with a relative lack of ice and other volatiles. One obvious way to lose volatiles is during a giant impact, so perhaps the Dysnomia-forming impact was more violent than the Charon-forming equivalent. Further exploration of this issue would be of considerable interest.

**Discussion and Conclusions:** Because convecting ice has a low viscosity, lateral variations in topography would be hard to maintain at Eris. Sputnik Planitia on Pluto is bright because it is a topographic low [5]; we would not expect similar features on Eris, and indeed this expectation is consistent with Eris’s relatively muted light curve. Similarly, long-wavelength departures from sphericity (e.g. an impact basin) would not be expected. Future occultation measurements of Eris’s shape [cf. 14] would thus be of great interest.

Although our results do not favour an ocean at Eris, refreezing of such an ocean would result in pressurization and potentially cryovolcanism. Detection of surface species that are unstable over geological timescales or are isotopically indicative of internal processes [15] would provide one way of testing for the presence of an ocean.

**References:**