VOLCANIC ERUPTION AND INTRUSION PROCESSES ON 4 VESTA: A REAPPRAISAL. K. Keil[1] and L. Wilson[2], [1]Hawai'i Institute of Geophysics and Planetology, University of Hawai'i, Honolulu, HI 96822, U.S.A, (keil@hawaii.edu) [2]Lancaster Environment Centre, Lancaster University, Lancaster LA1 4YQ, U.K.

Background: Extensive volcanism has been inferred [1] for the parent asteroids of the howardite-eucrite-diogenite meteorites (4 Vesta [2]), the aubrites [3], acapulcoites-lodranites [4-6], brachinites [7, 8], and ureilites [9-11]. If metal core formation in asteroids required much silicate melting [12-15], then the existence of meteorites from ~60 such cores [16] implies that at least this many asteroids formed early enough to exploit the short-half-life radionuclide ^{26}Al [17, 18] as a major heat source for differentiation.

After modeling explosive volcanic activity [19-25] we addressed melt migration in asteroid interiors [22, 26, 6] using a model of local melt accumulation and migration in isolated dikes propagating by fracturing host rocks, the fractures closing after the dike passed [27] (Fig. 1). These treatments did not consider the detailed effects on dike shape of pressure gradients linked to magma motion [28] and were developed before the full implications of the thermal histories of differentiated asteroids were appreciated.

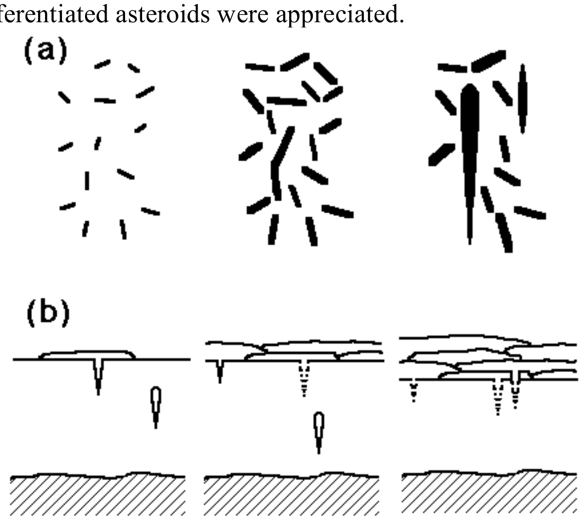

Fig. 1: Early models of (a) melt vein growth and interconnection into isolated propagating dikes and (b) crust accumulation from lava flows fed by these dikes.

New paradigm: If ^{26}Al is initially distributed near uniformly in an asteroid [29], and solid state convection does not occur in the interior at temperatures below the silicate solidus [30], the inefficiency of thermal conduction in silicates causes the temperature profile to be nearly uniform in the interior with a steep temperature gradient in the outermost layers of the asteroid [31]. The thermal boundary layer (effectively the elastic lithosphere) has thickness $B = \sim 1.4\ (\kappa\ \tau)^{1/2}$ [31] where τ is the time since asteroid formation and κ is the thermal diffusivity of silicates, ~10^{-6} m^2 s^{-1}. During the ~2 Ma lifetime of ^{26}Al, B grows to ~11 km. So all asteroids with radii > say ~30 km, have a nearly uniform temperature beneath the lithosphere. Low pressure gradients in asteroids mean that the solidus also increases only slowly with depth. As the temperature everywhere increases, melting begins over a wide range of depths at nearly the same time.

A new model of melt transfer in asteroid interiors [32] uses ophiolite evidence [33-37] showing that efficient networks of veins and dikes drain melt from partial melt zones in the Earth's mantle. In asteroids the near-uniform melting over a wide range of depths causes a similar efficient drainage network to form [32] (Fig. 2). Melt can migrate upward as fast as it is formed, with only a few % of melt being present in the mantle at any one time.

Implications: The melt formation rate in the interior of a differentiating asteroid is proportional to the heat generation rate and can be found as a function of time for any bulk composition and formation time. Values soon after the start of melting are 10-20 m^3 s^{-1} for a 50 km radius asteroid [38, 30] and ~200 m^3 s^{-1} for a radius of ~100 km [32]. The melting rate scales approximately as the asteroid radius cubed and declines exponentially with the ^{26}Al decay. Thus total magma generation rates in early-forming asteroids with radii 25-250 km will be ~3 to ~3000 m^3 s^{-1}. Buoyancy-driven melt migration must be approximately vertical, so melt from an asteroid interior cannot be transferred though a single pathway; ~5 active sites may be present on an asteroid at any one time [32]. Thus we expect peak volume flow rates at individual locations in the upper mantle, F_{MAX}, in the range ~1 to 600 m^3 s^{-1} for asteroids with radii in the range 25 to 250 km.

If melt is to be transferred directly to the surface of an asteroid, the dike carrying it must penetrate completely through the lithosphere thickness B. Rock elastic properties place a constraint on the dike's width and horizontal length and hence the volume flux through it. A second constraint is that the magma flow speed must be great enough that cooling is not excessive. A third constraint is that the pressure in the melt at the base of the lithosphere must be great enough to hold the dike open. Taken together, these factors mean that it is *not possible* to have a continuous flow of melt at a rate F_{MAX} from an asteroid interior to the surface unless the asteroid radius is greater than a critical value R_{CRIT} that varies between 250 and 300 km, depending on initial assumptions and modeling uncertainties.

As a result, the only option for virtually all differentiated asteroids is for melt to intrude at the base of the lithosphere, accumulating into large, sill-like magma reservoirs [32] (Fig. 2). Magma will be erupted to the surface from such a reservoir once the excess pressure in it is great enough to fracture the overlying lithosphere rocks and more than support the static weight of a column of magma as far as the surface.

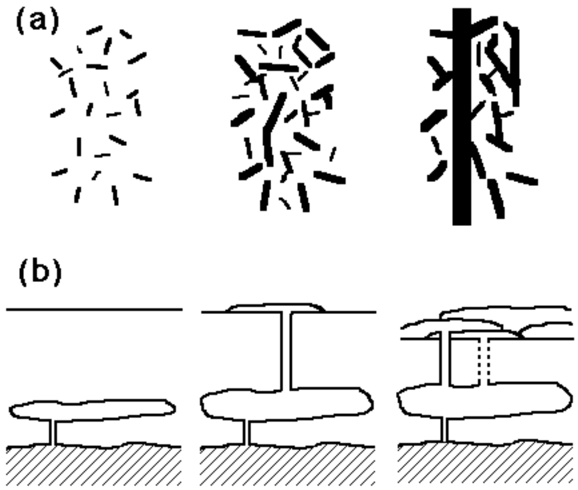

Fig. 2: New model: (a) static, mantle-wide melt vein network steadily transfers melt at low rates into major dikes feeding sill-like magma reservoirs at the base of the lithosphere. These erupt episodically to the surface at much higher magma volume fluxes.

Results: Typical asteroid eruptions involved fissures a few km long erupting magma to the surface at a maximum rate directly proportional to the asteroid radius, varying from ~4,000 m^3 s^{-1} (radius = 50 km) to ~20,000 m^3 s^{-1} (radius = 250 km). The smaller of these eruption rates is an order of magnitude greater than that commonly seen on Earth [~300 m^3 s^{-1}, Kilauea volcano, Hawai'i [39]) and rather more than that in the Laki eruption [40]. The rates for a Vesta-size body approach those of flood basalts on Earth [41] and the Moon [42] and imply lava flow lengths up to a few km (small bodies) and at least 100 km (Vesta) [26].

Explosive eruptions on Vesta at volume fluxes V of 10^3 -10^4 m^3 s^{-1} would have formed opaque fire fountains at all but the highest magma gas contents n depositing hot pyroclasts on the surface to form welded deposits [23]. Table 1 shows how the fractional deposit area that is hot depends on V and n.

Summary: Our new results support our earlier findings about sizes of lava flows and pyroclastic deposits expected on the surface of 4 Vesta. However, they argue against a magma ocean, suggesting instead that surface eruptions were fed by magma from large sill-like intrusions at the base of the lithosphere.

Table 1. Explosive eruption deposits on 4 Vesta. *n:* magma gas content, *u*; eruption speed, *R*: deposit radius, *A*: fraction of deposit area covered by hot clasts.

n /ppm	*u* /(m/s)	*R* /km	*A* if volume eruption rate is 3×10^3 m^3/s	1×10^4 m^3/s
100	20	1.6	99.8	99.9
300	34	4.4	99.0	99.7
1000	62	15	95.1	98.5
3000	106	43	75.5	92.3
10000	194	144	4.3	58.0

References: [1] Keil K. (2002) *Asteroids III*, 573. [2] Binzel R. P. & Xu S. (1993) *Science*, *260*, 186. [3] Keil K. (2010) *Chemie der Erde-Geochem.*, *70*, 295. [4] McCoy T. J. et al. (1996) *GCA*, *60*, 2681. [5] McCoy T. J .et al. (1997a) *GCA 61*, 623. [6] McCoy T. J. et al. (1997b) *GCA 61*, 639. [7] Warren P. H. & Kallemeyn G. W. (1989) *Proc. LPS Conf. 19*, 475. [8] Nehru C. E. et al. (1996) *Meteoritics 27*, 267. [9] Goodrich C. A. (1992) *Meteorit.*, *27*, 327. [10] Goodrich C.A. et al. (2001) *GCA*, *65*, 621. [11] Goodrich C. A. et al. (2002) *LPS XXXIII*, Abstract #1379. [12] Stevenson D. J. (1990) *Origin of the Earth*. OUP, 231. [13] Taylor G. J. (1992) *JGR*, *97*, 14,717. [14] Minarik W. G. et al. (1996) *Science 272*, 530. [15] Shannon M. C. & Agee C. B. (1996) *GRL*, *23*, 2717. [16] Chabot N. L. & Haack H. (2006) *Meteorites & the Early Solar System II*, 747. [17] Wadhwa M. et al. (2006) *Meteorites & the Early Solar System II*, 715. [18] McCoy T. J. et al. (2006) *Meteorites & the Early Solar System II*, 733. [19] Wilson L. & Keil K. (1991) *EPSL*, *104*, 505. [20] Muenow D. M. et al. (1992) *GCA*, *56*, 4267. [21] Keil K. & Wilson L. (1993) *EPSL*, *117*, 111. [22] Wilson L. & Keil K. (1996a) *EPSL*, *140*, 191. [23] Wilson L. & Keil K. (1997) *MAPS*, *32*, 813. [24] Wilson L. et al. (2010) *MAPS*, *45*, 1284. [25] Keil K. et al. (2011) *MAPS*, *46*, 1719. [26] Wilson L. & Keil K. (1996b) *JGR*, *101*, 18927. [27] Weertman J. (1971) *JGR*, *76*, 1171. [28] Lister J. R. & Kerr R. C. (1991) *JGR*, *96*, 10,049. [29] Lodders K. (2003) *Astrophys. J.*, *591*, 1220. [30] Mokovitz N. & Gaidos E. (2011) *MAPS*, *46*, 903. [31] Carslaw H. S. & Jaeger J. C. (1952) *Conduction of Heat in Solids. Secn. 9.8-I*, Clarendon. [32] Wilson L. et al. (2008) *GCA 72*, 6154. [33] Nicolas A. (1986) *J. Petrol.*, *27*, 999. [34] Sleep N. H. (1988) *JGR*, *93*, 10255. [35] Hart S. R. (1993) *Proc. Nat. Acad. Sci. U.S.*, *90*, 11,914. [36] Maaloe S. (2003) *J. Petrol.*, *44*, 1193. [37] Valentini L. et al. (2007) *JVGR*, *159*, 355. [38] Hevey P. J. & Sanders I. S. (2006) *MAPS 41*, 95. [39] Heliker C. et al. (2003) *USGS Prof. Pap. 1676*, 206 pp. [40] Thordarson Th. & Self S. (2003) *JGR*, *108*, #4011. [41] Swanson D. A. et al. (1975) *Amer. J. Sci.*, *275*, 877. [42] Wilson L. & Head J. W. (2008) *LPS XXXIX*, Abstract #1104.