DEFENDING HUMAN CIVILIZATION FROM SUPERVOLCANIC ERUPTIONS

Brian H. Wilcox, Karl L. Mitchell, Carolyn E. Parcheta,

Florian M. Schwandner, Rosaly M. Lopes

Jet Propulsion Laboratory,

California Institute of Technology

ABSTRACT

Large volcanic eruptions greater or equal to a magnitude 8 on the Volcanic Explosivity Index (i.e., supervolcanic eruptions) eject >10^{15} kg of ash and sulfate aerosols, sufficient to blanket sizeable fractions of continents and create a regional or global "volcanic winter." Such events could seriously reduce worldwide agricultural production for multiple years, causing mass famine. Supervolcanic eruptions occur more frequently than large asteroid or comet impacts that would have a similarly catastrophic effect to human civilization, especially now that many asteroid orbits have been mapped. We assess whether future supervolcanic eruptions could be dampened, delayed, or prevented by engineering solutions.

1. INTRODUCTION

There has been a significant effort over the last two decades to determine what threats exist to humanity on Earth and how to mitigate them under the umbrella term *planetary defense.*[1] Notable among these are the threats of asteroid and comet impacts and supervolcanic eruptions. For reference, a 2 km impactor would cause worldwide disruption similar to that of geologically mapped supervolcanic eruptions that have produced volcanic winters.[2] The efforts of astronomers worldwide have resulted in identification and successful orbit mapping of more than 98.3% of near-earth asteroids (NEAs) with diameter greater than 2 km,[3,4] (e.g. large enough to cause global climatic effects)[5] while considerable attention was given by other scientists, engineers, and hazard management coordinators to the study of major asteroid collisions. This effort concluded:

1) >2 km asteroid impacts occur half as often as supervolcanic eruptions,[6,7] and
2) no known asteroids will threaten Earth for at least a century.[8]

Since over 1.3% of these 98.3% of threatening NEAs were discovered in the latest two years,[9] it is reasonable to believe that all will be discovered in the relatively near future. Long-period comets are estimated to be <1% of the total threat,[10] and so do not add appreciably to the number. Given ~100 years warning for a NEA impact, and the considerable attention now focused on asteroid deflection techniques, it is also reasonable to expect that such collisions can be prevented in the future.[11] We propose that corresponding efforts be considered towards supervolcanic eruptions.

In this paper, we attempt to assess whether engineering solutions may dampen, delay, or prevent the negative effects of future supervolcanic eruptions on human civilization. We open-mindedly postulate the controversial hypothesis: *A system can be engineered to efficiently mitigate a supervolcanic, eruption-induced volcanic winter (regional or global) that would otherwise lead to mass starvation and a major population decline*.

2. BACKGROUND AND STARTING ASSUMPTIONS

While not all supervolcanoes are fed by mantle plumes from below, the most famous supervolcano, Yellowstone (Wyoming, USA) is by now unambiguously believed to be associated with a mantle plume. We recognize that mantle plumes and their existence is still a hotly debated topic.[12,13] We will use the term *mantle plume* less restrictively to simply mean a large volume of a spatially coherent magma (much greater than the minimum erupted volume of 1,000 km^3) that is transported toward the surface where some of that magma can erupt.

Explosive supervolcanic eruptions: The phrases *supervolcano* and *supervolcanic eruption* were widely introduced to the public by the BBC in 2000 to describe cataclysmic eruptions capable of *plunging the world into a catastrophe and push humanity to the brink of extinction*.[14] These eruptions are as rare as they are excessively catastrophic – a supervolcanic eruption has not occurred in the Holocene (past approx. 12,000 years) and, therefore, modern human civilizations have not witnessed such an eruption. Our knowledge of them comes strictly from the geologic record, and the impressively large deposits these eruptions produce. Supervolcanic eruptions register at magnitude 8 (10^{15} kg erupted) on the logarithmic Volcanic Explosivity Index (VEI) scale,[15] and correlate to ejected tephra volumes >1000 km^3 (>450 km3 dense rock equivalent (DRE) of stored magma).[16] Sparks et al., 2005 [17] (and references therein) concluded the same findings as did the astronomers mentioned in the introduction, namely that supervolcanic eruptions would threaten humanity's existence as they would dwarf the largest eruptions witnessed thus far. Laki (1783 Iceland),[18,19] Tambora (1815 Indonesia),[20,21] Krakatau (1883 Indonesia),[22,23] Novarupta (1912 Katmai, Alaska),[24,25] and Pinatubo (1991 Philippines)[26] are just a few large ultraplinian eruptions (VEI 6-7) that produced hemispheric climate anomalies several years in duration. An even larger, supervolcanic eruption would have a more noticeable global effect on climate, agriculture, and human infrastructure.[27,28,29] People living near the eruption would likely perish from roof collapses after approx. 10 cm of wet ash, which exceeds most critical rooftop loads.[30] Survivors living farther away would need to survive extreme famine or outright starvation for months, several years, or longer.[31]

What is a caldera volcano? Caldera volcanoes are not the "typical" volcano morphology – instead of being a positive landform (i.e. a cone protruding up from Earth's surface), calderas are negative landforms (i.e., ground that has sunk down below the surrounding surface due to previous eruptions or magma withdrawal). Calderas were first geologically defined by Leopold von Buch[32] and are now classified as volcanic collapse craters exceeding 1 km in diameter, having a depth that is shallower than their width, and can be circular, oval, or more complexly shaped depending on its origin and eruptive history. Calderas are currently viewed as the surface expression of collapsed magma reservoirs, and thus mostly reflect the extent to which the magma reservoir extended prior to its emptying. Regardless of what caused magma to leave the reservoir system (eruption or withdrawal for lateral intrusions), the network of rock remaining is not sufficiently strong enough to support the overlying coherent rock burden, so that overburden collapses and compresses the underlying rock over hours to days. For example, the 1991 VEI 6 Pinatubo caldera collapse likely lasted at least 6 hours.[33] There is evidence for 3 caldera forming mechanisms:

- Piecemeal – where the ground surface breaks up into smaller, potentially nested pieces as it descends,[34]

- Piston – where the whole ground surface lowers as a cohesive unit,[35] and
- Trapdoor – where the ground surface lowers and tilts from different degrees of drainage.[36]

Supervolcanoes are associated with silicic calderas, where the ground surface either slowly or catastrophically drops like a piston as the explosive eruption occurs.[37] In this study, we address two of the three known supervolcanic calderas in the U.S.: Yellowstone (Wyoming, 50 x 80 km, approx.0.8 km deep), Valles Caldera (New Mexico, approx. 22 km x 22 km, approx. 0.4 km deep, not discussed further in text), and Long Valley Caldera (California, 16 x 32 km, approx. 1.2 km deep).

2.1 CALDERA MAGMATIC RESERVOIRS AND SUPPLY RATES

Magma reservoir: *Magma* is a plastic mixture of melt, gases, and solids. Every magma reservoir resides, in part or in whole, underneath the volcano it produces. In the past, these reservoirs have been referred to as *magma chambers* and they were generally envisioned as a single, kilometer-scale, long-lived, ellipsoidal shaped, magma-filled cavity. However, traditional geophysical techniques are unable to unambiguously identify locations of significant melt volumes.[38,39] Magma "chambers" likely are neither true chambers,[40] or a large network of dikes and sills,[41] nor a crystal mush,[42] but a complex combination of all these. Regardless of geometry, however, the conductive heat transfer away from any of these reservoir geometries will appear as if it was coming from a single point source once measured at a sufficient distance from the magma reservoir. Therefore, we will calculate our engineered solutions around a single magma chamber for simplicity, but describe our work in the reference frame of a magma "reservoir" fed by a mantle plume.

Yellowstone case study: The shallow magma reservoir of Yellowstone is thought to comprise of at least 2 different bodies underneath the caldera.[43] The inferred rhyolitic volume of Yellowstone's magma reservoir is 15,000 km^3 [44] and resides between 5 and 16 km depth.[45] Huang et al. (2015)[46] give an updated estimated melt fraction and magma body size for Yellowstone from the variation in P-wave velocities. They found that a shallower unit has approx. 9% residual rhyolitic melt fraction in an approx. 10,000 km^3 body located at approx.4 -14 km depth and the deeper unit has a approx. 2% basaltic melt fraction in a approx. 46,000 km^3 body located at approx. 20 - 45 km depth. The current surface power output is 4.5 – 6 GW or approx. 2000 mW/m^2,[47] which is larger than surrounding regions. The deeper basaltic reservoir is thought to accumulate 0.3 km^3/yr (steady state average) of new material (including heat and gas components) based on CO_2 measurements.[48] This intrusion rate of 0.3 km^3/yr, would produce 22 GW of power if cooled completely without new intrusions. [49,50]

2.2 HEAT TRANSPORT AND DISSIPATION IN CALDERA SYSTEMS

Heat supply into the magma chamber from below: The heat and magma supply from the mantle plume into the magma chamber is likely not steady but periodic. Pulses of thermal energy could arrive, for example, if large "blobs" of basaltic magma were to periodically intrude upward into a silica-rich magma chamber. This is what would be expected from mantle plumes based on Rayleigh-Taylor instabilities.[51] The higher melting temperature of the basaltic magma can carry substantial heat energy up into the lower-temperature (but more viscous and more explosive) rhyolitic magma.

Heat transport mechanisms: There are three ways to move heat vertically from the mantle plume and lower basaltic reservoir into the upper silicic reservoir and out to the surface: conduction, advection, and convection. Between the deeper magmatic reservoirs, both heat conduction and advection (magmatic and/or gas) may dominate. From the uppermost magmatic body to the surface, all three mechanisms of heat transfer are involved. Convection plays a role in bringing heat upward within a magma reservoir. Advectively, ascending decoupled gases transport heat to the surface. On a larger scale, hydrothermal fluids efficiently transport heat by convection.

Magmatic gases: The dominant gas phases that exsolve from magmas are H_2O and CO_2, with minor components of SO_2, noble gases (mostly He), and halogens.[52] CO_2 and the noble gases are non reactive in geo/hydrothermal environments; the acid gases readily dissolve and contribute to deep fluids' chemistry and associated mineralizations. Like H_2O, CO_2 is a molecule with a strong dipole, which gives it a high heat capacity.[53] In active volcanic environments, CO_2 flux and heat flow are known to correlate, indicating that advective gas heat transport plays a dominant role in active volcanic environments.[54,55]

Magmatic heat loss – the rate of cooling within the silicic magma reservoir – competes with the rate of upward heat dissipation mechanisms of conduction, gas advection, and hydrothermal system convection. If these rates are balanced, or if the upward heat dissipation is at a larger rate than the magmatic heat loss at depth, then the magma reservoir cools in a naturally controlled way, with little risk of a major eruption.[56] Magmatic, hydraulic, or seismic forcing can alter this balance, causing a caldera system to become restless, and potentially unstable enough to erupt violently.

Magmatic forcing occurs when new magmatic heat is supplied from intrusions that are typically new basaltic magma ascending and mixing into the silicic reservoir.[57,58] The result of the additional heat and chemical material available may lead to an enhanced probability the silicic magma reservoir erupting, shifts in the overlying hydrothermal system, and increased precursory activity such as earthquakes, gas emissions, and more vigorous hydrothermal activity.[59,60] Magmatic forcing may cause an upward shift of the phase change isotherms and possibly even to a partial *drying out* of the hydrothermal system by reducing or removing the convective heat transfer mechanism in favor of conduction and gas advection. *Hydraulic forcing* may occur if the hydrothermal system cannot sustain a stable water volume and experiences a net water loss.[61] *Seismic forcing* occurs when earthquakes disrupt the hydrothermal system.[62,63] Earthquakes can create a pathway for magma or increase hydrothermal reservoir temperatures. All three forcing mechanisms occurred prior to the 1991 Pinatubo VEI 6 caldera-forming eruption. A basaltic intrusion preceded the eruption,[64] hydrothermal exploration was abandoned due to too corrosive reservoir conditions,[65] and seismic strike-slip faulting was detected above the magma reservoir directly prior to the climactic June 15th 1991 eruption.[66]

Hydrothermal fluids: Hydrothermal fluids are a multicomponent electrolytic solution that can derive from combining both infiltrated surface or groundwater (in hydrology this is called *meteoric* water, i.e., meteorologically derived) and a varying admixed component of magmatic fluids and gases.[67] The deepest, more magmatically dominated fluid compositions are highly acidic, hot, gas-rich, extremely corrosive fluids left over from volatile exsolution of magma (*magmatic fluids*). *Geothermal fluids* are meteoric waters which have interacted with hot rock material in magmatic environments, which greatly enhances its acidity and increases the load of solutes derived from the rock. Solutes induce a brine-like, corrosive, and occasionally supercritical character that can dissolve and transport even unreactive elements like gold.[68, 69]

Mixtures of meteoric water, magmatic fluids, and geothermal fluids with independently ascending gases can be encountered in most magmatic-hydrothermal systems.[70] Localized and rapidly changing thermodynamic conditions at subsurface mixing interfaces can lead to the formation of mineralized veins that seal or line (scale) subsurface fluid conduits, changing the chemistry of the fluid itself.[71,72] Surface expression of all these fluid types are known, however pure magmatic fluids and deep geo-/hydrothermal brines are rarely encountered directly at the surface. Roughly concentric zonation of conditions around volcanoes exist and lead to a gradual outward reduction in the magmatic component and heat flow. Conceivably, magma may rise up to shallower depths in large calderas months to years before a supervolcanic eruption. Analogous to volcanic systems dominated by magmatic fluids with shallow magma reservoirs (less than 2 km) central fracture-bound advective gas conduits may develop with magmatic gas temperatures near the surface.

Yellowstone case study: Yellowstone is situated in a tectonically active environment, and areas of major fracturing feature more acidic and gas-rich hydrothermal surface manifestations[73] than other areas. An estimated 50% of the gas coming out of

Yellowstone's hydrothermal field is derived from the degassing of CO_2 from the deeper basaltic intrusions. [74] The hydrothermal system at Yellowstone has surface manifestations of several types of hydrothermal fluids, including confined gas-poor neutral chloride waters (causing geysers as well as silicate and carbonate mounds), and gas-rich acid-sulfate hydrothermal.[75] Most of the water in the hydrothermal system is of meteoric origin and has likely taken decades to travel to and within the hydrothermal system.[76]

The temperature reported at a depth of 1 km at Yellowstone is 310°C,[77] which is very close to the boiling temperature associated at that hydrostatic pressure depth.[78] Any higher thermal gradients would exceed the *boiling point with depth curve* at these conditions and cause water to flash to vapor, thus driving any water column upwards until the vapor bubble reaches the surface (hence the many geysers in the area). We expect that supervolcanoes in water-rich areas will have thermal gradients over the magma chamber that are limited by the boiling point with depth curve, and heat fluxes that are dominated by convective hydrothermal transport.

It is estimated that the current natural heat flux out of the Yellowstone magma chamber is between 4.5 and 6 GW, with up to two-thirds of this heat flux associated with convective hydrothermal activity, as opposed to simple conduction through the rock.[79] In essence, this previous work on Yellowstone's heat budget suggests that gas and hydrothermal fluids efficiently remove heat from the magma reservoirs. The surrounding alpine setting as well as Yellowstone Lake ensure a steady supply of excess meteoric water for the hydrothermal system to maintain itself.

Long Valley Geothermal power plant:

At Long Valley Caldera in California, hydrothermal heat is harvested to generate electrical power. Two hydrothermal water reservoirs coexist here: A deeper brine in contact or very close to the magma reservoir, and shallow meteoric water that feeds surface hydrothermal vents. In this case, the two water bodies are not in contact due to an impermeable geologic unit. We assume that in both case study locations that the position of the boiling point with depth curve and with it the geothermal gradient changes depending on whether the two fluids mix, which would cause density and chemical changes.

2.3 ERUPTION MECHANISMS OF CALDERA SYSTEMS

Pre-existing structural features: Many (but not all) volcanic centers are situated along pre-existing cross points of regional fault systems.[80] These may become reactivated either by regional seismic activity providing a lower energy structural pathway for magma ascent, or by magmatic ascent triggering fault movement.

Fracture mechanics: If the pre-existing rock that the magma intrudes into (geologically termed *country rock*) is cooler than the magma's solidus temperature and the lithostatic pressure is in the brittle regime, then the country rock will behave in a brittle manner, allowing dikes to crack the rock and propagate towards the surface. The solidus (ductile to brittle transition) of silicic and basaltic rock is approx. 400° C and approx. 700° C, respectively. Initial, small-scale fractures within igneous rocks form as a result of cooling joints located at areas of the least compressive thermal stress.[81,82,83] These fractures may grow into a larger single fracture or a fracture network, in which they could A) remain inactive, B) become a fault plane, or C) allow magma to intrude and form a dike. Other types of fractures can be produced in rocks after a rock's formation, and these are a result of changing stress fields within the local or regional area due to tectonic forces, crustal unloading, and magmatic injections. These fractures will form along areas of either the greatest tensile stress or greatest tensile areas within a shear stress. Fractures, whether they become faults or dikes, will propagate along a path where their width opens in the direction of the most tensile, or least compressive, stress.[84] This stress can, and does change within volcanic edifices, both of positive (cone) and negative (caldera) constructions. As a result, the trajectories of fractures (both faults and dikes) adjust to the changing stress field, and are predetermined on a broad scale by the load of the edifice[85] if the stress field is accurately known. At caldera volcanoes, ring faults form around the edge of the caldera, which are due to two main factors: 1) cracks forming as a result of doming and tension[86] from the underlying magma reservoir inflating, or 2) near vertical or outward dipping tension fractures that merge together and become a shear fracture during a piston-like collapse of the surface rock into an emptied/emptying magma reservoir.[87]

Eruption trigger mechanisms: Supervolcanic eruptions tend to involve "monotonous intermediate" magmas (e.g. dacite and rhyolite), which are more viscous due to the high silica content, able to dissolve up to 7 wt% of volatiles, and are crystal rich (>35%).[88] Supervolcanic eruptions can be triggered internally or externally, but in both cases magmatic volatiles are the primary driver of an eruption's intensity. Internal triggering mechanisms include dike propagation from the magma reservoir to the surface, volatile super-saturation (and subsequent exsolution) during crystal growth, or injection of hot, fresh, basaltic magma that triggers the monotonous intermediate magma into an eruption. Dikes can propagate to the surface by creating their own fractures (internal trigger) or by following pre-existing ones (external trigger).[89,90] A rapid crystallization event could induce volatile super saturation of the crystal mush, forcing a bubble nucleation event that would over pressurize the reservoir and promote dike propagation to the surface.

Thermomechanical analysis of internal eruption triggering by Gregg et al. (2012)[91] found that the thermal "halo" around the magma reservoir brings the country rock into the ductile regime, which buffers the increasing over pressure by increased volume (due to volatile exsolution or injection of new material). An example of an external eruption trigger is a fault propagating from the surface down through magma reservoir roof, which then acts as the onset location for a supervolcanic eruption.[92] The thermo-visco-elastic numerical model used by Gregg et al (2012)[93] found that supervolcanic eruption onset is a function of the overburden rock thickness above the magma reservoir and the location of the ductile to brittle transition within that overburden.

Eruption mechanics: Some of the largest historic caldera-forming eruptions have occurred over weeks or months, and consist of a series of very large eruptions in sequence, culminating in a climactic eruption which often begins to wane at the onset of partial or full caldera collapse. Supervolcanic eruptions are thought to occur from one or several vents in one of two ways: within the caldera or along the entire boundary of the caldera, as a ringed vent. Generally, the magma is kept pressurized within the magma reservoir, where the volatiles remain in solution unless magma propagates towards the surface or a fault propagates downward and reduces the effective pressure at the top of the magma reservoir. Exposure to lower pressures during magmatic ascent allows volatiles to come out of solution, which gives the magma a buoyancy force (and fragmentation force) if the gas is trapped in highly viscous magma. As magma reaches the surface via dike or other conduit, and the internal bubble pressure exceeds the external magmatic pressure, the volatile bubbles will attempt to grow in size until the bubble pressure increase is sufficient to fragment the magma.[94,95] This fragmentation process produces a column of ash that can reach tens of kilometers in height - the more intense the process is, the higher the column of ash will rise. Supervolcanic eruptions would exceed the mass eruption of large historic eruptions such as Laki, Tambora, Krakatau, and Pinatubo – all of which ejected gas and aerosols through the tropopause into the stratosphere (>8-20 km, depending on latitude). This extremely high eruption column height and fine particle injection into the atmosphere is the sequence of events we seek to prevent.

2.4 HUMANITY DURING VOLCANIC WINTER

A large volume of ash (>100km^3) erupted over a period of days to months can cause (and has caused in the past) a regional "volcanic winter": a period of low or failed agricultural production as a result of excess atmospheric particulates, which diminish plants' ability to photosynthesize[96]. Since supervolcanic eruptions have geologically mapped units of >1000 km^3, the effect of agricultural failure would be significantly more widespread. Starvation is therefore a major concern when considering hazards from supervolcanic eruptions because prolonged volcanic winter could easily surpass civilization's amount of stored food worldwide. In fact, the U.N. Food and Agriculture Organization estimated the 2012 worldwide food storage to last for 74 days.[97] The effects of a volcanic winter could easily prohibit civilization from having enough food for the current population, let alone replenish the stored food currently saved up. In light of this, we pursued the following calculations to determine if it is possible to prevent such a cataclysmic eruption, or mitigate the eruption's intensity to dampen the effects of its hazards.

It has been reported that sulphate aerosols are of overwhelming importance in determining the volcanic winter effects of an eruption.[98] However, we will assume that the magnitude of sulphate aerosol released (>10^{15} kg of particulates and sulphate aerosols) is strongly correlated with the amount of ash released, and thus we will focus mainly on the amount of geologically documented ash for the Yellowstone super-eruptions in our calculations.

3. STATEMENT OF THE PROBLEM

The prospect of attempting to prevent a supervolcano from erupting seems daunting and perhaps impossible. For our case study, we have focused on Yellowstone to inform our understanding of possible supervolcanic eruption mitigation. It has been suggested that the hydrothermal circulation at Yellowstone may cool the underlying magma and may lead to decreased long-term volcanic hazards.[99] We aim to determine whether this is true and if it could be anthropogenically enhanced to dramatically reduce the risk of a major, VEI 8, cataclysmic supervolcanic eruption.

4. BOX MODEL CALCULATIONS: THE "LEAKY BATHTUB"

Yellowstone has erupted 3 times in the past approx. 2.1 million years, with total ejected ash approaching 10^{15} kg. While the heat energy associated with this ash is some 10^{23} J,[100] this averages to only approx. 1.5 GW continuous magmatic thermal power input to the volcano over the span of these eruptions, and is well within the capacity of humans to safely introduce heat into the environment – a typical electrical power plant commonly rejects more heat than this. However, this assumes that heat is provided to the magmatic system in a steady state manner, which is almost certainly not valid, although the degree to which it is invalid is unknown.

We consider a very simple model of the supervolcano by analogy with the "leaky bathtub" problem: a bathtub is filled by a faucet but also has a leak – how long does it take for the bathtub to fill? In this case the magma chamber is the bathtub, and the amount of heat energy in the magma is the "water in the bathtub". The faucet represents the rate of heat flow into the magma chamber from below. The leak in the bathtub is the heat flowing from the magma chamber to the surface by some combination of conduction, advection, and hydrothermal convection. In our Yellowstone example, the heat leak is 4.5 to 6 GW, but the average ejecta from the past 3 eruptions carried only one-third to one-quarter in excess of that. So this simple model suggests that, if we were to increase the steady-state heat leak out of the magma chamber by approx. 35%, the volcano would never erupt.

We assume a density of 2,700 kg m^{-3}. We can estimate the time that it will take for a long chilled hole to cool an expanse of rock: for a radius R around the hole, the heat capacity is proportional to R^2 and the thermal gradient is inverse with R, so the time to cool to distance R is approximately proportional to R^3. A one dimensional array of holes will have a thermal wave that radiates from each hole and then merges to spread out as a plane wave. For distances large compared to the hole separation, the heat capacity will grow proportional to R and the thermal gradient will drop inverse with R, so the time to cool will grow approximately proportional to R^2.

If we are to use the thermal conductivity of the rock (as opposed to hydrothermal transport) to bring heat to these cooled holes, a major issue is the thermal time-constant of the rock. The sorts of silica-rich rock associated with supervolcanoes (e.g. rhyolite) typically have a thermal conductivity of 2.6 W m^{-1} K^{-1}. The heat capacity is 23,000 J kg^{-1} K^{-1} (averaged over the range of temperatures up to molten rhyolite).[101]

We can estimate the absolute time required for this process. For a cube whose side is length x (in meters), and that has a temperature difference across parallel sides of 1,000K (e.g. the difference between molten rhyolite and surface temperatures), the mass will be $2700x^3$ kg, the heat capacity will be $23{,}000*2700x^3$ J K^{-1}, and the thermal gradient will be $1000x^{-1}$ K m^{-1}. With again a thermal conductivity of 2.6 W m^{-1} K^{-1}, the characteristic time for the block to cool will be $(23{,}000*2{,}700/(1{,}000*2.6))*x^2$ seconds, or about 24,000 x^2 seconds. For x=1,000 meters, this is 760 years. For x=5,000 meters, it is 19,000 years.

These long time constants are a concern because, although they are short compared to the time between supervolcanic eruptions of a given volcano, they are long compared to normal human decision cycles. A major concern with planetary defense has been that deflecting asteroids away from a future collision with Earth can take decades, but decision processes of governmental action often delays a decision until the last possible moment. For supervolcanoes, the characteristic times for action may be orders of magnitude longer than that for asteroid deflection, and the time and magnitude of the catastrophe will be much harder to precisely specify. This is a challenge more analogous to climate change than to asteroid impact in that regard.

This first estimate indicates that the task is not hopeless. Therefore, let us consider possible methods of draining heat away from a large caldera magma chamber complex.

5. POSSIBLE SOLUTIONS

5.1 DRILLING HOLES OVER THE TOP OF THE MAGMA CHAMBER

The first approach we modeled was to drill an array of holes over the top of the magma chamber, as deep as technology or risk-aversion to a possible triggering of an eruption would allow, and circulate cold water through the holes. In the case of Yellowstone, the top of the magma chamber is thought to be about 6,500 meters below the surface.[102]

Figure 1 shows the steady-state temperature distribution of an array of such convective loops on a spacing of 500 meters, simulated with 10x10 meter grid spacing. The upper figure shows an expanded temperature scale from 475K to 525K over a rectangular region spanning the horizontal distance between the loops (500 m) from a depth of 1,020 m to 1,390 m. The "top" leg of the loop is contained in the center of the cells at a depth of 1,230 m on the right and left extremes of the span. Similarly, the bottom figure shows an expanded temperature scale from 660K to 710K around the "bottom" legs of the loops from a depth of 2,680 m to 3,040 m, with the hot pipe in the center of the right and left cells at 2,870 m depth. The assumptions of this model are that the original temperature

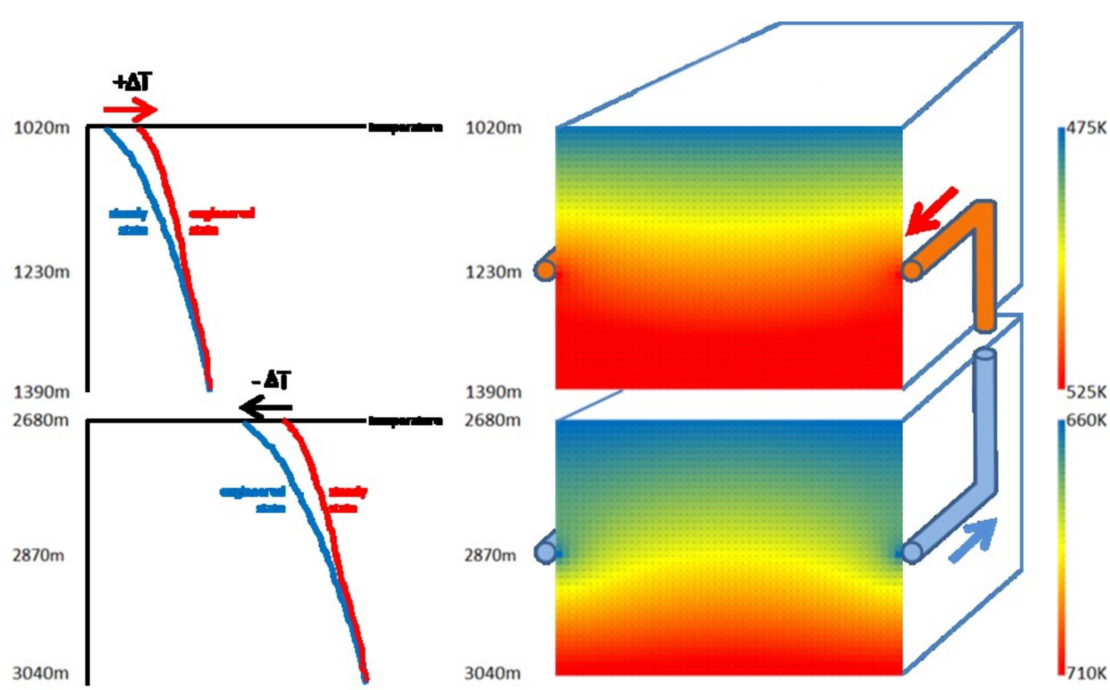

Figure 1: Steady-state temperature distribution for array of convective loops designed to passively cool the Yellowstone magma chamber. See text for details.

distribution was a uniform gradient from the surface at 300K to the magma chamber at 1,370K at 6,500 m depth. The loops are assumed to be 3,000 m long horizontally, such that natural convection is sufficient to drive the flow and carry sufficient heat with modest temperature drop. A constraint is enforced that nowhere in the convective loops can the water temperature exceed the boiling point with depth curve, as discussed previously, allowing the system to use only natural pressurization and to be refilled seasonally from groundwater near the surface. This constraint limits the depth and separation of the top and bottom legs of each loop, thereby limiting the effectiveness of the approach. The original thermal gradient above the magma chamber is (1,370K-300k)/6,500m = 0.165 K/m. The steady-state thermal gradient after installation of the convective loops is 0.190 K/m, a 15.4% increase over the natural flux.

Clearly this does not achieve our desired approx. 35% increase in the flux, and also is perilously close to the boiling point with depth curve with the attendant risk that water could flash to vapor anywhere in the system. If any vent or leak allowed sufficient water to escape such that the natural convection was interrupted, the whole system would presumably rapidly boil out, and it would be unlikely to be restored to operation with the introduction of any reasonable amount of new water flowing into the now-superheated pipes.

This simulation highlights a major problem with this approach – that the temperature gradient in the immediate vicinity of the pipes in the convective loop is a substantial bottleneck in the flow of heat. Here the exterior of the 10x10 m cells containing the bottom pipes of the loop are at 660K, while the exterior of the 10x10 m cells containing the top pipes of the loop are at 525K. Yet the actual ΔT of the water required to drive adequate heat transport and convective flow around the loop is only 15K. The rest of the 135K temperature difference is the thermal gradient within the 10x10 m cells containing the convective loops. This large thermal gradient is another consequence of the poor thermal conductivity of rock, representing a substantial impediment to the approach of using fully-contained convective flow as an effective method for extracting heat from the system.

While it is conceivable that reducing the 500 m separation of adjacent loops could achieve our goal of approx. 35% increase in heat flux out of the magma chamber, there is another issue that needs to be addressed. That issue is the relatively short lifetime of all known engineering materials in the hot brines and acids associated with the groundwater over magma chambers in areas like Yellowstone. The geothermal energy industry has been researching and seeking engineering materials that can survive such environments for sufficient durations to achieve acceptable return on investment, and has found that the "normal" materials one would choose for such purposes (e.g. "marine" stainless steel alloy

316) frequently are compromised by the hot corrosive brines and their acids in a matter of days.[103] Other materials considered for extreme environments fare similarly poorly.[104] Two adverse phenomena are associated with deep hydrothermal brines: scaling (mineral deposits) and corrosion. Hot and in some cases supercritical hydrothermal fluids interact with the hot rock material they circulate through and leach components out of the rock (*hydrothermal alteration* of the rock) and into solution. By the same process, these brines acidify by water-rock interaction. Hydrothermal brines are capable of dissolving, transporting, and depositing even gold,[105,106] and both scaling and corrosion are thought to be interrelated.[107] In volcanic-hydrothermal systems, different from high heat flow geothermal systems without acid gas components, brine physiochemistry determines corrosion behavior and choice of materials (e.g., this type of steel has Molybdenum in it for increased chloride resistance[108]). Both scaling and corrosion is affected by advecting and sparging (flushing) of CO_2 from the underlying magmatic body through carbonate deposition and cement corrosion (a factor to be considered when plugging wells).[109] The choice of engineering materials is a highly dynamic field of research.[110,111]

Deposition (scaling and veining) typically occurs when hydrothermal fluids suddenly expand into lower pressure areas, or when these fluids suddenly mix with colder meteoric fluids. In geothermal wells, pressure control devices may experience downstream depressurization and scaling, often as calcium carbonate,[112,113] but also as sulfates and sulfides.[114] Microbial deposition of extremophile organisms has only recently been considered as a scaling factor in mineral deposition.[115,116] Avoiding pressure drops at orifices is a concern – if the fluid doesn't separate (flash), then scaling might still happen if brines are used in production rather than neutral-chloride geothermal fluids. For that reason, geothermal exploration in highly active volcanic regions is costly and risky, and exploration drilling assesses the fluid chemistry in detail.[117] In geothermal operations, scaling inhibitors are added to the fluid, including pH adjusting agents.[118]

Corrosion occurs when brines are not yet saturated with respect to available potential solutes, and for metals specifically, complexation by organic and inorganic ligands plays a major role in dissolution and transport.[119] In geothermal exploration, acid-sulfate hydrothermal reservoir fluids, usually representing a deeper and hotter part of the hydrothermal convection complex, are seen as highly disadvantageous because they corrode typical engineering materials in very short time. On the magmatic gas dominated end of compositions, silica monolayers on stainless steel have shown promising corrosion resistance at pH <1.[120]

5.2 DRILLING HOLES AROUND THE PERIMETER OF SUPERVOLCANOS – SUCCESSIONS OF CONCENTRIC

RINGS OF HOLES AS THERMAL WAVE PROPAGATES INWARD (OVER THOUSANDS OF YEARS)

An alternative approach is, instead of having a 2-D array of convective loops over the top of the magma chamber, to have a 1-D array of pipes that would be constructed around the perimeter of the magma chamber. This has the advantage that the hot brines and acids formed over the magma chamber would be less likely to be encountered during drilling, assuming one drills a sufficient distance off the side of the magma chamber to avoid drilling technology challenges.[121]

The geothermal energy business has already addressed many of the key technical issues. One of the most popular approaches to modern geothermal energy extraction is "Enhanced Geothermal Systems" (EGS). An exhaustive compendium of information about EGS was compiled by an interdisciplinary panel led by MIT that studied future prospects for EGS as a source of energy on behalf of the DOE Idaho National Laboratory.[122] The next several citations are drawn from this work. The most common form of EGS is for pairs of wells to be drilled into hot, dry rock, with hydraulic pressure used to fracture the rock at depth around and between the two wells. Water is then injected into one of the wells and hot water is produced out of the other, greatly improving the heat transfer from the rock over simple radial conduction into an embedded pipe. Because the total surface area of all the fractures in the rock can be very great, it is possible to extract heat at a much higher rate, eliminating the bottleneck of heat flow converging on a single pipe. Water from the production wells is recycled into the injection wells, with only a small need for make-up water. The overall environmental footprint of EGS is small.[123] Of course the intentional hydraulic fracturing of rock near the magma chamber of a supervolcano is a risk factor that must be carefully evaluated.

The following picture emerges for the projected "mature" EGS technology: each production well should be capable of delivering 80 kg/s [124] of near-critical-point water at 340C,[125] from depths as great at 10 km.[126] Electric power can be generated from this resource at a cost of approx. $1/W installed capacity (including the cost of the wells, generator, heat exchangers, and all other hardware), generating 270 MW (electric) from each 1000 kg/s of water produced at 340C.[127] The change in enthalpy of water between 100C and 340C is 1.56 MJ/kg.[128] So the input thermal power for each 1000 kg/s of water flow at 340C, with the heat rejected at 100C, is 1.56 GW (thermal), and the overall thermal efficiency is 17.3%. Thus we can anticipate that an Enhanced Geothermal System extracting 20GW (thermal) from the perimeter of a supervolcano could generate 3.46 GW of electrical power for a cost of $3.46 billion, delivering power at under (perhaps well under) $0.10/kWh, even if the entire length of the holes must be drilled through rock as hard as granite.[129]

This is a competitive price for electric power - an unexpected bonus since our main purpose is to prevent a supervolcano catastrophe. However, this bonus may be the critical factor in getting decision maker approval for such an approach.

A total of 160 production wells would be required to extract the 20 GW in the form of 12,800 kg/s of water flow at 80 kg/s per well. The 20 GW of heat extracted would drain the approx. 10^{22} J of the "next" eruption in approx. 16,000 years, a very short time compared to the typical time between eruptions. The 20 GW extracted by these wells is large compared to the previously mentioned rates of existing heat flow to the surface and is comparable to the energy delivered from basaltic intrusions, estimated at 22 GW, as previously noted.[130]

So the strategy that emerges here would be to encircle the magma chamber with the needed 160 production and 160 injection wells and operate them for their useful life (say 50 years). This would cool a ring of rock around the magma chamber. Then the system would be replaced with a new ring, inside the previous ring (perhaps accomplished with directional drilling so the surface manifestation stays in place for several cycles). The original 20 GW extraction for 50 years would remove approx. $3x10^{19}$ J, the equivalent of $5x10^9$ cubic meters of rock cooled by 100C. This would cool out to a radius of over 50 m around each well pair, over a vertical span of 3000 m. So the next set of wells would be drilled 50 m in the direction of the center of the magma chamber. In this way, the cooling perimeter would close-in on the magma chamber at the rate of approx. 1 m/year. Even for a massive supervolcano such as Yellowstone, it would take less than 50,000 years for such a cooling system to completely drain the heat away from the magma chamber, all the while generating electricity at competitive prices. It is also straightforward to imagine that the entire capacity could be made several times the nominal 20 GW considered here, reducing proportionately the time to drain the heat from the magma chamber and further increasing the resilience of the system to sudden pulses of heat from below.

Since the circumference of Yellowstone is some 250 km, the initial 160 well-pairs around Yellowstone would be approx. 1.5 km apart. But each new ring of wells only cools to a radius of some 50 m around each well pair. So as new wells are drilled along lines toward the center of the magma chamber, they will begin to partition the magma chamber, with the span between the radial lines of wells largely uncooled for almost a millennium. The resulting radial "walls" of cooled magma will divide the magma chamber in such a way that it may be less likely for volatiles driven out of solution from any one partition to trigger a catastrophic eruption involving all the partitioned chambers.

Between the older, farthest facilities from the supervolcano and newer, closer facilities, a progressively steepening heat flow gradient exists. To mitigate the steepening gradient and transport heat not only vertically away but also laterally, we developed the concept of reinjection fluid swapping between hotter, inward facilities and colder, outward facilities. While some warming of the older outward facilities would be expected, this would pose little risk and could more easily be adjusted by mixing with fresh meteoric water. The even stronger benefit lies in the reinjection of even colder (outward derived production fluids) into the inward systems. In addition to the vertical heat exchange, we therefore facilitate heat loss by essentially creating an additional, horizontal heat exchange loop.

5.3 WATER RESOURCE MANAGEMENT

The above solutions would be massive scale endeavors that require major long term infrastructural investments. Yellowstone is very rich in water (like other “wet” volcanic systems) and water convection is believed to be the dominant upward heat transport process. It is a much greater challenge attempting to enhance dry conduction, rather than an engineered solution that enhances convection in an already water-rich system. If one does succeed in removing heat from the system, then most likely this will be competing first with the natural convection and doing relatively little to enhance cooling rates for the first few gigawatts. Conversely, a net removal of available water from a supervolcano hydrothermal system bears the risk of “drying out” of the hydrothermal convection heat removal mechanism, and should be avoided. A resulting dominance of conductive heat loss would most likely result in adverse effects (e.g., pervasive fracturing, magmatic gas ascent, etc.).

The hydrothermal system is described in more detail in Heasler et al. (2009).[131] Note that the depth of the water brine system (the deeper parts of the hydrothermal system is about 2.4 to 4.8 km below the surface. This is actually below the solid volcanic rock and sediment that extends to a depth of 900 to 1,800 m and is inside the hot but mostly solid part of the pluton that contains Yellowstone’s magma chamber,[132] where temperatures exceed 475 K. This is comparable with some of the depths of drilling discussed previously, and suggests that drilling below the existing hydrothermal systems might be extremely challenging.

Thus, an alternative mitigation strategy for supervolcanic eruptions might be simply to manage the water resources around the area. The current drought in California, as well as changing lake and water table levels, highlights how variable and delicate water resources can be. Historical records[133] suggest that both anthropogenic and natural climate variations can have profound effects on hydrologic systems, and furthermore, that growth of human activities, and attempts to impact changing water resources in order to sustain supply to urban, agricultural and industrial areas of demand, can enhance this greatly on a local level. It is, therefore, not unreasonable to think that the water supply around supervolcanoes may potentially vary greatly.

Some of the engineering challenges not discussed here, that will need to be addressed include:

- Can injecting water into a fault laden area lubricate existing faults in a way that makes eruptions more likely? Water has been known to lubricate existing faults (e.g., at Yucca Mountain).
- Can engineering materials be developed that survived the hydrothermal environment above a magma chamber (e.g., Yellowstone).

Given the engineering challenges, and the potential complications that relate to water supply, we consider a solution of simple water resource management to be a potentially useful alternative to an engineered solution. The most basic approach is simply to maintain the current system, preventing a build-up of heat due to suppression of advection and mechanical changes due to removal of water (and mass). However, if the “leaky bathtub” model is correct, then this simply provides a stable environment for filling the bathtub to the catastrophic level. Alternative approaches may be to deliberately increase or decrease water levels, either to enhance advection to remove heat as necessary, or to reduce it and potentially trigger eruptions under more controlled circumstances.

6. DISCUSSION OF POTENTIAL COMPLICATIONS TO ANY ENGINEERED SOLUTION

A key element of this model is the assumption that, if we were to instantaneously "freeze" the magma chamber by reducing its temperature and that of the surrounding rock to match the average thermal gradient around the planet, then no eruption would occur. While this may be plausible, there is a concern that such instantaneous freezing is not possible, and in any realistic system there is a possibility that, as we artificially extract heat energy out of the magma chamber, we could cause phase changes (e.g. volatiles coming out of solution) that would reduce the overall density causing expansion and cracking in the overburden, possibly opening a channel to the surface and precipitating an eruption.

One of the principles we would follow is "do no harm", so that the probability of inducing a violent eruption is as close to zero as possible. On the other hand, it is conceivable that smaller scale adverse effects like hydrothermal explosions and ground motion, as opposed to a global catastrophe, would be acceptable if we could somehow be confident in one over the other. Beyond human intervention, huge pulses of heat energy into the magma chamber may at times precipitate eruptions, with brief periods where the heat flux is so large that engineering solutions would be impractical. If heat flow were sufficiently massive then it may be impossible to mitigate supervolcanic eruptions.

6.1 CAN COOLING MAGMA TRIGGER ERUPTIONS BY VOLATILES BEING DRIVEN OUT OF SOLUTION? CAN WE ARRANGE IT SO THAT ANY RESULTING ERUPTION WILL NOT BE "SUPERVOLCANIC"?

Like with heat transfer and dissipation, the removal of volatiles which drive eruptions is desirably occurring at a controlled rate. The exsolution of volatiles in cooling magma is typically a gradual process which is triggered dominantly by either decompression, or addition of new volatiles from a deeper, volatile-rich injection. However, supersaturation can result in sudden violent exsolution events under certain circumstances (Mangan 2000).[134] For magma-volatile combinations in which solubility is less at lower temperatures, this could be problematic.

Also, as the case of Pinatubo's excess sulfur stored in anhydrite has shown, this is not a simple relationship but thermal triggering might play a role in exceptional circumstances as well. Care has to be taken to continuously instrumentally monitor the gas flux out of the geothermal mitigation system as well as out of the volcanic system as a whole. Changes in gas composition are usually the first indicator that a deep system is changing.[135]

6.3 CAN COOLING OF THE ROCK ABOVE A MAGMA CHAMBER MAKE IT MORE BRITTLE OR OTHERWISE SUSCEPTIBLE TO LOSS-OF-INTEGRITY?

Work by Gregg et al., 2012[136] uses a viscoelastic numerical model that assesses the fracture point of rocks in a temperature dependent way. While this topic needs further investigation by the community, their initial results find that the trigger mechanism of large caldera eruptions is highly dependent on the brittle-to-ductile transition and the temperature of the country rock. In essence, their model shows a "ductile halo" around a magma reservoir that helps reduce the likelihood of an eruption that is triggered from a dike propagating to the surface, or a fault propagating into the magma reservoir. An eruption with a brittle-to-ductile transition zone several kilometers below the surface would have to be initiated from the magma chamber through gas exsolution or a raising of this transition zone by roof uplift to the point of failure. In light of this, the other end of the spectrum in their model trend indicates cooling the country rock would allow fractures to propagate from the roof (ground surface) down to the magma reservoir more readily, potentially creating a pathway for eruption initiation that otherwise would have not occurred.

7. CONCLUSIONS

There are a number of unknowns about the nature of supervolcanic eruptions and how they are supplied that need to be addressed before attempting any engineered solutions, or favoring a water resource approach. In particular, more detailed "imaging" of the supply of melt to potentially supervolcanic systems is necessary in order to gain a better understanding of the how much heat supply varies from a steady state, experiments and models of how large magma reservoirs respond to cooling, and developing an improved understanding of the risks associated with the presence (and variable supply) of water above magma chambers. Given the potentially huge cost of supervolcanic eruptions on regional or even global scales, we consider these to be potentially valuable to stimulate future research.

For any of these techniques to be applied, government intervention would almost certainly be required. Some supervolcanoes are in government managed wilderness, as is the case for Yellowstone National Park. Even the concept of active water resource management would require planning, legislation to prevent private activities from adversely affecting water resource, and potentially infrastructure to transport water to or from the site. For an engineered solution, industry would likely require special encouragement to drill to the required depths and temperatures even when strict return-on-investment considerations may favor shallower wells and lower produced-water temperatures. Also, the "mature" EGS technology envisioned in the MIT report would have to become available, which might require government subsidy. But over all, it seems quite plausible that the entire enterprise could be self-supporting following initial investment by sale of the resulting electricity into the national power grid.

While Yellowstone is the greatest supervolcano threat to the U.S. (and perhaps also to all of human civilization), there are two other supervolcanoes within the contiguous 48 states. These are the Long Valley Caldera near Mammoth Lakes CA, and the Valles Caldera near Los Alamos, NM. Since Long Valley is located in a popular recreation and vacation-housing area without national park status, it seems to be a likely candidate for an initial test of the system. Indeed there is already a geothermal power plant at Long Valley, the Casa Diablo plant operated by Ormat (Ormat Technologies, Inc., Reno NV). Being in population-rich California, it is also closer to a large customer base for increased electricity production capacity.

ACKNOLWLEDGEMENTS:

In addition to the co-authors of this publication, the PI for the task would like to thank David Pieri for helpful discussions and pointers into the literature early in the formulation of this effort. U.S. Government sponsorship acknowledged.

REFERENCES

[1] See, for example, 4th IAA Planetary Defense Conference; http://www.pdc2015.org/.

[2] Rampino, M.R. "Super-eruptions as a Threat to Civilizations on Earth-like Planets" Icarus 156, 562–569 (2002) doi:10.1006/icar.2001.6808, available at http://www.idealibrary.com.

[3] National Research Council, Committee to Review Near-Earth Object Surveys and Hazard Mitigation Strategies, Space Studies Board, "Defending Planet Earth: Near-Earth Object Surveys and Hazard Mitigation Strategies: Final Report", 2010, http://www.nap.edu/catalog/12842/defending-planet-earth-near-earth-object-surveys-and-hazard-mitigation, Figure 2.4 shows that 2 km diameter corresponds approximately to absolute magnitude 16.5.

[4] Harris, Alan and D'Abramo, Germano "Population of Near-Earth Asteroids" ICARUS 257 (2015) 302-312, Table 1, average of column R(H) to magnitude 16.5 weighted by column N(2014), http://www.sciencedirect.com/science/ article/pii/S0019103515002067.

[5] NRC, 2010, Figure 2.5 shows onset of global effects between 1 and 2 km diameter.

[6] Rampino, M.R., "Volcanic Winters", Ann. Rev. Earth Planet. Sci. 1988. 16: 73-99.

[7] Rampino, 2002.

[8] Study to Determine the Feasibility of Extending the Search for Near- Earth Objects to Smaller Limiting Diameters, National Aeronautics and Space Administration, Report of the Near-Earth Object Science Definition Team, page 16, http://neo.jpl.nasa.gov/neo/neoreport030825.pdf

[9] Harris, 2015, Table 1, ratio of cumulative nD to n(<H) up to H=16.5.

[10] http://neo.jpl.nasa.gov/neo/report.html

[11] U.S.Congress (19 March 2013). "Threats From Space: a Review of U.S. Government Efforts to Track and mitigate Asteroids and Meteors (Part I and Part II) - Hearing Before the Committee on Science, Space, and Technology House of Representatives One Hundred Thirteenth Congress First Session" (PDF). *United States Congress*. p. 147. Retrieved 3 May 2014.

[12] Foulger, G.R., Plates vs Plumes: A Geological Controversy, 2010, Wiley-Blackwell, 364 pages ISBN: 978-1-4051-6148-0

[13] Sigurdsson H., Origin and Transport of Magma, in: Eds. Sigurdsson, Houghton, McNutt, Rymer, Stix, 2015, The Encyclopedia of Volcanoes, Second Edition, pg 33 or 1421.

[14] Miller, C.F., and Wark, D.A., 2008, Super-volcanoes and their explosive super-eruptions, Elements, 4, 11-15.

[15] Newhall, C. G., & Self, S. (1982). The volcanic explosivity index (VEI): An estimate of explosive magnitude for historical volcanism. *Journal of Geophysical Research*, *87*(C2), 1231-1238.

[16] Self, S. 2006, The effects and consequences of very large explosive volcanic eruptions, Phhilosophical Transactions of the Royal Society, 364, 2073-97.

[17] Sparks,R.S.J., Self, S., Grattan J.P., et al., 2005, Super-eruptions: Global effects and future threats, Geological Society of London, 28.

[18] Thordarson, T., & Self, S. (2003). Atmospheric and environmental effects of the 1783–1784 Laki eruption: a review and reassessment. *Journal of Geophysical Research: Atmospheres (1984–2012)*, *108*(D1), AAC-7.
[19] Witham, C. S., & Oppenheimer, C. (2004). Mortality in England during the 1783–4 Laki Craters eruption. *Bulletin of Volcanology*, *67*(1), 15-26.
[20] Oppenheimer, C. (2003). Climatic, environmental and human consequences of the largest known historic eruption: Tambora volcano (Indonesia) 1815. *Progress in physical geography*, *27*(2), 230-259.
[21] The Economist (2015) *Volcanoes and climate - After Tambora*. The Economist, 11 April 2015. http://www.economist.com/news/briefing/21647958-two-hundred-years-ago-most-powerful-eruption-modern-history-made-itself-felt-around .
[22] Simkin, T. & R.S. Fiske (1983), *Krakatau 1883. The volcanic eruption and its effects*. Smithsonian Institution Press, Washington DC (2nd ed.), 464pp. ISBN 0-87474-841-0.
[23] Dörries, M. (2003). Global science: the eruption of Krakatau. *Endeavour*, *27*(3), 113-116.
[24] Oman, L., Robock, A., Stenchikov, G., Schmidt, G. A., & Ruedy, R. (2005). Climatic response to high-latitude volcanic eruptions. *Journal of Geophysical Research: Atmospheres (1984–2012)*, *110*(D13).
[25] Adleman, Jennifer. "The great eruption of 1912." *Alaska Park Science* (2002): 5.
[26] Hansen, J., Lacis, A., Ruedy, R., & Sato, M. (1992). Potential climate impact of Mount Pinatubo eruption. *Geophysical Research Letters*, *19*(2), 215-218.
[27] Gathorne-Hardy, F. J., & Harcourt-Smith, W. E. H. (2003). The super-eruption of Toba, did it cause a human bottleneck?. *Journal of Human Evolution*, *45*(3), 227-230.
[28] Handler, P. (1989). The effect of volcanic aerosols on global climate. *Journal of Volcanology and Geothermal Research*, *37*(3), 233-249.
[29] Self, S. (2006). The effects and consequences of very large explosive volcanic eruptions. *Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, *364*(1845), 2073-2097.
[30] USGS Fact Sheet 027-00 , 2000, Volcanic Ash Fall—A "Hard Rain" of Abrasive Particles
[31] Finnsson, H., 1796, Um mannfækkun ı´ Hallærum (Decimation of the population in Iceland due to famines), in Rit Thess konunglega ı´slenska lærdo´mslistafe´lags, XIV, pp. 30– 226, Copenhagen
[32] von Buch, L. (1820), Ueber die Zusammensetzung der basaltischen Inseln und ueber Erhebungs-Cratere. Kgl. Pruss. Akad. D. Wiss., Berlin 1820, 18p.
[33] Scott, W.E., Hoblitt, R.P., Torres, R.C., Self, S., Martinez, M.M.L., Nillos Jr., T. 1996. Pyroclastic Flows of the June 15, 1991, Climactic Eruption of Mount Pinatubo. *In*: C.G. Newhall and R.S. Punongbayan (eds) *Fire and Mud – Eruptions and Lahars of Mount Pinatubo, Philippines.* Philippine Institute of Volcanology and Seismology and University of Washington Press, 1996.
[34] Howard, K.A., 2010, Caldera collapse: Perspectives from comparing Galápagos volcanoes, nuclear-test sinks, sandbox models, and volcanoes on Mars, GSA Today 20(10): 4-10. DOI: 10.1130/GSATG82A.1
[35] Anderson, E.M., 1937, Cone sheets and ring dikes: the dynamical explanation. Bulletin Volcanologique, 2, 1, 35-40.
[36] Steven, T.A. and Lipman, P.W., 1976, Calderas of the San Juan volcnic field, southwestern Colorado, USGS prof paper 958, 35
[37] Scott, 1996.
[38] Iyer H., Evans J., Dawson P, Stauber D., Achauer U., 1990. Differences in magma storage in different volcanic envrionments as revealed by seismic tomography; silicic volcanic centers and subduction related volcanoes. Magma Transport and Storage pp 293-316.
[39] Chu R., Helmberger D.V, Sun D., Jackson J.M, Zhu L., 2010. Mushy magma beneath Yellowstone. Geophys Res. Lett. 37, L01306.
[40] Malfait, W.J., Seifert, R., Petitgirard, S., Perrillat, J.-P., Mezouar, M., Ota, T., Nakamura, E., Lerch, P., Sanchez-Valle, C., 2014a. Supervolcano eruptions driven by melt buoyancy in large silicic magma chambers. Nature Geoscience 7, 122e125.
[41] Koyanagi et al from Hawaii -> plexus of dikes an sills paper
[42] Gualda G.A.R, Ghiorso M.S., 2013 The bishop tuff giant magma body: an alternative to the standard model. Contrib Mineral. Pertol. 166, 755-775.

[43] Huang, H.H., Lin, F.C., Schmandt, B., Farrell, J., Smith, R.B., Tsai, V.C., 2015, The Yellowstone magmatic system from the mantle plume to the upper crust, Science, 348, 6236, 773-776.

[44] Lowenstern J.B., Smith R. B. Hill D. P, Monitoring super-volcanoes: geophysical and geochemical signals at Yellowstone and other large caldera systems. Philosophical transactions of the Royal Society A364:2005-2072. 2006

[45] Farrell, J., Smith, R.B., Husen, S>, Diehl, T., 2014, Tomography from 26 yrs of seismicity revealing that the spatial extent of the Yellowstone crustal magma reservoir extends well beyond the Yellowstone caldera, Geophy. Res. Lett., 41, doi:10.1002/2014GL059588

[46] Huang, 2015.

[47] Fournier, R. O. (1989). Geochemistry and Dynamics of the Yellowstone National Park Hydrothermal System. *Annual Review of Earth and Planetary Sciences*, 17(1), 13-53.

[48] Werner, C., & Brantley, S. (2003). CO2 emissions from the Yellowstone volcanic system. *Geochemistry, Geophysics, Geosystems*, 4, 1061.

[49] Jacob B. Lowenstern and Shaul Hurwitz "Monitoring a volcanoesuper volcano in Repose: Heat and Volatile Flux at the Yellowstone Caldera", DOI: 10.2113/GSELEMENTS.4.1.35, http://pubs.er.usgs.gov/publication/70033744, February 2008, p36

[50] See e.g. "Ellis B., Branney M.J. (2010) Silicic explosive phreatomagmatism from the Yellowstone hotspot track: the Deadeye Member. Bulletin of Volcanology 72, 1241–1257 DOI 10.1007/s00445-010-0400-9" for an example of this at Yellowstone/SNP.

[51] See, for example, Morgan, W. J. (1972). "Deep mantle convection plumes and plate motions". *Bull. Am. Assoc. Pet. Geol.* **56**: 203–213

[52] Robert B. Symonds, William I. Rose, Gregg J. S. Bluth, Terence M. Gerlach. Volcanic-gas studies: Methods, results, and applications, Rev. Mineral. Geochem 30, 1-66 (1994).

[53] Eggers Jr, D. F., & Crawford Jr, B. L. (1951). Vibrational intensities. III. Carbon dioxide and nitrous oxide. The Journal of Chemical Physics, 19(12), 1554-1561.

[54] Schwandner F. M., Seward T. M., Gize A. P., Hall P. A., and Dietrich V. J. (2004): Diffuse emission of organic trace gases from the flank and crater of a quiescent active volcano (Vulcano, Aeolian Islands, Italy). Journal of Geophysical Research D - Atmospheres 109, D04301, 20pp.DOI:10.1029/2003JD003890.

[55] Chiodini, G., Granieri, D., Avino, R., Caliro, S., Costa, A., & Werner, C. (2005). Carbon dioxide diffuse degassing and estimation of heat release from volcanic and hydrothermal systems. Journal of Geophysical Research: Solid Earth (1978–2012), 110(B8).

[56] Jacob B. Lowenstern* and Shaul Hurwitz* "Monitoring a volcanoesuper volcano in Repose: Heat and Volatile Flux at the Yellowstone Caldera", DOI: 10.2113/GSELEMENTS.4.1.35, http://pubs.er.usgs.gov/publication/70033744, February 2008, p36

[57] Eichelberger, J. C. (1980). Vesiculation of mafic magma during replenishment of silicic magma reservoirs. Nature 288: 446-450.DOI 10.1038/288446a0.

[58] Murphy, M. D., Sparks, R. S. J., Barclay, J., Carroll, M. R., & Brewer, T. S. (2000). Remobilization of andesite magma by intrusion of mafic magma at the Soufriere Hills Volcano, Montserrat, West Indies. Journal of petrology, 41(1), 21-42.

[59] Delaney, P. T., Sass, J. H., Duffield, W. A., and Kauahikaua, J. P., 1995, Temperature and response time of magma hydrothermal systems, with an example from Kilauea, Hawaii: Proceedings, World Geothermal Congress, Florence, Italy, 18-31 May, 1995, v. 2., p. 1239-1244.

[60] Lowell, R. P., Farough, A., Hoover, J., & Cummings, K. (2013). Characteristics of magma-driven hydrothermal systems at oceanic spreading centers. *Geochemistry, Geophysics, Geosystems*, *14*(6), 1756-1770.

[61] Symonds, R. B., Gerlach, T. M., & Reed, M. H. (2001). Magmatic gas scrubbing: implications for volcano monitoring. *Journal of Volcanology and Geothermal Research*, *108*(1), 303-341.

[62] Nakano, M., & Kumagai, H. (2005). Response of a hydrothermal system to magmatic heat inferred from temporal variations in the complex frequencies of long-period events at Kusatsu-Shirane Volcano, Japan. *Journal of volcanology and geothermal research*, *147*(3), 233-244.

[63] Hill, D. P., Pollitz, F., & Newhall, C. (2002). Earthquake-volcano interactions. *Physics Today*, *55*(11), 41-47.

[64] Pallister, J.S., Hoblitt, R.P., Meeker, G.P., Knight, R.J., and D.F. Siems (1996). Magma mixing at Mount Pinatubo: Petrographic and chemical evidence from the 1991 deposits. In: C.G. Newhall & R.S. Punongbayan (eds) Fire and Mud – Eruptions and Lahars of Mount Pinatubo, Philippines. Philippine Institute of Volcanology and Seismology and University of Washington Press, p. 687–731.

[65] Delfin, F.G., Villarosa, H.G., Layugan, D.B., Clemente, V.C., Candelaria, M.R., & Ruaya, J.R. (1996). Geothermal Exploration of the pre-1991 Mount Pinatubo Hydrothermal System. In: C.G. Newhall & R.S. Punongbayan (eds) Fire and Mud – Eruptions and Lahars of Mount Pinatubo, Philippines. Philippine Institute of Volcanology and Seismology and University of Washington Press, 1996, p. 197-212.

[66] Bautista, B. C., L. P. Bautista, E. S. Barcelona, R. S. Punongbayan, E. P. Laguerta, A. R. Rasdas, G. Ambubuyog, E. Q. Amin, and R. S. Stein, Relationship of regional and local structures to Mount Pinatubo activity, in Eruptions and Lahars of Mount Pinatubo, Philippines, edited by C. G. Newhall, and R. S. Punongbayan, pp. 351–370, PHIVOLCS, Quezon City, and Univ. of Wash. Press, Seattle, Wash., 1996, pp. 351-370.

[67] Barnes, H.L., 1997, Geochemistry of Hydrothermal Ore Deposits. Wiley, 972p. ISBN 047157144X

[68] Seward, T.M. (1993), The hydrothermal geochemistry of gold. In: Foster, R.P. (Ed.), *Gold Metallogeny and Exploration*, Springer, 1993, 37-62. ISBN: 978-0-412-56960-9

[69] Stefánsson, A., & Seward, T. M. (2004). Gold (I) complexing in aqueous sulphide solutions to 500 C at 500 bar. *Geochimica et Cosmochimica Acta*, *68*(20), 4121-4143.

[70] Symonds et al, 1994

[71] Symonds et al 2001

[72] Lowell, J. D., & Guilbert, J. M. (1970). Lateral and vertical alteration-mineralization zoning in porphyry ore deposits. Economic Geology, 65(4), 373-408.

[73] Fournier 1989

[74] Jacob B. Lowenstern* and Shaul Hurwitz* "Monitoring a volcanoesuper volcano in Repose: Heat and Volatile Flux at the Yellowstone Caldera", DOI: 10.2113/GSELEMENTS.4.1.35, http://pubs.er.usgs.gov/publication/70033744, February 2008, p36

[75] Fournier 1989.

[76] Rye, R. O., & Truesdell, A. H. (2007). The Question of Recharge to the Deep Thermal Reservoir Underlying the Geysers and Hot Springs of Yellowstone National Park. *US Geological Survey professional paper* (1717), 235-270.

[77] Lowenstern, 2008, p36.

[78] http://www.engineeringtoolbox.com/water-thermal-properties-d_162.html

[79] Jacob B. Lowenstern* and Shaul Hurwitz* "Monitoring a volcanoesuper volcano in Repose: Heat and Volatile Flux at the Yellowstone Caldera", DOI: 10.2113/GSELEMENTS.4.1.35, http://pubs.er.usgs.gov/publication/70033744, February 2008, p36

[80] Branney, M. and Acocella, Valerio (2015). Calderas. In: Sigurdsson, H., Houghton, B., McNutt, S., Rymer, H., and Styx, J. (Eds.), "The Encyclopedia of Volcanoes". 2015. pp. 299-315.

[81] Pollard, D. D., & Aydin, A. (1988). Progress in understanding jointing over the past century. *Geological Society of America Bulletin*, *100*(8), 1181-1204.

[82] Holland, M., Urai, J. L., & Martel, S. (2006). The internal structure of fault zones in basaltic sequences. *Earth and Planetary Science Letters*, *248*(1), 301-315.

[83] Ryan, M. P., & Sammis, C. G. (1978). Cyclic fracture mechanisms in cooling basalt. *Geological Society of America Bulletin*, *89*(9), 1295-1308.

[84] Pollard, D. D., Segall, P. A. U. L., & Delaney, P. T. (1982). Formation and interpretation of dilatant echelon cracks. *Geological Society of America Bulletin*, *93*(12), 1291-1303.

[85] Gudmundsson, A. (2002). Emplacement and arrest of sheets and dykes in central volcanoes. *Journal of Volcanology and Geothermal Research*, *116*(3), 279-298.

[86] Gudmundsson, A. (2007). Conceptual and numerical models of ring-fault formation. *Journal of Volcanology and Geothermal Research*, *164*(3), 142-160.

[87] Roche, O., Druitt, T. H., & Merle, O. (2000). Experimental study of caldera formation. *Journal of Geophysical Research: Solid Earth (1978–2012)*, *105*(B1), 395-416.

[88] Cashman, K. V., & Giordano, G. (2014). Calderas and magma reservoirs. *Journal of Volcanology and Geothermal Research*, *288*, 28-45.

[89] Delaney, P. T., Pollard, D. D., Ziony, J. I., & McKee, E. H. (1986). Field relations between dikes and joints: emplacement processes and paleostress analysis. *Journal of Geophysical Research: Solid Earth (1978–2012)*, *91*(B5), 4920-4938.
[90] Valentine, G. A., & Krogh, K. E. (2006). Emplacement of shallow dikes and sills beneath a small basaltic volcanic center–The role of pre-existing structure (Paiute Ridge, southern Nevada, USA). *Earth and Planetary Science Letters*, *246*(3), 217-230.
[91] Gregg, P. M., De Silva, S. L., Grosfils, E. B., & Parmigiani, J. P. (2012). Catastrophic caldera-forming eruptions: Thermomechanics and implications for eruption triggering and maximum caldera dimensions on Earth. Journal of Volcanology and Geothermal Research, 241, 1-12.
[92] Gregg et al. 2012.
[93] Gregg et al. 2012.
[94] Papale, P. (1999). Strain-induced magma fragmentation in explosive eruptions. *Nature*, *397*(6718), 425-428.
[95] Kaminski, É., & Jaupart, C. (1997). Expansion and quenching of vesicular magma fragments in Plinian eruptions. *Journal of Geophysical Research: Solid Earth (1978–2012)*, *102*(B6), 12187-12203.
[96] Robock, Alan (2000), Volcanic eruptions and climate. Reviews of Geophysics 38 (2): 191-219. doi:10.1029/1998RG000054.
[97] http://www.theguardian.com/global-development/2012/oct/14/un-global-food-crisis-warning
[98] Rampino (1988).
[99] Lowenstern and Hurwitz (2008)
[100] Neuville, Daniel R. et al, "Thermodynamic and rheological properties of rhyolite and andesite melts", Contributions to Mineralogy and Petrology. December 1993, Volume 113, Issue 4, pp 572-581, http://link.springer.com/article/ 10.1007%2FBF00698324
[101] Neuville, D.R., Courtial, P., Dingwell, D.B., Richet, P., "Thermodynamic and rheological properties of rhyolite and andesite melts" Contributions to Mineralogy and Petrology, 1993, Volume 113, Issue 4, pp 572-581, http://link.springer.com/article/10.1007%2FBF00698324#page-1
[102] Lowenstern, Smith, Hill 2006.
[103] Lichti, Keith A., Susan J. Swann, Stephen P. White, Norio Sanada, Y. Kurata, Hiroshi Nanjo, Jun Ikeuchi, and B. Christenson. "Corrosion in Volcanic gases." In *Proc. NEDO International Geothermal Symposium*. 1997.
[104] Lichti et al (1997).
[105] Benning, L. G., & Seward, T. M. (1996). Hydrosulphide complexing of Au (I) in hydrothermal solutions from 150–400 C and 500–1500 bar. *Geochimica et Cosmochimica Acta*, *60*(11), 1849-1871.
[106] Simmons, S. F., Brown, K. L., Browne, P. R., & Rowland, J. V. (2015). Gold and silver resources in Taupo Volcanic Zone geothermal systems. *Geothermics*.
[107] Mundhenk, N., Huttenloch, P., Sanjuan, B., Kohl, T., Steger, H., & Zorn, R. (2013). Corrosion and scaling as interrelated phenomena in an operating geothermal power plant. *Corrosion Science*, *70*, 17-28.
[108] Keserovic, A., & Bäßler, R. (2015). Suitability of UNS S31603 steel for geothermal brines in volcanic areas–Influence of different physicochemical conditions on its corrosion behavior. *Geothermics*, *53*, 479-487.
[109] Scherer, G.W., et al., 2005, Leakage of CO2 through abandoned wells: role of corrosion of cement, In: Thomas DC & Benson SM (Eds.), Carbon Dioxide Capture for Storage in Deep Geological Formations, Elsevier, pp. 827-848.
[110] Keserovic et al 2015
[111] Nogara, J.B., Zarrouk, S.J., 2014, Evaluation of Corrosion Resistant Alloys as Construction Material for Acidic Geothermal Wells. Proceedings 36th New Zealand Geothermal Workshop 24 - 26 November 2014 Auckland , New Zealand.
[112] e.g., Corsi, R. 1986, Scaling and Corrosion in Geothermal Equipment: Problems and preventative measures. Geothermics 15: 839-856.
[113] Brown, K. 2013, Mineral Scaling in Geothermal Power Production. United Nations University, Geothermal Training Programme Report 39, Reykjavik (Iceland). ISBN 978-9979-68-337-7.
[114] Nitschke, F., Scheiber, J., Kramar, U., & Neumann, T. (2014). Formation of alternating layered Ba-Sr-sulfate and Pb-sulfide scaling in the geothermal plant of Soultz-sous-Forêts. *Neues Jahrbuch für Mineralogie-Abhandlungen: Journal of Mineralogy and Geochemistry*, *191*(2), 145-156.

[115] Hreggvidsson, G. O., Petursdottir, S. K., Björnsdottir, S. H., & Fridjonsson, O. H. (2012). *Microbial speciation in the geothermal ecosystem* (pp. 37-67). Springer Vienna.

[116] Windman, T, Zolotova, N., Schwandner, F., and Shock, E. (2007): Formate as an Energy Source for Microbial Metabolism in Chemosynthetic Zones of Hydrothermal Ecosystems. *Astrobiology* 7(6):873-890.

[117] e.g. Stober, I., Bucher, K. 2013, The Chemical Composition of Deep Geothermal Waters and Its Consequences for Planning and Operating a Geothermal Power Plant. In: Stober & Bucher (Eds.), Geothermal Energy, Springer, New York, pp. 255-277. ISBN: 978-3-642-13351-0.

[118] e.g. Baba, A., Demir, M. M., Koç, G. A., & Tuğcu, C. (2015). Hydrogeological properties of hyper-saline geothermal brine and application of inhibiting siliceous scale via pH modification. *Geothermics*, *53*, 406-412.

[119] Wahrenberger, C., Seward, T. M., & Dietrich, V. (2002). Volatile trace element transport in high temperature gases from Kudriavy volcano (Iturup, Kurile Islands, Russia). Water–Rock Interaction: a Tribute to David A. Crerar. Geochem. Soc. Spec. Publ, 7, 307-327.

[120] Schwandner, 2004

[121] See also patent US 7284931 B2, Stratford, B.S. (2002), *Magma evacuation systems for the prevention of explosions from supervolcanoes*. https://www.google.com/patents/US7284931

[122] Tester, Jefferson W.(Chair) et al, "The Future of Geothermal Energy: Impact of Enhanced Geothermal Systems on the United States in the 21st Century - An assessment by an MIT-led interdisciplinary panel", DOE Idaho contract DE-AC07-05ID14517, ISBN: 0-615-13438-6, 2006, http://geothermal.inel.gov/publications/future_of_geothermal_energy.pdf

[123] Tester, 2006, p 8-16 to 8-18.

[124] Tester, 2006, p 1-29

[125] Tester, 2006, p 1-26

[126] Tester, 2006, p 2-3.

[127] Tester, 2006, p 7-11 to 7-16.

[128] http://www.engineeringtoolbox.com/water-thermal-properties-d_162.html

[129] Tester, 2006, p 1-29.

[130] Assuming 300K increased temperature for the basalt over the silicic magma, heat capacity of 23 kJ/kg, and density of 2800 kg/m^3.

[131] Heasler, HP, Jaworowki C, Foley D (2009) Geothermal systems and monitoring hydrothermal features. In: Young, R., and Norby, L., eds., Geological Monitoring: Boulder, Colorado, Geological Society of America, p. 105–140. http://www.nature.nps.gov/ geology/monitoring/files/geomon-05.pdf).

[132] Smith, R. B. & Siegel, L. J. (2000). Windows into the Earth: The Geologic Story of Yellowstone and Grand Teton National Parks., 242 p.

[133] http://www.water.ca.gov/waterconditions/docs/California_Signficant_ Droughts_2015_small.pdf

[134] Mangan, M., Sisson, T., 2000. Delayed, disequilibrium degassing in rhyolite magma: Decompression experiments and implications for explosive volcanism. Earth Planet. Sci. Lett. 183: 441-455.

[135] Andrea Cannata, Gaetano Giudice, Sergio Gurrieri, Placido Montalto, Salvatore Alparone, Giuseppe Di Grazia, Rocco Favara, Stefano Gresta, Marco Liuzzo, Relationship between soil CO_2 flux and volcanic tremor at Mt. Etna: Implications for magma dynamics, Environ. Earth Sci. 61, 477-489 (2010). doi: 10.1007/s12665-009-0359-z.

[136] Gregg et al, 2012.