Chakoumakos and Lumpkin (1990) Pressure – temperature constraints on the crystallization of the Harding pegmatite, Taos County, New Mexico

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Introduction

In essay #10 of this series, I cited text from Swanson (1977) and Fenn (1977) that gave experimental corroboration to a long-held belief among geoscientists: that the slow cooling of silicate melt fosters the growth of large crystals, whereas rapid cooling promotes fine-grained crystal textures. Strictly speaking, the crystal sizes in the experiments of Swanson and of Fenn correlated with the magnitude of liquidus undercooling, not the rate of cooling, which was nearly instantaneous in both cases. Both authors made an intuitive connection between the crystal sizes and the environments of crystallization in which the rates of cooling are slow (plutonic) and fast (volcanic). With that correlation, the giant size of crystals in pegmatites could be inferred to have formed over very long time frames of very slow cooling (Figure 1). That inference has been stated as fact in some definitions of pegmatite (see Chapter 1 of London, 2008).

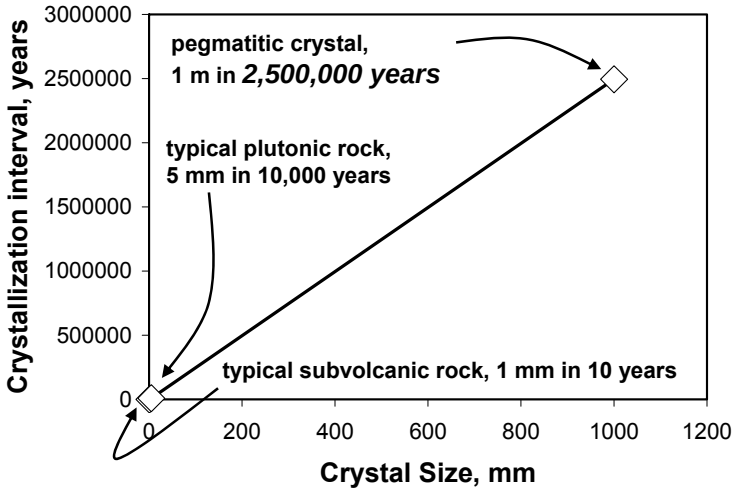

Figure 1. An inferred relationship between crystal size and the duration of crystallization in pegmatites, in relation to the same properties of volcanic and plutonic rocks. From London (2008) and sources therein.

Jahns (1953) may have been the first geologist to suggest that the giant crystals in pegmatites form rapidly, and hence that the melts from which they grow cool rapidly as well (see essay #5 of this series). He did not explain the evidence that might have led him to that conclusion, except that he recognized that pegmatite-forming melts that form dikes in other types of host rocks have migrated from their source, presumably upward, and down a geothermal gradient to lower temperatures. Jahns and Tuttle (1963, essay #8 of this series) remarked on "*chilled margins*", but they attributed those to what they called "*compositional quench*" rather than to the more conventional explanation via rapid cooling. In a recent paper, London et al. (2020a) documented the texture of granophyre – a microscopic graphic intergrowth of feldspar and quartz – in the border units of thin and thick pegmatite dikes. Heretofore, granophyre has been associated only with rapidly cooled volcanic rocks (e.g., Barker, 1970; Morgan and London, 2012).

The notion that pegmatites cool slowly as a means to explain their giant crystal dimensions changed with a publication by Bryan Chakoumakos and Gregory Lumpkin:

- Chakoumakos, B.C. and Lumpkin, G.R. (1990) Pressure-temperature constraints on the crystallization of the Harding pegmatite, Taos County, New Mexico. *Canadian Mineralogist*, **28**, 287-298.

That article and its consequences are the principal subjects of this essay.

Pressure-Temperature Constraints

Chakoumakos and Lumpkin (1990) chose the Harding pegmatite, near Dixon, Taos County, New Mexico as the case study for their model. The geology and mineralogy of the pegmatite had previously been described by Jahns and Ewing (1977), and the metamorphic grade of the rocks that host the pegmatite had been characterized by Holdaway (1977) and Grambling and Williams (1985). The latter two studies put the peak metamorphic conditions at 500°-550°C and 325-375 MPa (mega-pascals).

Figure 2 of Chakoumakos and Lumpkin (1990) conveyed some of the constraints in pressure and temperature that were incorporated into their cooling model. The intersections of the hydrous solidus and liquidus of the Harding pegmatite bulk composition (Jahns and Burnham, 1957) with the reaction of spodumene + 2 quartz = petalite from London (1984) gave a region of overlap in the range of 550°-675°C and 310-370 MPa, which is close to the peak metamorphic condition cited above. With those data, Chakoumakos and Lumpkin (1990) selected a pressure of 350 MPa as the regional pressure at the time of pegmatite emplacement, and 550°C as one of two temperatures for the host rocks to the pegmatite upon emplacement. Two inferences were implicit in these choices: (1) the equilibrium boundaries for the solidus and liquidus of the Harding bulk composition defined the temperature interval of crystallization, and (2) the pegmatite-forming melt intruded host rocks at the peak of their metamorphism. In the first case, Chakoumakos and Lumpkin (1990) took the equilibrium solidus to represent the temperature at which the melt would have crystallized completely. In the second case, the pegmatite-forming melt would have cooled to the temperature of the host rocks, 550°C, but not lower until the host rocks began their regional uplift and its consequent drop in pressure and temperature.

Fluid Inclusion Isochors

Chakoumakos and Lumpkin (1990) cited work by Cook (1979) to calculate isochors based on the homogenization of aqueous-carbonic fluid inclusions in quartz and beryl from three zonal assemblages in the pegmatite. In their Figure 5, they delineated a path of cooling that starts at ~ 650°C and culminates at 350°C for what Jahns and Ewing (1977) regarded as the core assemblage of cleavelandite (coarse platy albite) plus rose muscovite. The path they plotted falls between 250-300 MPa in pressure. Had they extrapolated those isochors to the pressure of 350 MPa, their inferred pressure at the point of pegmatite emplacement, the temperatures would have been substantially higher, especially at the high end (> 800°C).

Numerical Simulation of Cooling of the Harding Pegmatite

The most significant contribution of the article by Chakoumakos and Lumpkin (1990) was a numerical model for the cooling of the pegmatite body (their Figure 7). They used a solution to the heat conduction equation from Carslaw and Jaeger (1959), which models the conductive loss of heat from a hot body (magma) in cooler surroundings (host rocks). The model does not include the effects of convection within the melt body, nor the possible loss of heat from an escaping aqueous solution, and the authors deemed the contribution from the heat of crystallization of the melt to be negligible. As boundary conditions, they chose a temperature of 650°C for the pegmatite-forming melt upon emplacement (the liquidus temperature of the hydrous melt from Jahns and Burnham, 1957), and a host rock temperature of 350°C. That host rock temperature was determined by a thermal gradient of 30°C/km to the depth of 11-12 km at which the lithostatic pressure was ~ 350-400 MPa. That geothermal gradient is high in relation to other determinations of a continental geotherm of 18°C/km (e.g., Burke and Kidd, 1978; Catlos et al., 2001). The model considered that the emplacement of the melt from its source into the host rocks occurred in a single step, not incrementally.

Figure 7 presented several cooling curves at different locations with the pegmatite from its margins to center. The most significant of those curves was the cooling history at the center of the dike. The center of the dike would have cooled to 550°C, the equilibrium solidus of the melt, in ~ 3 years. The cooling curves are concave and parabolic in shape, and they converge after 100 years at a temperature of ~ 385°C. From this model, Chakoukmakos and Lumpkin (1990) concluded that:

"*For host-rock temperatures below 550°C, the magmatic crystallization (giant-crystal texture) would have occurred in 100 years or less. For a host-rock temperature near the pegmatite solidus (550°C), the magmatic portion of crystallization would have taken much longer (> 1000 years).*"(p. 296).

They phrased the cooling rate somewhat differently in the abstract:

"*The magmatic portion of crystallization began at 650°C and 330-350 MPa (11-12 km depth) and continued isobarically to 550°C… If the host-rock temperature at the time of intrusion was 0 to 200°C below the solidus temperature, then cooling models for a finite sheet indicate that the magmatic crystallization (giant-crystal texture) occurred in 1000 years or less.*"(p. 287)

Chakoumakos and Lumpkin (1990) did not discuss what their model of Figure 7 showed: that with host rocks at 350°C, the dike would have cooled to 550°C at its center in ~ 3 years. At that rate of cooling and a dike thickness of 20 meters, a crystallization front would have had to advance at a rate of ~ 2-3 meters per year. That translates to an advance of the crystallization front at a rate of 10^{-5} cm/second, or about an order of magnitude faster than the minimum crystal growth rates (10^{-6} cm/second) reported by Swanson (1977). Cooling to the solidus temperature over 100 years would drop the necessary crystal growth rate to ~ 10^{-7} cm/second, which is in line with Swanson's (1977) results. Moreover, the melt that formed the Harding pegmatite would have cooled continuously, which means that the driving force to form crystals would not diminish with time as it did in the isothermal experiments of Swanson (1977) and Fenn (1977).

Whether 3 years or 100 years, the modeled time frame for the magmatic interval of crystallization in the pegmatite was very short, especially in its relationship to the notional correlation of crystal size with time (Figure 1 above).

The HEAT3D Thermal Diffusion Model of KWare

The solution developed by Carslaw and Jaeger (1959) for the conductive flow of heat from one body to another is an algebraic series based on successive terms of the error function, a complex integral, for temperature as a function of time. It incorporates that rate of change in the spatial dimensions. It is a cumbersome algorithm to type, but it can be solved in spreadsheets such as Excel™. As input variables, it requires the initial temperatures of the two bodies in question (e.g., melt and host rock), their heat capacities at constant pressure (the heat content necessary to raise the temperature of a body by 1 Kelvin), the thermal diffusivity (the rate of heat flow in response to a gradient) of the two materials in question, and the spatial units of measure (e.g., meters, kilometers). Following the lead of Chakoumakos and Lumpkin (1990), Morgan and London (1999) and Webber et al. (1999) employed the algorithm to calculate the cooling histories of two thin pegmatite dikes in southern California. Their results conveyed what was by then the obvious conclusion that the dikes in question, from 1 to 2 meters in thickness, would have cooled to their likely solidus temperatures in hours to days. The results put the crystallization in pegmatite-forming melts, and their large crystals, on a short timescale, but one that is comparable to the timescale of laboratory experiments.

In 1995, Dr. Ken Wohletz developed the program HEAT3D: A 3-D Heat Flow Simulator[1] (also refer to the Appendix). He made that program freely available in 2003, with updates as recently as 2016, on the website:

https://www.lanl.gov/orgs/ees/geodynamics/Wohletz/KWare/Index.htm[2]

The user interface is straightforward, and as Wohletz noted (Appendix), it allows for the calculation of much more complex and variable (in space and time) heat flow simulations than the original one-dimensional model of Carslaw and Jaeger (1959). The input parameters for the model include:

- the units of the spatial scale and dimensions (2D or 3D)
- the geothermal gradient (rate of increase of temperature with depth)
- the depth to the top of the 2D or 3D matrix, which is called a “mesh” in the HEAT3D program
- the bulk density, heat capacity and thermal diffusivity of the rock or rocks (any number of rocks in any geometric shape and at any scale can be chosen)
- the composition (mafic, intermediate, silicic) and initial temperature of the magma body, its heat capacity and thermal diffusivity.

Appropriate values for the bulk densities, heat capacities, and thermal diffusivities of the materials in question are available through various professional publications[3]. They are mostly derived from laboratory measurements, and the HEAT3D program treats them as constants. As is evident from the effort of Chakoumakos and Lumpkin (1990), the most important variables of

the thermal model, and those that are least well constrained, are the initial temperature of the magma and especially the temperature of the host rock at the time of emplacement.

A feature of granitic pegmatites that has likely been observed but not noted by countless geologists is that they contain no inherited crystals (London, 2008). The pegmatite-forming melts, therefore, are injected entirely in the liquid state. If one accepts that the pegmatite-forming melts arise from larger plutonic bodies of granitic magma, and that the pegmatite-forming melts represent the late-stage residual melts in such bodies, then the temperatures of those residual melts must be on or near the liquidus of the granite system close to its eutectic temperature, which is also its solidus temperature for those compositions. These are the minimum or eutectic temperature values, well known through laboratory experiments, that lie between 650°-700°C for H_2O-saturated granitic melts. For the lithium-rich Harding pegmatite, Chakoumakos and Lumpkin utilized the liquidus temperature of 650°C as the initial temperature of the pegmatite-forming melt upon emplacement. For less evolved granitic compositions, a value of 700°C would be appropriate (London, 2008). For the HEAT3D model, therefore, the magma can be chosen as "silicic" and its initial temperature can be set between 650°-700°C with little uncertainty, correct? Not so fast: the liquidus temperatures are for H_2O-saturated granitic melts. For the H_2O-saturated Harding pegmatite bulk composition, Figure 1 of Fenn (1977) showed it to crystallize over an interval of 150°C; it is, therefore, not of a eutectic composition. The liquidus temperature rose rapidly as the H_2O content of the melt decreased below saturation. The same relations held for the composition of the Spruce Pine pegmatite (Figure 2 of Fenn, 1977), which is closer to the eutectic composition of a simple granitic liquid. For the many reasons that have been mentioned in previous essays of this series (also see London, 2008, 2018), there is reasonable uncertainty as to whether pegmatite-forming melts are or are not H_2O-saturated upon their emplacement.

The variable that is most uncertain, as is evident from the assessment by Chakoumakos and Lumpkin (1990), is the initial temperature of the host rocks at the site of pegmatite emplacement. This temperature is determined by the geothermal gradient, which is the increase in temperature with depth, but which may be perturbed locally by the heat released from a proximal magma body. In their calculation, Chakoumakos and Lumpkin (1990) utilized a geothermal gradient of 30°C/km from an undisclosed source. Calculations of continental geotherms from ancient to modern times yield a value of 18°C/km (e.g., Burke and Kidd, 1978; Catlos et al., 2001). Pegmatites of all ages are affiliated with continental mountain-building events (e.g., McCauley and Bradley, 2014), and therefore the continental geotherm seems to be most appropriate. To calculate the temperature of the host rocks, their depth must also be known. Chakoumakos and Lumpkin (1990) chose a depth of pegmatite emplacement to coincide with the estimated pressure at the peak of metamorphism of the host rocks; that was 350 MPa (their Figure 1). The rate of increase in pressure with depth through the continental crust is ~ 33.3 MPa/km; at that rate, the depth would be 10.5 km, which is slightly less than the value (11-12 km) cited in their paper. In their assessment of fluid inclusion data, they put the pressure closer to 250 MPa, which would correspond to a depth of emplacement of 7.5 km. The consequent range of host rock temperatures is presented in Table 1.

Table 1.

Geotherm:	18°C/km	30°C/km	Other
Depth, km	Temperature at depth, °C		
7.5	160	250	550
11	223	355	

In another calculation that was not presented in their paper, Chakoumakos and Lumpkin (1990) chose 550°C as the host rock temperature, presumably to represent the case that the emplacement of the pegmatite coincided with the peak of metamorphism of the host rocks. However, with a geothermal gradient of 30°C/km, the depth to reach this temperature is 18.3 km; with a gradient of 18°C/km, that depth would be 30.5 km, close to the base of the continental crust. Neither value is plausible because the Harding dike, like most pegmatite bodies, intruded its host along fractures (Jahns and Ewing, 1977). Beyond a depth of ~ 9 km, rocks become ductile and planar fractures cannot exist (Brisbin, 1986).

The magnitude of the uncertainty that this range of initial host rock temperatures introduces can be quickly assessed through the HEAT3D model. As an example, two runs of the program were conducted for a horizontal silicic body 20 meters thick at an initial temperature of 650°C, at the depth of 7550 meters to the dike centerline, and geothermal gradients of 18°C and 30°C/km. The time to reach 550°C, the solidus of the hydrous Harding composition, is 1.92 years with the lower geothermal gradient, and 2.74 years in the second case (same as shown in Figure 7 of their article). The answer, then, is that the choice of the geothermal gradient has a negligible effect on the rate of cooling. It is very fast in both cases.

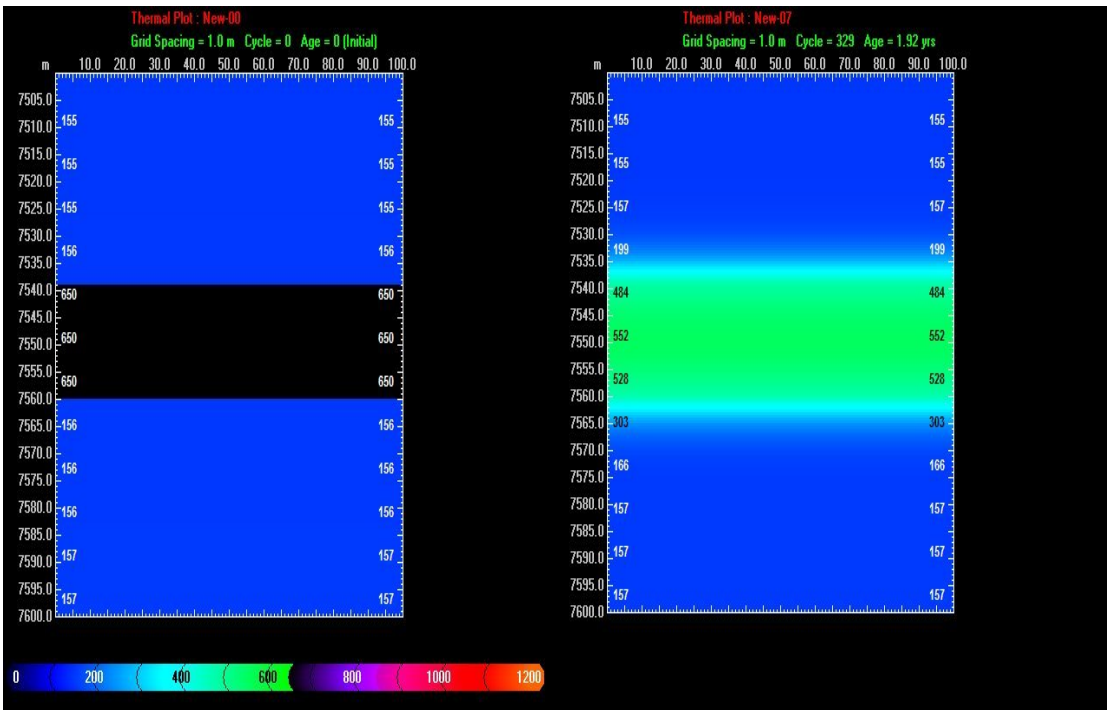

Figure 2. Initial (left) HEAT3D model using a geothermal gradient of 18°C/km, and final cycle (right) to 550°C at the center of the dike..

The heat from a large and nearby magma body – for example, the pluton from which the pegmatite-forming melt was derived, and possibly a much older and larger batholithic complex beneath it – can cause a perturbation to the normal geothermal gradient. The effect of an elevated thermal aureole around a granitic body can be modeled approximately in the HEAT3D program by choosing a geothermal gradient that gives the desired temperature at depth[4]. The same HEAT3D model was run for the same depth, rock, and magma properties except for the geothermal gradient, which was set at 70°C/km to achieve a temperature of 550°C at depth to the top of the magmatic sill. In that case, the center of the dike cools to 560°C in 250 years; the center of the dike reaches 557°C after 450 years, and it remains at that temperature beyond 3,000 years later, when the model run was terminated. The examples illustrate that the heat from the pegmatite body – even a large pegmatite of 20 meters thickness – is negligible in relation to the heat sink that the semi-infinite mass of the host rock represents. For this reason, the initial temperature of the host rock at the time and site of emplacement is the more critical parameter.

Rubin (1995) attempted to model the extraction of a dike of granitic melt from a larger magma body, exactly as Jahns and Burnham (1969) envisioned it for the granite-pegmatite paradigm. A granitic pluton was allowed to crystallize and dissipate its heat to the surrounding host rocks. Rubin's (1995) model employed variables that included the temperatures of melt and host rocks, the viscosity of the melt as functions of composition and temperature, and a term for the magma driving pressure, which at its lowest value is the gravitational buoyancy of the residual melt, and at its greatest might be caused by filter pressing – the compaction of a crystalline mush by compression – and the upward extraction of the melt. Rubin's (1995) general conclusion was that dikes have insufficient heat to propagate beyond the thermal aureole of their source pluton. The size of the aureole and hence the distance that melts can travel is determined by the volume of the source pluton, the quantity of its heat loss, and the initial temperature of the rocks that host the igneous body.

So the question remains, what is the temperature of the host rocks upon emplacement of a pegmatite dike if the temperature is higher than the normal geothermal gradient, lower than the peak metamorphic grade of the host rocks, but elevated above the geotherm by a pluton or batholith that might not be known? There are few good answers. Chakoumakos and Lumpkin (1990) chose to represent the cooling path of the Harding pegmatite (their Figure 6) with two segments: a segment of isobaric cooling to 350°C, followed by a path along a geothermal gradient to the surface upon uplift. The isobaric segment followed from their analysis of the fluid inclusion isochors (their Figure 5). Though their Figure 5 shows that isobaric path at ~ 250 MPa, they chose to represent it at 350 MPa in their Figure 6. The inflection in the cooling path occurred at 350°C, their calculated temperature of the host rocks based on the geothermal gradient. London (1986) constructed a cooling curve for the giant Tanco pegmatite, Manitoba, in much the same fashion. London (1986) and Brisbin (1986) arrived at the same pressure (depth) of emplacement – 280 ± 10 MPa – through entirely different means. At that pressure, the cooling history was approximately isobaric to 350°C, at which point the cooling curve was forced to deflect into the stability field of eucryptite +quartz, and hence along (down temperature and pressure) a geothermal gradient with uplift. In that case, 350°C represents the temperature of the host rocks at the time of pegmatite emplacement. That is higher than the temperature of a normal continental geotherm (Table 1) calculated to a depth of ~ 8.5 km (280 MPa). It implies that the temperature of the host rocks to the Tanco pegmatite was elevated by heat from a nearby pluton

or batholith, which might have been the giant Lac du Bonnet batholith or the Birse Lake pluton just to the south of the pegmatite (Gilbert et al., 2008).

Temperatures of Crystallization in Pegmatites

Another approach to the problem of time and temperature is to determine the actual temperatures at which pegmatites crystallize. The temperature of crystallization along the margin is especially informative because the heat flow models predict that the temperature of melt and host rock along their contact immediately reaches the midpoint of their respective body temperatures. London et al. (2012) conducted a detailed study of feldspar compositions and crystal structure on samples from the Main (2.5 meters thick) and Swamp dikes (1 meter thick) of the Little Three mine near Ramona, California. Feldspar thermometry reflected nearly isothermal crystallization at ~420°–430°C from the margins to center, though the temperatures recorded by the feldspars increased slightly toward the core of the Swamp dike section. London et al. (2020a) added detailed chemical and structural data for feldspars from three other thin pegmatite dikes in San Diego County, California. They summarized the results in the abstract:

"*Feldspar solvus thermometry from these three dikes indicates that their pegmatite-forming melts crystallized at ~ 375–475 °C. Those low temperatures are consistent with the occurrence of granophyric plagioclase– quartz intergrowths along the borders of pegmatites, thick and thin, that arise from thermal quenching of their melts against much cooler host rocks, and hence at much shallower depths than the igneous sources of the pegmatite-forming melts*."

Again, the cooling history recorded by the feldspars was nearly isothermal but increased slightly toward the core of some bodies.

These pegmatite are chemically simple, and their compositions are close to that of granite (London et al., 2012). Therefore, if the temperature of the melt was originally 700°C, and the feldspars record the initial cooling temperature at the margin of 425°C, then the host rock temperature had to be close to 150°C. That is the temperature of rocks at a depth of 5 to 8 kilometers, depending on a choice of the two geothermal gradients cited in this essay. The pegmatites are miarolitic, which would be favored by shallower levels of emplacement.

London et al. (2020b) assessed the time-temperature relations of crystallization in miarolitic pegmatites based on feldspar solvus thermometry and the known stability limits of some of the mineral assemblages (Table 2). For a dike 1 meter wide, the main stage of cooling from the liquidus to the temperature of 435C, which was the temperature record by feldspars at the margins of the pockets and the same as that of massive pegmatites (London et al., 2020a), spanned 12 to 28 days depending on the host rock temperature (160°C at the normal continental geothermal gradient, or elevated to 300°C by heat from a buried pluton). if the dike crystallized completely over those intervals, then the rates of advance of the crystallization front would be ~ $2*10^{-4}$ to $4*10^{-4}$ cm/sec, which is not unrealistic in relation to the minimum growth rates noted by Swanson (1977) for the reasons already stated. The interval over which large primary crystals of tourmaline, quartz, beryl, etc. might have formed within the miarolitic cavities, from 435°C to 355°C, might have spanned 28-183 days. Again, for a crystal dimension of 10 cm, its modeled growth rate would have been ~ $4*10^{-6}$ to $8*10^{-7}$ cm/sec, which is in line with the growth rates measured from experiments. The example of the demonstrates that the modeled cooling rates of

pegmatites, though rapid, appear to be in line with the rates of growth measured from experiments. However, it is important to note that although the experiments of Swanson (1977) and Fenn (1977) were conducted in a highly undercooled melt, the actual temperatures of those melts (~ 700°C) were substantially higher than those that are indicated for natural pegmatites.

Table 2. Duration of crystallization in a miarolitic pegmatite[1]

Depth:	7 km	7 km
Host temperature:	160°C	300°C
Initial melt temperature:	700°C	700°C
	Time after emplacement	Time after emplacement
Massive pegmatite, margin to center @ 435°C:	12 days	28 days
Primary minerals in cavities to 355°C:	26 days	183 days
Closure of perthite exsolution @ 335°C	40 days	2.7 years
Laumontite stable @ 285°C:	55 days	> 13.2 years
Clays follow stilbite @ 185°C	> 1 year	
Thermal equilibrium with host rocks:	> 200,000 years	13.2 years

[1] Refer to the text of London et al. (2020b) for additional boundary conditions used in the model

Concluding Remarks

One of my longstanding favorite maxims has been "don't do the work if you're not prepared to accept the results." Scientists are often guided by intuition or by fact-based knowledge that leads them to a testable hypothesis. In other cases, an idea comes along for which there is no preconceived expectation of an outcome. That led me to ask Bryan Chakoumakos if he was surprised by the result shown in their Figure 7. In a recent email exchange, he replied:

"*As far as the cooling calculations for the Harding pegmatite I was not surprised by the result. The physics of heat flow are well established, and pegmatite bodies are relatively small as compared to say the average size pluton*."

One could say that this conclusion should have been obvious all along, and perhaps it was to those like Jahns who advocated for rapid crystallization that results from rapid cooling. The textures of pegmatite minerals are those of melts that have crystallized at a highly undercooled and supersaturated state with respect to the minerals that can form. On a field trip to the Himalaya pegmatite in 1986, I asked Professor Ichiro Sunagawa, a renowned expert in crystal growth, what he thought of the textures. He said "*they look like devitrified glass*." Sunagawa's comment carries an implication of crystallization at low temperature from a very viscous melt or glass. He was referring to the pronounced unidirectional and radial fabric of elongate crystals in the pegmatite that is similar to spherulitic textures seen in glasses, including obsidians. That spherulitic texture in obsidians is sometimes referred to as "snowflake", but the radial spherical aggregates of crystals in obsidians can reach meter-scale dimensions (Smith et al., 2001).

The recent studies of crystallization temperatures based on feldspar solvus thermometry (e.g., London et al., 2020a, 2020b) have provided the factual basis for the interpretation that pegmatites cool quickly and crystallize at temperatures well below their liquidus and even their solidus temperatures. Pegmatites are where the conventional correlation between crystal size and the duration of crystal growth (Figure 1) falls apart.

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Footnotes

[1] Copyright 1995 by the Regents of the University of California.

[2] Last accessed 01 December 2021.

[3] For those who would like to create and run HEAT3D, here are the input parameters that I normally use:

For the mesh:

- size: in meters, scale depends on model; the larger the mesh size, the slower the program runs
- depth: 7500 m
- surface temperature: accept the default of 20°C
- geothermal gradient: 0.020°C/m (same as 20°C/km)
- hit "OK"

For the model, Rock 1:

- bulk density: 3000 kg/m^3 (applies to a dense, mafic host rock, such as greenstone or gabbro)
- thermal conductivity: 3 W/m*K (applies to a dense, mafic host rock, such as greenstone or gabbro)
- heat capacity: 1100 J/kg*K (applies to a dense, mafic host rock, such as greenstone or gabbro)

For the model, Magma 1:

- magma type: select "silicic" note: select "silicic" before entering any other inputs
- magma temperature: input 700°C
- thermal conductivity: 1.5 W/m*K
- heat capacity: 2300 J/kg*K
- hit "OK"

From the link "Calculate", hit "Run", or change the frequency with which the program calculates a cycle using "Acceleration" before "Run" In the "Run" dialogue box, you can reset the dump interval, and you can set the program to run continuously until you stop it.

[4] Actually, the HEAT3D program is far more versatile than this application implies. For example, see Figure 15-5 of London (2008). In that example, a large granitic sill (Magma 1) was allowed to cool. At the point that 75% of its heat content had dissipated to the host rocks, the HEAT3D run was halted, and a vertical pegmatite dike (Magma 2) was added to extend from the pluton margin to beyond the thermal aureole of the pluton. When the run was restarted, the pluton and the pegmatite continued to cool simultaneously. The portion of the pegmatite dike within the thermal aureole of the pluton cooled more slowly than did the distal portion of the dike.

Appendix: readme.tex document in the HEAT3D download folder

HEAT3D: A 3-D Heat Flow Simulator, by Ken Wohletz

Built around solution of conservation of energy by finite differencing, HEAT is a user interface that allows design of a computational mesh representing rock geometry, properties, and magma body emplacement. The following text gives some more background.

The original version of HEAT was a FORTRAN program described in the book "Volcanology and Geothermal Energy" by K. Wohletz and G. Heiken (University of California Press, 1991,

432 pages). Generally following the solution scheme described in Appendix E of that book, this HEAT version has a number of improvements.

HEAT3D is a graphically interfaced application written for a PC running Microsoft WindowsTM. Designed to study 3-D, transient thermal regimes in and around volcanic and magmatic intrusions, HEAT models a variety of geologic structures and rock properties and their effect on both conductive and convective heat flow. The graphical interface is readily used to develop and tailor the simulation to represent most geological conditions of magma intrusion and geological structure. Calculated thermal regimes are color encoded and updated in graphical display with each update stored as a file for future playback. This modeling goes well beyond that done by analytical solution of 1-D linearized expressions of thermal diffusion in that it calculates the nonlinear effects of heterogeneous media, temporal, spatial, and thermally varying properties. As with most geophysical modeling, the results of these calculations are not mathematically unique for inversion applications. However, the geological constraints applied from the first part of this study will greatly reduce the number of simulations that might satisfactorily fit observations.

This modeling goes well beyond that done by analytical solution of 1-D linearized expressions of thermal diffusion in that it calculates the nonlinear effects of heterogeneous media, temporal, spatial, and thermally varying properties. HEAT employs an explicit finite differencing scheme rather than an alternating direction implicit one in order to insure that the original differential equation solved is exactly reproduced by the finite difference equation when time and spatial steps are infinitesimal. Truncation errors that might evolve when using very short time steps are minimized by utilizing double precision. Continuous thermal gradients are assigned along the boundaries and initial conditions use a designated regional thermal gradient. Latent heats of fusion/crystallization are solved for all rocks including magma where temperatures are in that range. Convective heat transfer in the magma is calculated as a function of temperature and composition and constrained by calculated Rayleigh numbers; convective heat transfer in fluid saturated rocks is solved by the Darcy equation and constrained by Nusselt number. New magmas may be introduced into the calculational mesh at any time during the simulation. All rock/magma properties are assigned by the user and they include: density, porosity (fluid saturation), heat capacity, initial temperature, spatially and thermally varying thermal conductivities, and location. As mentioned earlier, the code has been applied to several geologic areas to test its suitability. A version of heat has been adapted for laboratory rock melting experiments involving thermal variations of rocks melted by a moving hot molybdenum probe. Results thus far have shown the method predicts measured temperatures with enough accuracy to make engineering designs based on it.

[Note: the HEAT3D download folder contains a much lengthier PDF file in which Wohletz goes through the derivation of the algorithms that are used in the EXE program.]