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The Formation and Cooling of Dykes

DR ANDREW SNELLING

INTRODUCTION

When molten (igneous) rock forces its way, or is squeezed up, into a vertical or near-vertical crack in other rocks, be they flat-lying sedimentary rocks such as sandstone or shale, or folded metamorphic rocks such as schist or gneiss, or other igneous rocks, the crack is usually forced apart and the molten rock cools in the space to form a tabular igneous intrusion cutting across the surrounding rocks that is known as a dyke (or dike, depending on which country you live in!). The molten rock (magma) crystallizes as it cools, the most common dyke rocks being dolerite (or diabase to North Americans), basalt and fine-grained granitic rocks. Although dyke formation and magma cooling in dykes have not been observed and timed, prolific studies have still been directed towards understanding the mechanism of magma intrusion to form dykes and the timing of magma cooling in them.

Issues such as the rate of cooling of magmas are critical to young-earth creationists, since the timescales usually envisaged for the cooling of such large bodies of magma as granite batholiths are usually quoted in the order of from hundreds of thousands of years to tens of millions of years. Anti-creationists and theistic evolutionists have pointed to such apparent contradictions with the young-earth timescale, thus insisting that young-earth creationists and Flood geologists are 'obviously' wrong and so their viewpoint is untenable. It is thus clearly appropriate that the data on these topics be reviewed. Dykes perhaps represent the simplest example of a cooling magma body because of their limited dimensions, and so are an appropriate place to start in our investigations.

MAGMA INTRUSION

Szekely and Reitan[1] maintain that the width of a dyke places a limit on the possible vertical distance of magma transport (that is, the distance from the source of the dyke material), provided that the magma viscosity, and the temperature and thermal properties of the surroundings and the magma, are known or can be estimated. In many cases dyke compositions indicate that the source of the magmas is the lowermost crust or uppermost mantle. The geological environment of emplacement may then dictate a distance of transport of only about 30 kilometres.

In order to define the combination of variables compatible with dyke formation by magma intrusion, Szekely and Reitan considered a simplified physical picture, as shown in Figure 1. A fissure through the earth's crust is represented by a rectangular slot of width W, which connects a magma chamber or source (with the magma at h = 0 when t = 0) to the free surface at h = H. At time t = 0, which corresponds to the opening of the fissure, the static pressure acting on the magma ΔP will force the magma upward through the fissure.

As the magma rises it encounters wall rock of progres-

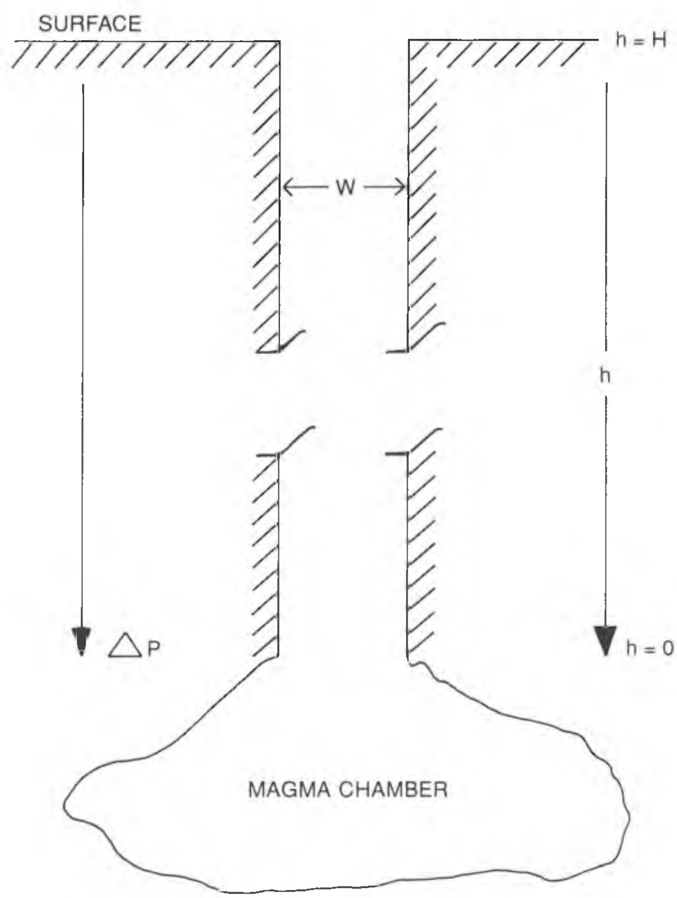

Figure 1. Physical model for the intrusion of a dyke into a fissure from a magma chamber at depth below the crust (after Szekely and Reitan[1]).

sively lower temperature. The resultant heat loss causes a substantial increase in viscosity and at least partial crystallization. The increase in viscosity, the partial blockage of the slot, and the reduction in the effective driving force will hinder upward flow. Should crystallization be completed before the free surface (or any other elevation of h above the magma chamber that is of interest) is reached, the mechanism is ruled out for the given geometry.

A rigorous formulation of the problem, where allowance is made for the two-dimensional nature of the velocity and temperature fields, and simultaneously, account is taken of the interaction of the velocity field with rate of crystallization and inward-moving (from the walls of the fissure) solidification front, results in a rather complex set of partial differential equations. Thus Szekely and Reitan considered it preferable, at least as a first approximation, to examine the asymptotic behaviour of the system. They first considered the rise of magma in a rectangular slot without crystallization and developed expressions relating the rate of rise of the magma toward the free surface to other parameters, such as width of the slot, viscosity, etc. Then they determined the time required for complete crystallization as a function of dyke width, for given values of the latent heat of fusion and the difference of temperature between the wall rock and the solidus temperature of the magma (that is, that temperature at which the originally liquid magma has completely solidified).

Using the parameters already depicted in Figure 1, at t = 0 magma enters the slot from the magma chamber and is forced upward owing to the pressure of the overburden acting on the walls of the magma chamber. To describe the rise of the magma within the slot, Szkeley and Reitan made the following assumptions:

(1) The flow is laminar throughout;
(2) Entrance and inertial effects can be neglected;
(3) There are no radial gradients within the slot;
(4) For the effect of gradual cooling of the magma throughout its rise, the viscosity can be considered as a given function of the vertical position h, varying over a specified permitted range; and
(5) Frictional heat generated by viscous flow can be neglected.

Although restrictive, the assumptions represent a reasonable compromise between faithful representation of reality and the desired mathematical simplicity. Assumptions 1, 2 and 5 can be justified in hindsight from their computed results, which show that the rise time is only significantly affected by the passage of the magma in the upper portion of the slot, where the assumptions would appear to be valid. Assumption 3 is very reasonable for most cases, in view of the fact that the ratio between H (the height the magma rises from the magma chamber to the free surface) and W (the width of the dyke) is very much greater than 1.

Within the framework of these assumptions and their physical model, Szkeley and Reitan concluded that the position of the top of the magma within the slot is readily described by a first-order differential equation which relates the rate of rise of the magma to the width of the dyke, the static pressure due to the overburden load, the magma density, the acceleration due to gravity, the distance risen by the magma, and the position-dependent viscosity. The variation of viscosity as a function of position in the crust is dependent on the following facts:

(1) The viscosity is an exponential function of temperature;
(2) The ambient temperature increases with depth; and
(3) Viscosity is dependent on pressure, which increases with depth.

After integration of their first-order differential equation, Szekely and Reitan derived a simplified equation which they then used to perform computations, substituting in appropriate property values:-

- The static pressure, $\Delta P = 8 \times 10^9$ dynes cm^{-2}
- The magma density, $\rho = 2.7$g cm^{-3}
- The viscosity at $h = 0$, $\mu_o = 10^5$g cm^{-1}sec^{-1}
- The viscosity at $h = H$, $\mu_H = 10^{10}$g cm^{-1}sec^{-1}

Their computations were then used to construct curves relating the time of rise of the magma to the width of a dyke for six different heights above the top of the magma chamber (see Figure 2). Also shown in this figure are curves relating time required for complete crystallization to the width of a dyke, for two different values of the latent heat liberated by crystallization.

In their computation of the time of crystallization, Szekely and Reitan assumed that heat transfer is by conduction only and solely in the direction perpendicular to the dyke in a semi-infinite medium. Parameters used were:-

- The thermal diffusivity of the magma
  4×10^{-3} cm^2 sec^{-1}
- The thermal diffusivity of the surrounding rock
  6×10^{-3} cm^2 sec^{-1}
- The initial temperature of the magma, taken to represent the liquidus temperature (the lowest temperature at which the magma is still totally liquid)
  1200°C
- The temperature at which crystallization is assumed to be completed
  1000°C
- The latent heat of fusion liberated uniformly over the temperature range 1200°C to 1000°C
  150 or 100 cal g^{-1}
- The initial temperature of the surrounding rock
  600°C

These assumptions made by Szkeley and Reitan avoid **underestimating** the time required for complete crystallization.

Assessing the results as plotted on Figure 2, it should be noted that the region to the lower left of the lines

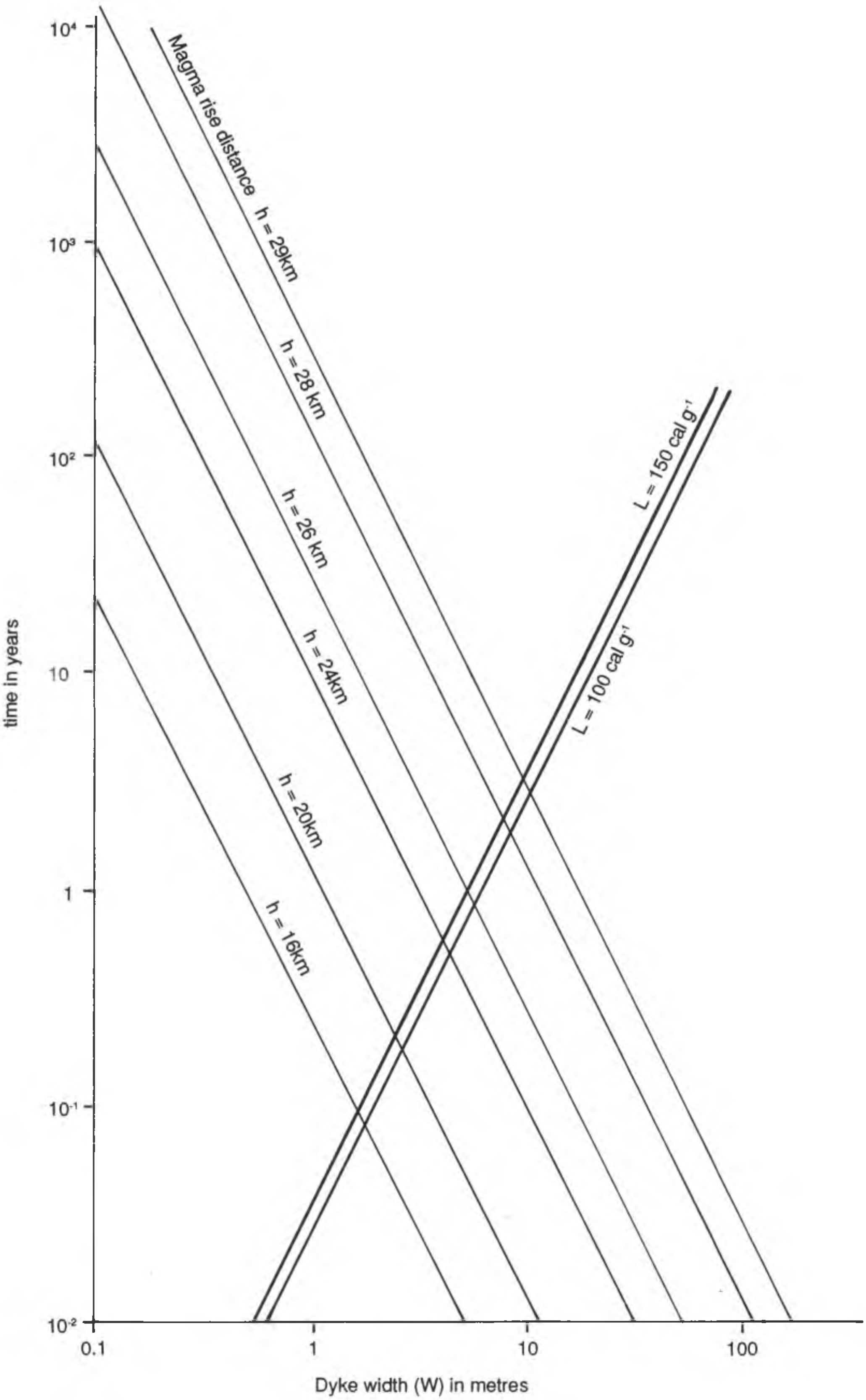

Figure 2. *Graph showing the time of rise of magma versus dyke width W, with height of rise of the magma through the earth's crust as the parameter h, and also time of complete crystallization versus dyke width, with the latent heat of crystallization as the parameter L (after Szekely and Reitan[1]).*

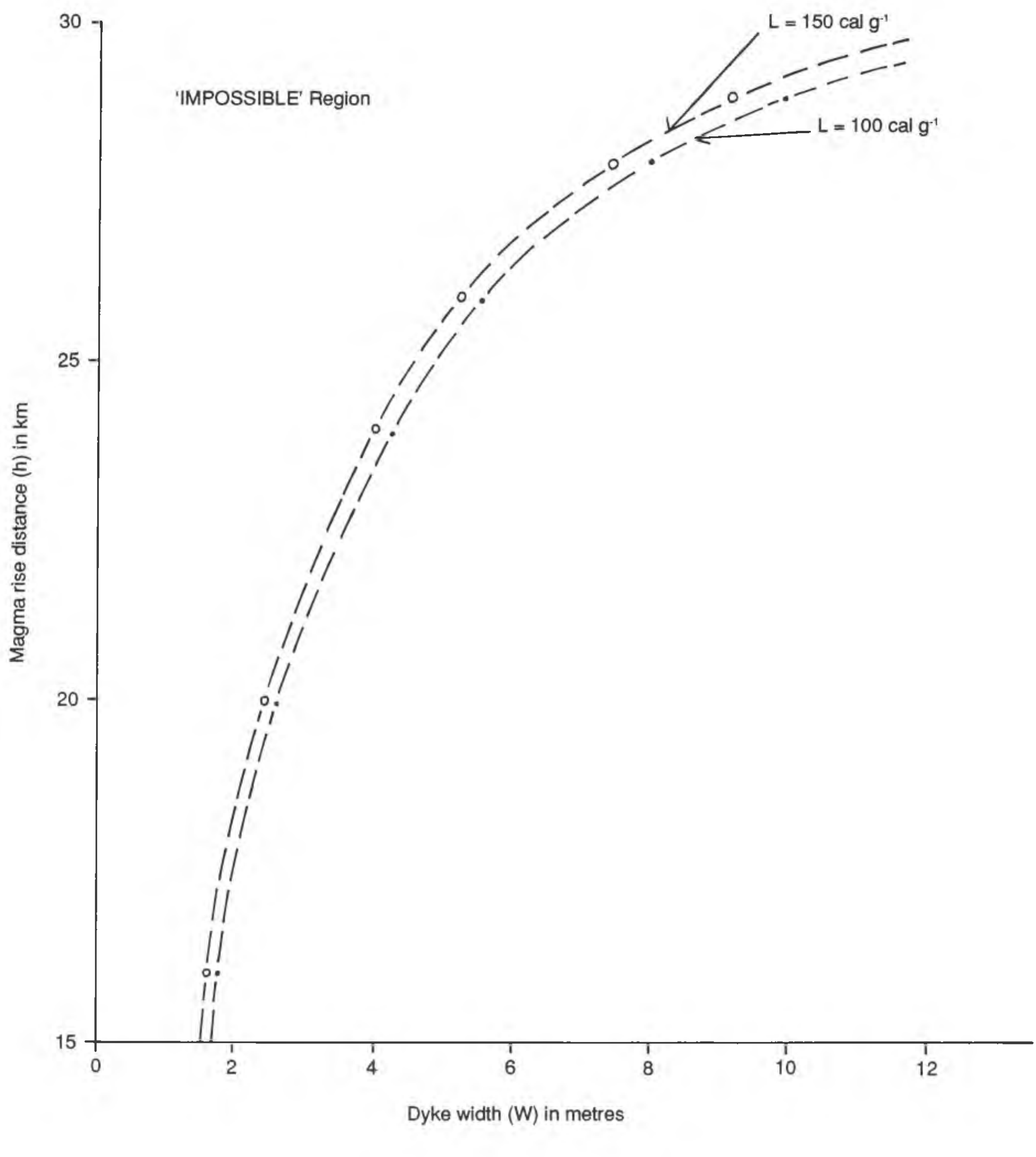

Figure 3. Graph showing the locus of intersections of height of rise of the magma h versus width of the dyke W in Figure 2, defining an 'impossible' region above and/or to the left of the curves and a 'permitted' region below and/or to the right of the curves (after Szkeley and Reitan[1]).

labelled with values of h is ruled out, as the magma would not have reached the height in the time available. Similarly, the region to the upper left of the lines labelled with values of L (the latent heat) is ruled out, since complete crystallization of the dyke will have occurred. Critical limiting values of dyke width for any combination of latent heat fusion and height above the magma chamber are given by the intersections of the lines. The locus of such intersections, relating height of rise of magma to dyke width for values of L, is shown in Figure 3. The curves define an 'impossible' region to the upper left and a 'feasible' region to the lower right.

Of particular interest are the timescales involved in the rise of the magma from the considerable depths of the lowermost crust and for crystallization to occur. Since the region to the left of the lines labelled with values of latent heat L is ruled out, this imposes a **maximum** time for the rise of a magma into a fissure and subsequent crystalliza-

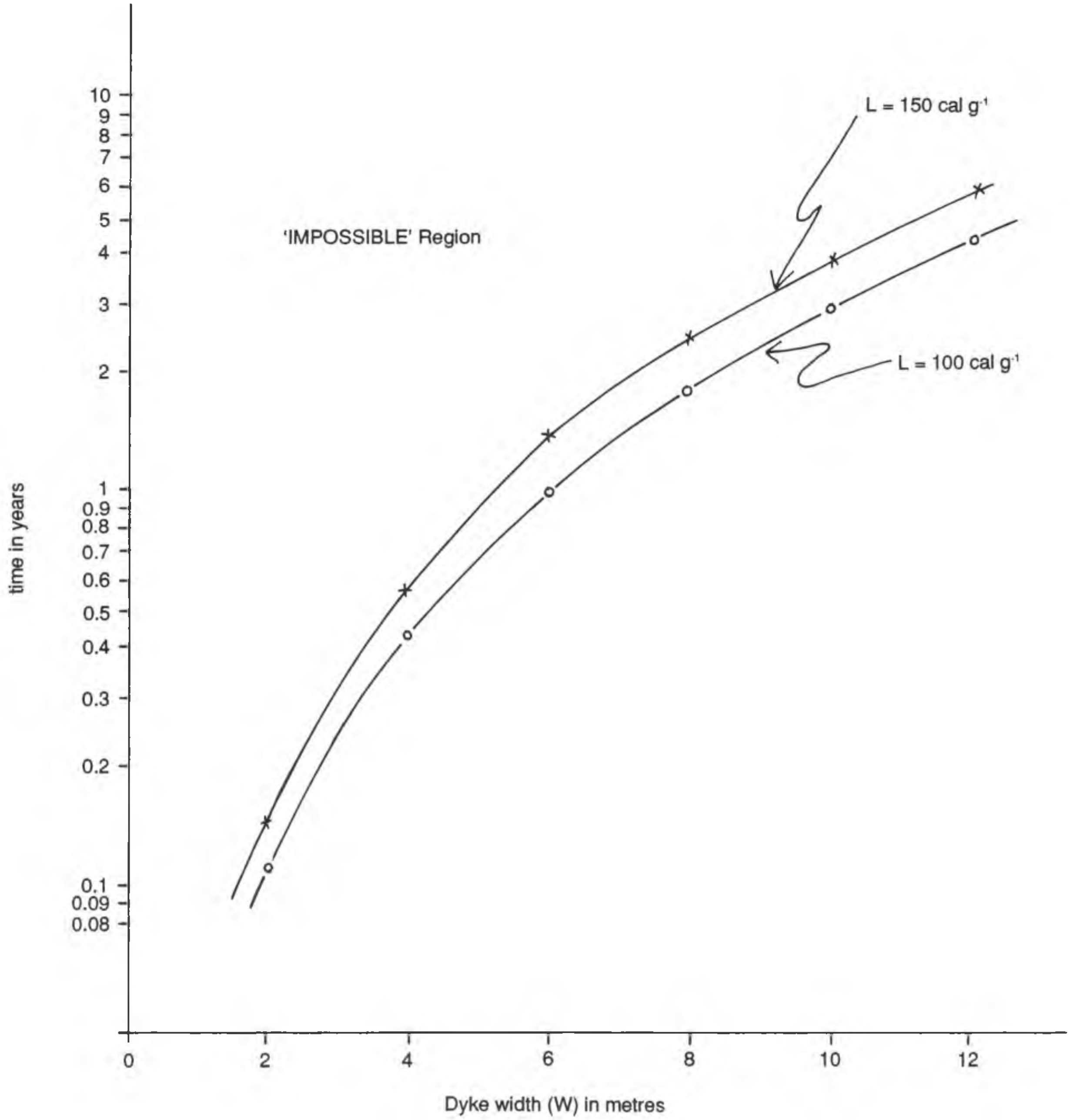

Figure 4. Graph showing the locus of intersections between time in years for rise and cooling of the magma versus width of the dyke W as derived from Figure 2, again defining an 'impossible' region above and/or to the left of the curves and a 'permitted' region below and/or to the right of the curves.

tion. Taking the critical limiting values of dyke width for any combination of latent heat fusion again (but this time the intersections of the lines give the timespans involved) the locus of such intersections, relating time of rise and crystallization to dyke width for values of L, is shown in Figure 4. Again, the curves define an 'impossible' region to the upper left and a 'feasible' region to the lower right.

Comparing Figures 2 and 4, it is thus possible to calculate that for a dyke of width 12 metres, the time taken for the magma to rise from a depth of around 29 kilometres and then crystallize is definitely less than five or six years, and probably less than three years. Significantly, most dykes are narrower than 12 metres, so this represents a limiting timespan for dyke formation. Furthermore, Szekely and Reitan concluded that their approach, and their selection of values for their computations, tended to underestimate.

At dyke widths greater than 12 metres, Szekely and

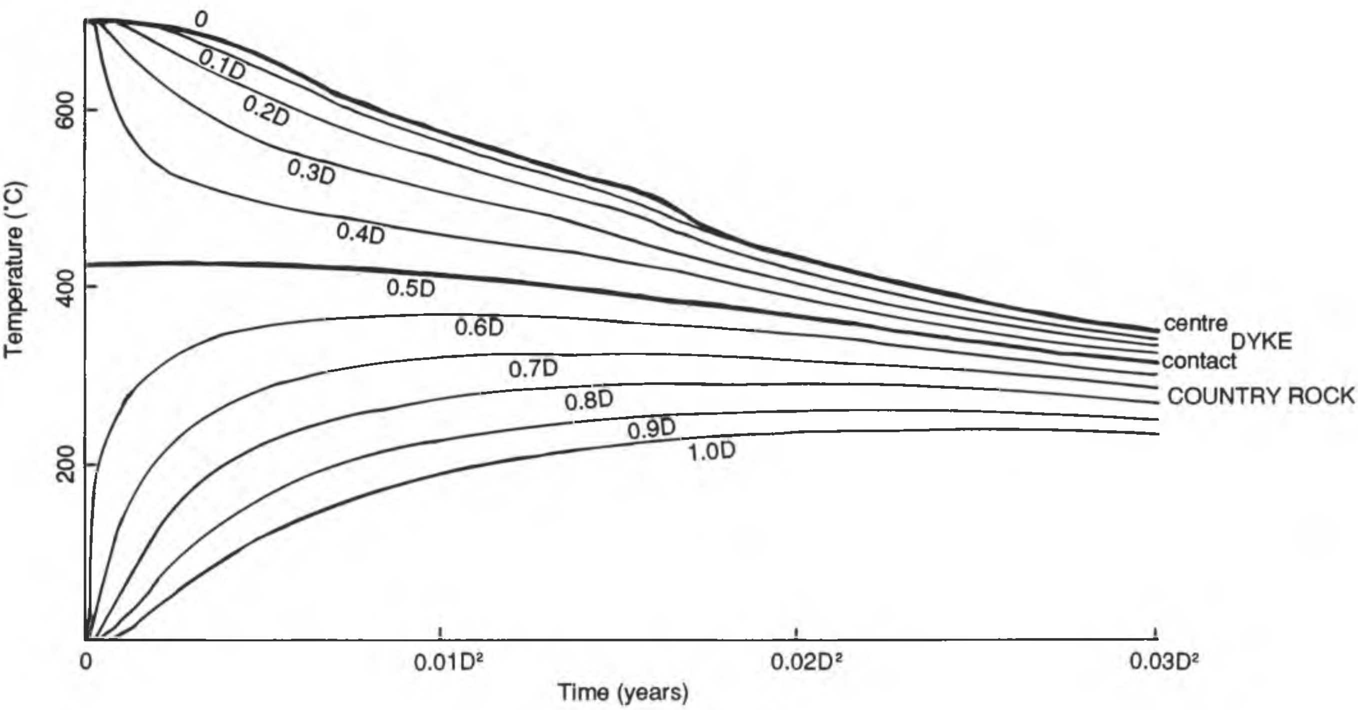

Figure 5. Graph showing the family of curves for the cooling history of a dyke of thickness D metres for which the intrusion temperature is 700°C, the solidification temperature 500°C, and the latent heat of solidification is 80 cal g^{-1}. The labels on the curves are distances from the centre of the dyke measured as fractions of the thickness D (after Jaeger[2]).

Reitan's analysis appears to be inadequate. On Figure 2 it can be seen that it is impossible to relate the height of the magma rise parameter to the time and width parameters, but even a dyke as wide as 100 metres, ignoring the height of rise parameter, can be seen, according to the graph, as taking less than 300 or 400 years to form and crystallize.

COOLING TIMES

In an earlier study, Jaeger[2] modelled the cooling by conduction of an intrusive sheet of magma — a dyke (or sill, in the case of a horizontal sheet) — taking into account the effect of heat of solidification. As in any analysis of this type, a number of general assumptions must be made. Jaeger assumed that:

(1) The intrusive body is in the form of a flat sheet extending indefinitely in the perpendicular direction — a reasonable approximation to most dykes.

(2) The intrusion takes place rapidly, that is, in a time which is very short compared to the time of solidification.

(3) The country rock is at constant temperature which will be taken to be zero. (If it is not zero, Jaeger concluded that the results of his analysis should be added to the initial country rock temperature.)

(4) The magma is supposed to be intruded at a constant temperature, which, for simplicity, will be taken to be the liquidus temperature appropriate to its composition. (There is good geological evidence that this is frequently the case.)

(5) The thermal properties of the country rock and solidified magma are the same and are independent of temperature. (This assumption was again made for simplicity, but it is a reasonable one in view of the uncertainty of these quantities.)

(6) Heat removed by escaping volatiles need not be considered. (Since the effects of volatiles are dispersed over a region whose thickness is many times that of the magma sheet, it is reasonable to assume that the heat conveyed by them is also dispersed over a large region and so will make little contribution to the temperature in the narrow region which is heated by conduction from the cooling sheet.)

(7) The effects of convection in the magma can be neglected and cooling is by conduction only.

After developing a numerical solution from first principles, Jaeger presented his results as families of curves of temperature against time. Solutions for three typical cases are presented in Figures 5, 6 and 7. In each case the temperatures are plotted as functions of the time in years for selected points of the dyke, measured from the centre of the dyke as distances 0, 0.1D, 0.2D, 0.3D and 0.4D (in

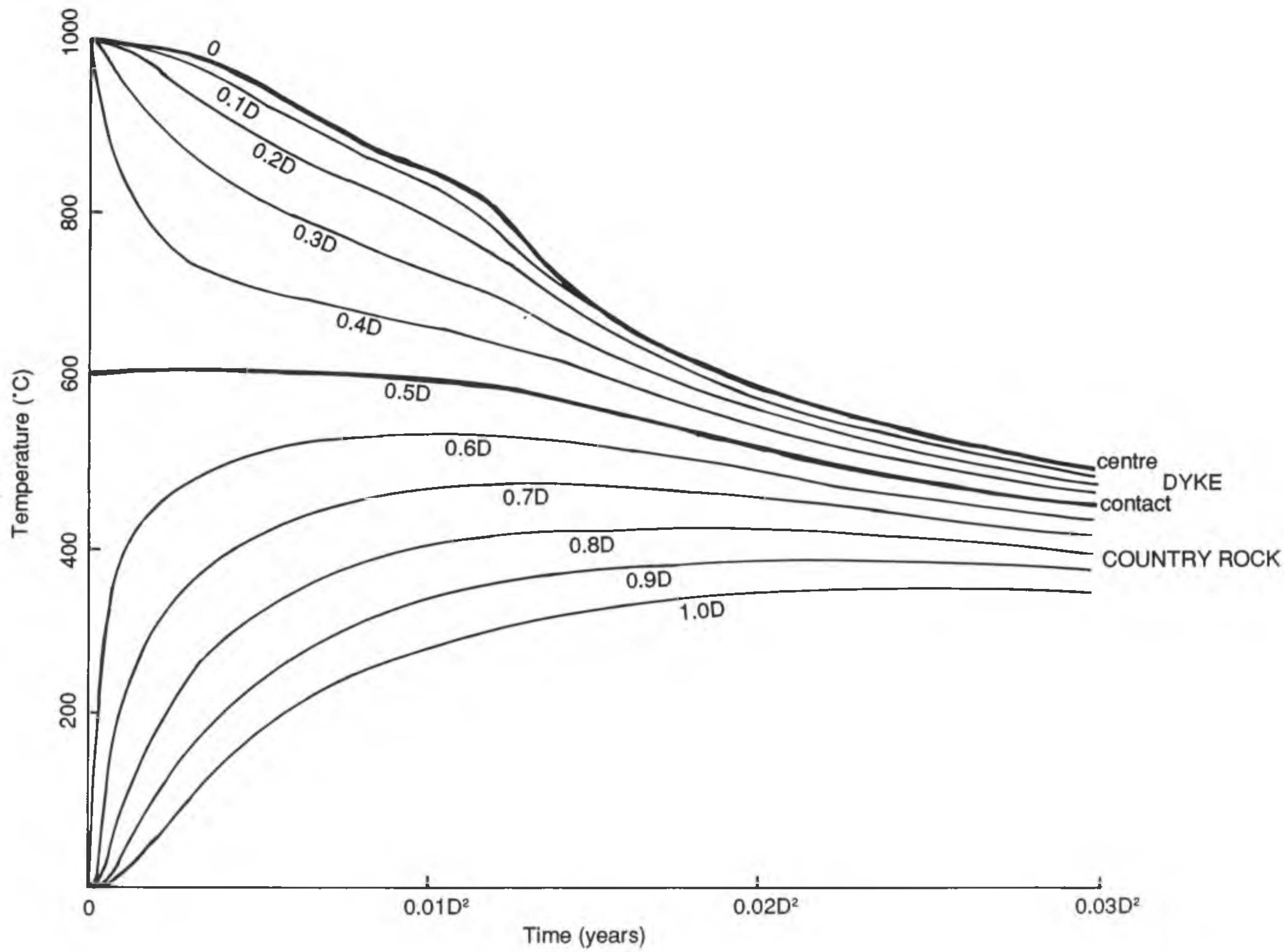

Figure 6. Graph showing the family of curves for the cooling history of a dyke of thickness D metres for which the intrusion temperature is 1000°C, the solidification temperature 800°C, and the latent heat of solidification is 100 cal g^{-1}. The labels on the curves are distances from the centre of the dyke measured as fractions of the thickness D (after Jaeger[2]).

the dyke), 0.5D (which is the edge of the dyke and the contact with the country rocks), and then 0.6D, 0.7D, 0.8D, 0.9D and 1.0D (in the country rock), where D is the thickness of the dyke in metres. The timescale is expressed as a factor of D^2.

Figure 5 represents the family of cooling curves for a typical granitic dyke, where the temperature of the magma at the time of intrusion T_2 = 700°C, the temperature of solidification T_1 = 500°C, the latent heat of solidification (or fusion) L =80 cal. g^{-1}, and the density of the magma ρ = 2.7 g cm^{-3}. The cooling time for complete crystallization (or solidification) is the time taken for the temperature of the magma at the centre of the dyke to fall from its initial temperature T_2 to the temperature of solidification T_1. In this case it is the time taken for the magma at the centre of the dyke to fall from 700°C to 500°C, and can be read directly from the graph as $0.016D^2$ years, where D is the thickness of the dyke. Thus a granitic dyke 1 metre wide would take 0.016 years to crystallize, a 2 metre wide granitic dyke would take 0.064 years, a 5 metre wide granitic dyke would take 0.4 years, and a 10 metre wide granitic dyke would take 1.6 years to crystallize, etc.

However, this does not take into account complete cooling to ambient temperatures from the temperature of crystallization of 500°C. Once the granitic dyke reached that crystallization temperature at its centre, it would then take still further time for the dyke and the surrounding country rock to cool right down again to ambient temperatures. This of course is depicted in the diagram by the partial extension of the family of curves beyond the time of crystallization. If the curves are projected (extrapo-

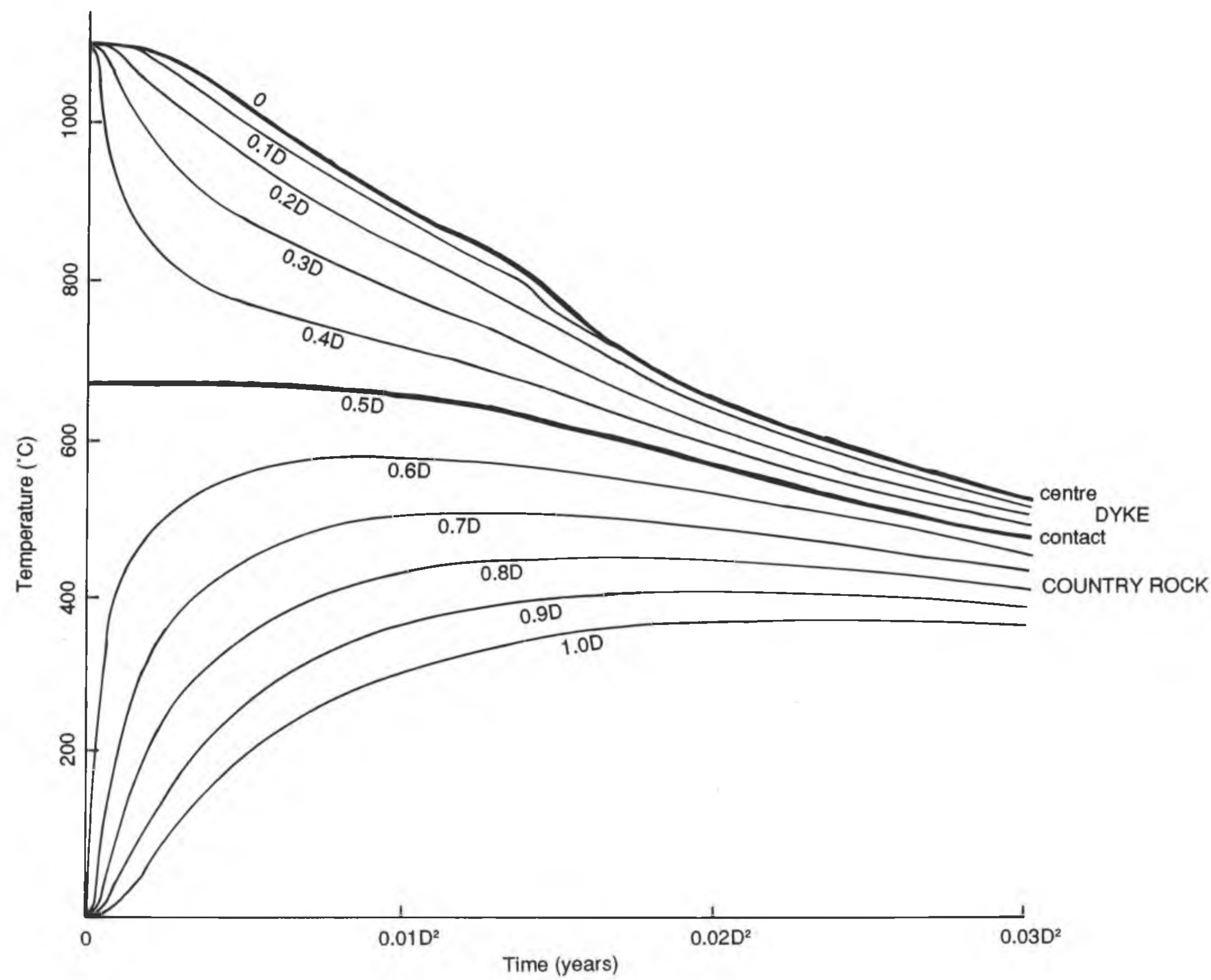

Figure 7. Graph showing the family of curves for the cooling history of a dyke of thickness D metres for which the intrusion temperature is 1100°C, the solidification temperature 800°C, and the latent heat of solidification is 100 cal g^{-1}. The labels on the curves are distances from the centre of the dyke measured as fractions of the thickness D (after Jaeger[2]).

lated) even further until the dyke is completely cooled, the time taken would appear to be about $0.09D^2$ years. In the case then of a 10 metre wide granitic dyke, complete cooling would appear to take about 9 years.

Figures 6 and 7 correspond to a typical dolerite (diabase) dyke. In Figure 6 the initial temperature of the magma T_2 is 1000°C, while in Figure 7 it is 1100°C. Otherwise, the other parameters — the temperature of solidification T_1 = 800°C, the latent heat of solidification L = 100 cal g^{-1}, the density of the magma ρ = 2.8g cm^{-3} — are the same in both examples. The times for complete solidification (crystallization) are $0.012D^2$ and $0.014D^2$ years respectively. Thus, a dolerite dyke intruded at a temperature of 1000°C which is 1 metre wide would take 0.012 years to cool to 800°C, a 2 metre wide dyke would take 0.05 years, a 5 metre wide dyke would take 0.3 years, and a 10 metre wide dyke would take 1.2 years, etc. Similarly, a dolerite dyke 1 metre wide intruded at 1100°C would take 0.014 years to cool (crystallize completely) to 800°C, a 2 metre wide dyke would take 0.056 years, a 5 metre dyke would take 0.35 years, and a 10 metre wide dyke would take 1.4 years, etc.

Again, the dykes need to then cool from 800°C to ambient temperatures, and so some further time is needed for complete cooling. If the families of curves are again projected until the dykes are completely cool, the time taken in both the examples of Figures 6 and 7 would again be about $0.09D^2$ years. Thus a 10 metre wide dolerite dyke would also appear to take about 9 years to completely cool.

The results set out in Figures 5, 6 and 7 apply to deeply buried dykes, where the magma has not reached the surface and there is a considerable amount of rock cover above the top of the dyke. Of course, nearer the surface cooling will be more rapid, so the results of this analysis represent an upper time limit on crystallization. In any case, Jaeger concluded that his results should be applicable at depths below the surface greater than the thickness of the dyke.

In a subsequent study, Jaeger[3] confirmed his earlier calculations and results. His additional comments are of importance here too. For example, he states that if volatiles (principally gases) are liberated on solidification and escape rapidly (for example, through cracks), the heat carried by them away from the dyke will have the effect of accelerating the process of cooling of the dyke. Additionally, if the dyke has intruded into sediments or metamorphic rocks that are wet, because of either ground or pore water, the effect will be to lower the contact temperatures by amounts of up to 50°C and slightly shorten the times of solidification and complete cooling.

Jaeger also reported a simplified expression for cooling time, which encompasses all the relevant parameters:-

$\tau = \kappa t/a^2$

where

τ is a Fourier number or dimensionless time;

κ is the thermal diffusivity (and accounts for thermal conductivity, magma density and the specific heat);

t is the time in years; and

a is half the width of the dyke (D = 2a).

Substituting in what Jaeger described as a reasonable value for κ (0.01 cm^2sec^{-1}), and rearranging the equation to calculate cooling time, he found that

$$t \approx \frac{\tau a^2}{31.5}$$

Furthermore, he suggested that for $\tau<0.01$ cooling is only superficial; for $\tau = 0.1$ it will have penetrated to the centre of the dyke; for $\tau = 1$ there is substantial cooling at the centre; and for $\tau > 10$ cooling is practically complete. Consequently, if a value of τ for complete cooling is set at 12, then a 10 metre wide dyke (a=5) would cool completely in approximately 9.5 years, a similar estimate to the extrapolation of Jaeger's families of curves.

A FIELD EXAMPLE

The calculations and results reviewed here, derived as they are from theoretical considerations and first principles, have successfully been used in practical situations where they correlate well with other field observations.

Perhaps one of the most unusual field examples of dykes are those studied by Wiebe[4] along the northern coast of Cape Breton Island, Nova Scotia, Canada. At this locality are found a number of dykes, some consisting of homogeneous basalt, and some of homogeneous granite, but yet others consisting of a composite granite-basalt mixture. The latter are unusual, since basaltic and granitic magmas are not normally known to co-exist like this. The geological setting of these composite dykes allows reasonable estimates to be made of the time required for solidification of each magma. Indeed, the solidification time had to have been short, otherwise the basaltic and granitic magmas in these composite dykes would have begun to mix, which the field observations, petrographic observations and geochemical analyses indicate has only just begun on a very small scale.

Wiebe observed that at least one of these composite dykes varied gradually over a distance of 25 metres along its length from homogeneous basalt, to pillowed basalt (rounded blobs of basalt that look like pillows) with granitic veins between the pillows, to granite with less than 50% pillow-like inclusions of basalt. In all the composite dykes, mixing of the two magmas appears to have formed by a progression from simple curved contacts to intimate granite veining of the basalt, to ultimate subdivision of basalt into smaller pillows and lenses. These and many small-scale features suggest that the basalt and granite were magmas when they initially made contact with each other in the composite dyke. Because the granite clearly remained mobile longer than the basalt, the latter must have cooled first, even though the basalt magma may have been intruded at a higher temperature than the granite.

Temperature and pressure in the country rock at the time of dyke emplacement were difficult for Wiebe to estimate, the chilled margins of most dykes suggesting low temperatures and probably low pressures. The existence of the mineral hornblende in the basalt suggested load pressures greater than 1 kilobar (kb) were likely. Wiebe thus estimated the temperature of the country rock to have been 200°C and the pressure 2kb. Initial temperatures of the basaltic and granitic magmas would have been 1100°C and 750°C respectively, such high temperatures in the dykes thus ensuring that even if there was a variation of as much as 200°C in the temperature of the country rock, this would affect the cooling times of these magmas by less than 50%. Mineralogical observations suggested that at the time of injection the temperatures of the magmas were slightly below these liquidus temperatures for each rock.

Wiebe then proceeded to use the standard cooling calculations of Jaeger as already outlined above. In so doing, he adopted Jaeger's values for the other properties of the magma and country rock. Other parameters used were an average dyke thickness of 50cm (some portions of these dykes were over 1 metre wide), an average dyke composition of about 70% granite and 30% basalt, an initial average temperature of the dyke segments of approximately 850°C (even though the initial temperature of the basalt and granite magmas would have been 1100°C and 750°C respectively), solidification temperatures of the basalt and granite of 900°C and 700°C respectively, a solidification temperature therefore of the dyke as a whole of 700°C, an overall density of 2.6g cm^{-3}, a thermal conductivity of 0.005 cal cm^{-1} sec^{-1} deg^{-1}, a specific heat of 0.28, and a latent heat of crystallization of 80 cal g^{-1}.

Based on the 70:30 ratio of granite and basalt, and the average initial temperature estimated at 850°C, solidification of the composite (or average) dyke is thus assumed to have been complete when the 700°C isotherm reached the centre of the dyke, that is, when all the granitic component was solid. Using the methods of Jaeger and the conditions already specified, Wiebe thus calculated that the composite dyke would have solidified completely in 27.8 hours, although he cautioned that this time must be considered only as an estimate value for a dyke of homogeneous average composition and width, since it cannot be applied specifically to any point in the real dyke. Portions of basaltic composition would have solidified more rapidly, and granite very near to the basalt would have solidified more slowly than this average value. Thus the calculated time of solidification based on the average initial temperature is probably a moderate overestimate of the actual time.

The basalt pillows particularly would have become solid much more rapidly than the granite host, and because of their discrete sizes a cooling history timeframe can be calculated. A basalt pillow would have solidified when the 900°C isotherm reached its centre, so in contact with a granite magma at 750°C the approximate time necessary for solidification of a pillow with a 5cm radius would have only been 35 minutes. Similarly, a pillow with a 10cm radius would have cooled in about 140 minutes. Thus Wiebe concluded that the 900°C isotherm would have reached the centre of a typical pillow within less than 1 hour. He also found that these short timeframes for cooling of both the basalt and granite magmas were supported by the geochemical data on the transfer during magma cooling of various elements between the basalt and granite components of these composite dykes.

CONCLUSIONS

For creationists the cooling times for fine-grained igneous rocks such as basalt and dolerite (diabase) have never been considered to be a real problem, although the timeframes involved have never been really documented in the relevant creationist literature. This review therefore has endeavoured to supply this documentation of the cooling times involved in such thin igneous bodies as dykes.

We have seen that by the application of standard cooling theory for solid bodies, the timeframes involved are indeed relatively short, and that even for a granitic dyke 10 metres wide, which is fairly rare, the cooling time for crystallization (solidification) to occur is only of the order of 1.6 years, and cooling to ambient temperatures is completed in about 9 years. It is also good to see that these estimates are confirmed by field studies such as the example cited. Since the intrusion of dykes into other rocks is an event which can be much later than the formation of the rocks into which the dyke intrudes, dyke formation in many instances could still have occurred after Noah's Flood, so these timeframes greater than the duration of the Flood year do not generate an additional problem. We have also seen that even the time taken for a magma to rise from depths as much as 30 kilometres down in the earth's crust and crystallize is also only a matter of a few years.

While such thin bodies of igneous rock would appear to be a simple solvable challenge for creationists, they provide the easiest place to start in this whole investigation of the cooling of igneous rocks. Further studies will need to delve into the cooling times for larger bodies of igneous rock, such as 1 kilometre thick sills and granite batholiths up to tens of kilometres wide.

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Dr Andrew Snelling is a geologist with a B.Sc.(Hons) from The University of New South Wales and a Ph.D. from The University of Sydney. He has worked in the mining industry and is still a consultant geologist, but now also works full-time with the Creation Science Foundation where he contributes to **Creation Ex Nihilo** magazine and edits the **Creation Ex Nihilo Technical Journal**. He resides in Brisbane, Australia.