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PLUMBING THE DEPTHS OF BATHOLITHS

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ABSTRACT. Knowledge of the pressure of consolidation of a pluton and the pressure-time history of magmatic evolution should allow better understanding of the tectonic and thermal history of the crust. Available methods to estimate pressures of plutons are mainly for those of consolidation. These are either *extrinsic,* based on geological context, or *intrinsic,* based on mineral texture, mineral assemblage, fluid inclusions, mineral inclusions, apparent cooling rates, and mineral chemistry. The methods of lattice-dimension matching of mineral inclusions and of detailed chemistry for zoned minerals could lead to pressure-time reconstructions. Future barometers based on mineral chemistry should use atomic species that have low diffusion coefficients and whose values are not sensitive to computational schemes and cumulative analytical errors. Aluminum and silicon in coexisting hornblende, biotite, pyroxene, plagioclase, or garnet are reasonable candidate phases for barometry. Pressure-temperature-time data on plutons and their crustally-derived magmas can yield insights on the paleogeophysical properties of the crust. In fact, the existence of country rocks not at grade with the pluton could constrain the thermal structure of the crust and, combined with thermal modelling, could allow us to estimate, for the time of magma generation, the minimum thickness of the crust beneath the pluton.

INTRODUCTION

In recent years, petrologists have made great strides toward understanding the history of the continental crust, stimulated in part by an enhanced appreciation of its typically complex history and constitution as revealed by seismic methods. We now recognize that a section of the continental crust may be an aggregation of many discrete blocks juxtaposed by tectonic processes and is not simply stratified by composition and/or density. We also know that the processes of crustal aggregation may assume many forms involving homogeneous strain, thrusting, underplating, or accretion and diachronous juxtaposition of units. Our impetus to understand better the history of the crust is also augmented by improved methods to detect details of crustal constitution. These methods include geochronology (especially isotopic geochronology), use of trace elements and radio-isotopes to define the nature of the source rocks, use of stable isotopes to define processes of open-system reactions in both metamorphic and igneous rocks, and use of mineral assemblages and mineral chemistry to decipher the pressures, temperatures, and fluid compositions during different stages of mineral paragenesis. Combinations of pressure (P) and temperature (T) data with the time (t) of passage of the portion of the crust through those P and T values have permitted the reconstruction in favorable circumstances of P–T–t records for metamorphic rocks.

Full understanding of the evolution of the crust must include knowledge of the paleogeophysical properties of the crust through time. For example, the thermal structure, state of internal stress, and thickness of the crust at some particular place and time when the crust was undergoing deformation (subduction, rifting, extension by decóllement, thickening by underplating, or by stacking of thrust sheets, et cetera) must influence the kinematic and dynamic response of the crust to external tectonic stresses.

The barometry of metamorphic rocks has received considerable recent attention, but the barometry of plutonic rock is less developed (see Bohlen and Lindsley, 1987 for review). Plutonic rocks can be helpful in investigating the paleogeophysical crustal properties, because the generation, transport, and solidification of a magma reflect not only the chemical nature of the source terrane but also aspects of the physical state of the crust through which the magma passed and into which it came to rest. Comparison of the P–T–t records of the country rocks and of the plutons could provide insight into the record of crustal paleogeophysics (Zen, 1988a).

This paper surveys our knowledge of the barometry of plutonic rocks. I will mention our ability to read temperature and time records of plutonic rocks only in passing. Thanks to the growing body of experimental data over a range of pressure, our ability to decode the temperature of plutonic rocks is serviceable. This in part reflects the fact that the succession of mineral assemblages within the melting intervals of silicate magmas mainly depends on the temperature; the phase-in and phase-out boundaries are less sensitive to pressure (see, for example, Wyllie, 1977). In contrast, our ability to record even the *relative* time for plutonic mineral assemblages is virtually nonexistent, partly because any sample of nonmetamorphosed plutonic rocks is polychronic; every mineral and every grain of a given mineral very likely represents a different time record. Defining *simultaneity,* that is, meaningful values of t, for mineral assemblages in plutonic rocks must remain a major challenge to petrologists who would use the rocks as P-T-t records.

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PLUTONIC BAROMETRY

How do we measure pressure in plutonic rocks? A barometer could be *extrinsic,* depending on the geological context, or it could be *intrinsic,* depending on features preserved in the rocks.

Extrinsic Plutonic Barometers

These barometers include: stratigraphic and structural reconstruction leading to estimates of the load above the pluton; nature of the contact metamorphic assemblage; and estimates of the amount of erosion or tectonic denudation during the time interval between the emplacement of the pluton and a datable and nonconformably overlying sedimentary or volcanic sequence. The last method is useful only if the age difference across the nonconformity is small enough to bound the estimated amount of erosion to acceptable values. The extrinsic barometers only yield pressure estimates for the consolidation of the magma body and preserve no earlier history.

Intrinsic Plutonic Barometers

The intrinsic barometers are of five kinds and depend on (1) the textural relations of the plutonic rocks, (2) the nature of the mineral assemblages, (3) the mineral chemistry of the assemblages, (4) the nature of inclusions in the minerals, and (5) discordant argon cooling spectra.

Textural indicators.—A common textural indicator of the depth of emplacement of a pluton is the presence of miarolitic cavities. Such cavities, if formed during the consolidation of the magma, would show that a vapor phase was present, and that the pressure was low enough to allow the cavities to form and be kept open. However, unless we know the composition of the vapor phase, we cannot measure the pressure quantitatively.

Granophyric, porphyritic, and graphic-intergrowth textures in granitic rocks have all been used to suggest shallow depth of consolidation. Porphyritic textures where the phenocryst phases are also in the groundmass show that the magma sustained two different rates of nucleation, the younger of which was rapid; thus a mixed cooling and depth history might be reasonably inferred. This is not surprising, as the porphyritic texture is common in dike rocks.

In contrast, occurrence of plutons in a high-grade metamorphic and migmatitic terrane suggests possible local anatexis, and if other evidence is permissive, local generation of the magma might be surmised. Such a setting suggests reasonably deep burial at the time of magma generation.

Mineral assemblages.—Use of mineral assemblages in plutonic rocks as indicators of pressure takes several forms.

Use of magmatic muscovite as pressure estimator (Whitney, Jones, and Walker, 1976; Chamberlain and England, 1985) relies on the experimental data for the intersection of the thermal breakdown curve for ideal, endmember muscovite ($H_2KAl_3Si_3O_{12}$), in the presence of quartz, to sanidine + Al silicate (andalusite or sillimanite) with the curve for the H_2O-saturated solidus for the model haplogranite of Tuttle and Bowen (1958). The intersection occurs at a total pressure between 3 and 4 kb. The chemical complexity of the naturally occurring systems precludes any straightforward use of this barometer, however; in fact the minimum pressure of 3 to 4 kb in many instances contradicts other valid indications of pressure of emplacement (Zen, 1988b). The phengite barometer (Massonne and Schreyer, 1987) may be useful for plutonic rocks but its application to natural muscovite compositions remains to be explored.

Magmatic epidote is a potential indicator of pressure (Zen and Hammarstrom, 1984a) in rocks having a suitable mineral assemblage (quartz, high-Al hornblende, biotite, plagioclase, K-feldspar, sphene, and an iron-titanium oxide phase such as magnetite or ilmenite) and suggests crystallization at pressures exceeding 6 kb. The pressure estimate was based on the scanty experimental data of Naney (1983) at 8 kb, but the data have since been extended by Johnston and Wyllie (1988) to 15 kb. The presence of magmatic epidote by itself does not demonstrate high pressure; the pressure estimate requires the proper mineral assemblage and a proper pistacite (ps) content of the epidote. A high-ps epidote (ps $\simeq$ 33 as opposed to $\simeq$ 20–25 for the situation described) in a two-mica, hematite-bearing granite may be stable at pressure as low as 3 kb (Zen, 1988c). Eventually we may be able to establish a barometer based on epidote of variable Fe content and differing mineral assemblage as a continuous rather than an on-off barometer, but we are not yet there.

A third kind of barometer is based on the pressure dependence of the composition of the H_2O-vapor saturated granite minimum-temperature melt. An experimental system that has been widely used is the ternary haplogranite system, Or–Ab–Q, at $P_{H2O} = P_{total}$ (that is, a H_2O vapor phase is present; see for example, Whalen, 1980; Bateman, 1988; and Bateman and others, 1962, who considered some of the conceptual problems and use of this barometer). Either modal or normative proportions of the three minerals, as endmembers, are computed and projected onto the ternary system. Comparison of the clustering of data with the ternary minimum as a function of pressure is used to infer the pressure of consolidation of the rock.

The liquidus relations of the ternary feldspar system have been utilized as a barometer also. For example, Whitney and Stormer (1985), by comparing the matrix composition of the feldspars of the Tertiary Fish Canyon Tuff of the San Juan region, Colorado, with experimental data in the system Ab-An-Or-Q-H_2O, concluded that the matrix equilibriated at moderately high pressures.

Use of both of the above methods assumes (1) that the rock in question consolidated in the presence of a vapor phase dominated by H_2O, and (2) that the *liquid* part of the magmatic system has a composition given by the modal or normative mineral proportions. The first assumption may be questioned, especially for highly evolved granitic rocks, because the presence of other components such as the halogens is likely to change significantly the location of the pseudo-ternary minimum and thus the pressure estimate. The second assumption limits the valid use of the method to highly evolved, haplogranite systems (that is, the rock contains no antecedent phase compositions). For most fairly coarse-grained plutonic rocks, every mineral, and perhaps every grain of the same mineral, may have formed over a different span of time during which the liquid portion of the system could have evolved in major-element composition. A different set of problems bedevils the use of the method for partly phyric, partly microlitic volcanic rocks (see discussions by Grunder and Boden, 1987; Stormer, Whitney, and Dorais, 1987). Barring the unlikely prospect that we could accurately measure the modal proportions of the rim parts of minerals within the ternary system (with the implied simultaneity of crystallizations of these modal portions) *and* could ascertain the presence of a vapor phase heavily dominated by H_2O, the use of retrofitting quantitative mineral assemblage data to the liquidus configuration to estimate pressure should be avoided.

The relative sequence of "hornblende in" and "plagioclase in" phase changes in the melting interval of gabbro and tonalite is pressure sensitive (Wyllie, 1977) as well as composition dependent. Textural indications of the order of crystallization of these two minerals could yield at least qualitative information on the pressure, but much calibration remains to be done.

Mineral chemistry.—Use of the chemistry of mineral or minerals for pressure measurement holds the promise that we may have a means to determine pressure as a function of time, provided we can decipher the time, t, of formation of the minerals in question. The method is well-based for metamorphic rocks, and its application has become popularized by the "Gibbs method" advocated by Spear and his coworkers (Spear, Ferry, and Rumble, 1982; Spear and Rumble, 1986; Spear and Selverstone, 1983); for plutonic rocks, both the methods and their conceptual basis are just beginning to receive adquate attention (Young, Barth, and Anderson, 1986; see review by Essene, 1982). Major problems in applying the method to plutonic rocks include a concern for the ease of cation reequilibration at magmatic temperatures and a lack of assurance that the bulk chemistry of a given mineral represents the equilibrium value at *any* time. Unlike metamorphic rocks, for which an assumption of whole-rock equilibrium at the peak stage of metamorphism may be justified at least at medium to high grades of metamorphism, the chemistry of plutonic minerals must be regarded, as a first approximation, as time-averaged values.

Whitney and Stormer (1985) and Stormer and Whitney (1985) used a combination of two pairs of minerals, the Fe-Ti oxides whose equilibrium compositions are insensitive to pressure and the coexisting feldspars to infer the temperature and pressure of crystallization in the large magma chamber immediately preceding its eruption to form the

Tertiary Fish Canyon Tuff. The potential problems with this barometer are being debated (Grunder and Boden, 1987; Stormer, Whitney, and Dorais, 1987; Johnson and Rutherford, 1987a, b); the actual pressure may be significantly less (see discussion below, on the hornblende barometer). The compositions of the various minerals used may be too sensitive to subsolidus changes to form a generally useful geobarometer (see also Fuhrman and Lindsley, 1988).

Almandine-rich garnet, in equilibrium with biotite, muscovite, and plagioclase, was a barometer developed for metamorphic rocks (Ghent and Stout, 1981) but is useful for peraluminous granites (J. L. Anderson, 1989, written commun.). Magmatic garnets which are enriched in the grossular component ($Ca_3Al_2Si_3O_{12}$) and are principally a solid solution between grossular and almandine endmembers are encountered in numerous tonalitic plutons in southeast Alaska. Zen and Hammarstrom (1984b) summarized the experimental data on the stability of these grossular-rich almandinic garnets relative to a mixture of anorthite and pyroxene and suggested that these garnets crystallized at pressures possibly as high as 13 to 15 kb. Bohlen, Wall, and Boettcher (1983) and Bohlen and Liotta (1986; see also Bohlen and Lindsley, 1987) used the presence of grossular-almandine solid solutions, in combination with ilmenite-sphene equilibria rather than pyroxene equilibria, to establish comparable pressure ranges under granulitic metamorphic conditions. Even though application of this barometer to plutonic rocks still requires calibration, the presence of grossular-rich almandinic garnet suggests high pressure of equilibration.

Hammarstrom and Zen (1986) proposed that the total aluminum content of hornblende in specified mineral assemblages (2 feldspars, biotite, sphene, quartz, an iron-titanium oxide, and a former melt) could be used as a barometer for calc-alkalic plutonic rocks. The regression equation describing this barometer was confirmed by Hollister and others (1987); the two will be referred together as the HZH+ version of the hornblende barometer.

Using a variety of calc-alkalic starting material, Johnson and Rutherford (1989a) experimentally confirmed the covariant relation between pressure and the Al content of hornblende in the appropriate buffered assemblage. However, their regression equation yields a lower pressure for any given Al^T. Table 1 compares the results of the two regression equations in three applications. The pressure estimates based on the Johnson and Rutherford equation seem minimally adequate to accommodate the country-rock pressure data; part of the discrepancy might be explicable by time lag between metamorphic peaks and consolidation of the plutons in a setting of crustal uplift. As the HZH+ barometer gives pressure values consistent with those for pelitic contact metamorphic assemblages using several different barometers (M.L. Crawford, 1989, oral commun.), the discrepancy may be real and requires resolution even though the error bands about the various barometric values do overlap.

TABLE 1

Comparison of versions of the hornblende geobarometer

Pluton	Round Valley Idaho	Moth Bay Alaska	Ecstall British Columbia
Range of Al in hornblende*	2.57–2.62	1.78–2.66	2.17–2.28
Pressure range, HZH+**	9.0–9.3	5.0–9.5	7.0–7.6
Pressure range, Johnson and Rutherford***	7.5–7.7	4.1–7.8	5.8–6.2
Country rock pressure	8–9†	8††	8†††

*Hammarstrom and Zen, 1986.

**Hammarstrom and Zen, 1986: P (kb) = $5.03Al^T-3.92$. Note that use of this equation is not formally correct as the equation was partly based on the three data sets, set at P = 8 kb!

***Johnson and Rutherford, 1989a. P (kb) = $4.28Al^T-3.54$.

†Selverstone and others, 1987.

††Cook and others, 1987.

†††Crawford and others, 1987.

The above discussion highlights the point that in the HZH+ regression equation, the effect of temperature of equilibration was not explicitly taken into account. For rocks having the appropriate assemblage, this is not a major drawback because the pressure-dependence of the temperature range of crystallization, at pressures greater than about 2 kb, is small (see Hollister and others, 1987; Wyllie, 1977). Eventually, experimental determination of the temperature effect is needed to calibrate the barometer fully.

An interesting use of the hornblende geobarometer is to the Fish Canyon Tuff the hornblende of which contains a maximum of 8.2 wt pecent Al_2O_3, equivalent to about 1.4 Al^T per 23 oxygen formula (Whitney and Stormer, 1985), indicating a crystallization pressure of about 3 kb (Hammarstrom and Zen, 1986; Johnson and Rutherford, 1989b). It seems unlikely that the Al content of hornblende could have reequilibrated without at the same time changing the compositions of coexisting feldspars or the Fe and Ti oxide phases.

The hornblende geobarometer is conceptually akin to the Al-in-pyroxene barometer (Boyd, 1970, 1973), experimentally calibrated by MacGregor (1974). The barometer was designed to apply to mantle-derived ultramafic rocks. MacGregor (1974) determined the univariant four-phase boundary Al-pyroxene+spinel+pyrope+forsterite in the ternary system Al_2O_3-MgO-SiO_2. For this system, the SiO_2 activity was buffered by the assemblage rather than fixed by quartz, as is the case for the hornblende geobarometer.

Inclusions in minerals.—Both fluid and mineral inclusions in minerals have been tested as potential barometers. Pressure estimation based on fluid inclusions has been widely applied for metamorphic rocks (see Hollister and Crawford, 1981; Ferry and Burt, 1982, for summary of the method and examples of application). Pressures so estimated are almost always considerably lower than those based on the silicate phase assemblages and suggest either leakage or late trapping and retention of the fluid phase (Selverstone and others, 1984). The usefulness of fluid inclusions to estimate pressure for plutonic rocks remains to be shown.

Rosenfeld and his coworkers (Rosenfeld and Chase, 1961; Rosenfeld, 1969; Adams, Cohen, and Rosenfeld, 1975a, b; Cohen and Rosenfeld, 1979) suggested that when a mineral grain was being included in its host, the lattices of the two minerals must dimensionally fit. The fit may become poor at room conditions because the two minerals generally have different values of thermal expansion and bulk compressibility. However, if it can be shown that the differential strain across the grain contact has not been permanently relieved, for example by development of microcracks, then we should be able to estimate the P and T values during the growth of the mineral inclusion by measurement of the differential strain as a function of P and T. This may be done by measurement of null strain along the grain contact, indicated by optical properties, at different pressures and temperatures, and comparing the results with calibration curves to establish univariant "isomekes" ("equal lengths," that is, pressure-temperature trajectories of null strain; Adams, Cohen, and Rosenfeld, 1975a). The linear elastic constants can be directly utilized if inclusions and hosts have proper crystallographic orientations. If two such mineral pairs can be found in the same specimen, both the P and T during the inclusion process may be determinable. Although Rosenfeld's examples are metamorphic rocks, the method might apply to the much hotter plutonic rocks and deserves further scrutiny.

Argon spectra.—Discordant apparent mineral ages based on the argon release spectra have been used to estimate pressure. For example, if the biotite "age" is significantly younger than the hornblende "age" for the same rock, then it is inferred that the difference reflects the time it took to cool the rock from the Ar diffusion closing temperature for argon in hornblende, about 500°C, to that in biotite, about 300°C (Spear and Harrison, 1989; Harrison, Spear, and Heizler, 1989). Snee (ms.), using this approach, calculated the depth of intrusion for some plutons in the Pioneer batholith in southwest Montana.

Caution is necessary in the application of this method. The location of the sample with respect to the intrusive margin affects its cooling history, especially for large plutons. The age of intrusion must be separately established before the heat-conduction equation can be applied. The depth estimate is sensitive to the physical model assumed for the crust, including the mode of heat transport and the presence of concurrent decompression, as discussed by Snee (ms.). The closure temperatures for argon diffusion are subsolidus for all minerals; even a valid depth estimate therefore does not necessarily refer to that at the time of solidification. In the present state of knowledge of paleogeophysical properties for any plutonic region, it seems prudent to use the difference in argon "age" discordance between sites primarily to alert us to a possible pressure difference, but to use other means to measure it.

DISCUSSION OF PRESSURE GAUGES

Summary Critique of Existing Barometers

As petrologists, we would like to know not only the pressure of final consolidation of a pluton, but any other depth information during the evolution of a magmatic system, for instance, the depth or depths at which the magma was generated, and the depths at which the magma was pooled, differentiated, or reacted with the country rock. Most of the barometers discussed above measure the depth of pluton consolidation—the depth of death of the magma system.

Late-stage magmatic muscovite could eventually yield information on the depth of final consolidation; further, if the muscovite is a relic from an earlier phase of crystallization, additional, earlier thermal and depth history might be extractable. The same can be said of magmatic epidote. Phenocrysts of epidote do occur in fine-grained dike rocks and indicate a polybaric history (Evans and Vance, 1987; Dawes, 1988; Dawes and Evans, 1988). As a proper buffering mineral assemblage is lacking, however, such epidote phenocrysts do not convey much more *depth* information than does the presence of high-pressure crustal xenoliths (Evans and Vance, 1987).

The fit of bulk modal or normative data to the haplogranite liquidus surface does not convey meaningful information at *any* pressure except for granite truly in the chemical system Ab–Or–Q, as discussed previously.

The use of fluid inclusions or of argon release spectra best yield pressure estimates during cooling of a pluton and conveys no information on previous pressure history.

The use of mineral chemistry of coexisting phases suffers the handicap that, away from the rims of minerals, the nature of the buffering assemblage is commonly unknown even if there is no post-consolidation alteration.

Thus at the present stage of our understanding of plutonic rocks, the barometers apply mainly to depths of consolidation; the method of solid-inclusion barometry is one possible exception. How, then, can we hope to reconstruct the history of magmatic systems?

What are the Desiderata of a Barometer?

A good mineral-chemistry barometer should be sensitive to pressure changes; it should be applicable to common rock types; it should also be *stable* against subsequent alterations. Thus for a barometer based on mineral chemistry, the atomic species and their crystallographic sites should be such as to have very low solid-state diffusion coefficients, should be insensitive to redox conditions, and should be resistant to hydrothermal alteration.

A mineral-chemistry based barometer should encode temperature information for the sample, so that if the barometer does depend on temperature, the effect of this variable may be compensated for. The temperature effect ideally should be small; in other words, a multiphase reaction serving as a good mineral-chemistry barometer should have a large volume change relative to enthalpy change. A mineral-chemistry based barometer should also encode time information, for example, by chemical zoning or by zonal mineral inclusions, so that the P–T–t history may be reconstructed. Thus, species that readily respond to post-consolidation events such as ferrous iron, calcium, sodium, potassium, hydroxyl, and halogen should be avoided for barometric purposes. The alkalis, the halogens, and hydroxyls are also all components whose values are subject to large analytical and computational errors. On the other hand, components such as silica, alumina, and titania should be explored for their barometric promises.

Some minerals can be expected to show strong compositional zoning, reflecting changes of chemical activity in the magma chamber during crystallization. If these minerals do not show such zoning, later annealing may have occurred. Thus potassic feldspar and biotite are possibly troublesome candidate barometers, but plagioclase, epidote, hornblende, pyroxene, and garnet, all of which do show zoning, are more promising on this count.

Biotite deserves special comment. It has been used in numerous barometers and thermometers for metamorphic rocks, principally by exploiting its Fe″/Mg ratios. The lack of Fe-Mg zoning in many igneous biotites, on the scale of microprobe analysis, leads one to suspect that these biotite compositions were changed by post-magmatic homogenization. Magmatic biotite, however, does show different total aluminum contents that in part reflect the degree of aluminum saturation in the magma as well as the nature of the AFM phase assemblages (Zen, 1986). The Al content of biotite may be useful for barometry; it is ironic that attempts to extract information on the intensive variables from biotite compositions commonly use the less stable compositions whereas the readily determined Al^T is neglected.

Likewise for hornblende, the Al (and probably Ti) contents are probably much more resistant against post-crystallization changes than are Fe″/Mg ratios or the alkali contents. Use of concentrations of various elements in amphibole of metamorphosed mafic rocks as barometers and some of the pitfalls are discussed by Laird (1982) and Laird and Albee (1981). Arguments against using the alkali contents of the mineral, especially using the A-site (alkali) occupancy to monitor the intensive variables, include the fact that A-site occupancy assignment is made at the end of the computational scheme, so that errors in chemical analyses of all the elements, or in their oxidation states, or in the assumed substitution schemes (including possible coupled vacancy substitutions in the M(1-3) sites such as $2Al^{vi} = 3(Fe, Mg)$) are reflected in the amount and nature of the A-site occupancy. In contrast, data on the diffusion coefficient at magmatic temperatures of Al in diopside (Freer, 1981; Sauttner, Jaoul, and Abel, 1988), in which it occupies crystallographical positions analogous to those in hornblende, indicate that redistribution of Al in hornblende is likely a negligible factor (see Hammarstrom and Zen, 1986 for further discussion).

Diffusion of species in garnet and diffusion of Al in the coupled substitution $M'' + Al = M' + Si$ in plagioclase are probably sufficiently low that these minerals may be reasonable phases for barometric purposes. Experimental data on the distribution of these components in coexisting phases over the appropriate P, T ranges and in the proper assemblages are eagerly awaited.

FROM PRESSURE ESTIMATES TO PALEOGEOPHYSICAL RECONSTRUCTION

The pressure of consolidation of plutonic rocks is useful for geological reconstruction. Even more useful, however, is the pressure history (P-t record) that could allow us to obtain other kinds of geological and geophysical information. I will discuss some of these pies-in-the-sky in this section.

An obvious use of P-t data is to reconstruct the average decompression rates between time intervals that bracket the pressure data. Even if P-t data are not available for a single pluton, composites of P-t tracks for a series of dated plutons from the same tectonic block could give useful results. An example of this approach was given by Zen (1988c; see also Anderson, Barth, and Young, 1988); it is being currently applied by us to the study of decompression history of southeast Alaska.

Reliable pressure estimates in conjunction with temperature of a locally derived (possibly migmatitic) magma could permit calculation of the *average* vertical component of the pressure-temperature gradient, $dP/dT|_z$, between the earth surface and the pluton at the time of crystallization. Values of $dP/dT|_z$ in turn would provide the starting point to construct the paleothermal profile.

Knowledge of the pressures of plutons could permit extrapolation of the thermal structure of the crust below the present level of erosion and thereby improve estimates of the depth of generation of the magma (Zen, 1988a), assuming conductive heat transfer. This effort relies on thermal modelling and requires two kinds of model-based estimates: first, to combine the model variables to create a *metamorphic field gradient* that is consonant with the observed data; second, to extend the model parameters below the exposed pluton in order to estimate the minimum depth for the generation of the magma.

The *metamorphic field gradient* describes the apparent spatial rates of change of the metamorphic temperatures and pressures for adjacent rock bodies, provided they are not separated by faults. It differs from the true thermal gradients, dP/dT, for a variety of reasons, including: (1) each packet of rocks probably followed a different P-T-t trajectory, (2) the metamorphic mineral assemblages and especially the chemistry of coexisting minerals may have become frozen in at different times and under different P-T conditions, (3) the attainment of maximum T and maximum P probably was not coeval for any one body of rock because pressure change by loading and unloading is instantaneous but the thermal effects are delayed, and (4) the maximum P and/or T values for *different* bodies of rocks were probably not attained at the same time. *Metamorphic field gradient* is what we observe; each point along the gradient commonly records a different t value. Thermal gradients, by contrast, are what we speculate on and, in desperation, try to model.

Reconstruction by modelling of the observed *metamorphic field gradient* requires some consistent assumption on the meaning of the observed P-T values for the rock bodies. If, for instance, we deduce that the rocks preserve maximum values of T, then the *metamorphic field gradient* becomes the "metamorphic geotherm" of England and Richardson (1977; see also Spear and Selverstone, 1983) and is identified with the loci of maximum T values of P–T–t paths for a crustal section undergoing vertical motion (fig. 1). Different modelling parameters can then be tested against the observed metamorphic geotherms, within limits of tolerances. A model of P–T–t paths not only obeys the boundary conditions (for example, the P-T trajectory must pass through the surface at T = 0°C, P = 0 kb) but also contains information of the time, t, at which each point on the geotherm was reached. Thus the model contains implicit information on the instantaneous thermal gradient, dP/dT. This information *must be consistent with the thermal and barometric information provided by the plutons.* Such consistency would in turn allow estimation of other physical parameters.

I will develop this idea by a pair of geological models simplified from real situations. Model 1 (fig. 2A) shows a tonalitic pluton of known age, 100 Ma. From isotopic or other kinds of evidence, we have concluded that the magma had a sialic crustal source. We have further inferred that during the final consolidation of the pluton, both it and its surrounding country rocks were at the same temperature (say 760°C) and pressure (say 9 kb; or 27 km of crust). Let us suppose that the contact relations of the pluton and the nature of the metamorphic country rock were such that the pluton was most likely derived by partial melting of the local country rock (an actual example is the high-pressure, magmatic epidote-bearing Round Valley pluton of western Idaho discussed by Zen and Hammarstrom, 1984a; Zen, 1988c; Hammarstrom and Zen, 1988. See also Manduca, 1988). We then conclude that: (1) the average geothermal gradient at the time of plutonism was about 28°/km, (2) the observed geotherm was regional rather than imposed by the intrusion, (3) the metamorphic geotherm for similar country rocks away from the pluton must have been no more than this value, (4) the crust was at least 27 km thick at the time of intrusion, and (5) the thermal structure of the crust was such that the metamorphic geotherm must have passed through the P–T–t point of 9 kb, 760°C, 100 Ma. These constraints significantly help in selecting plausible models of the thermal evolution of the crust.

However, the model contains no information on the thickness of the crust that existed below the level of the pluton, nor on the thickness

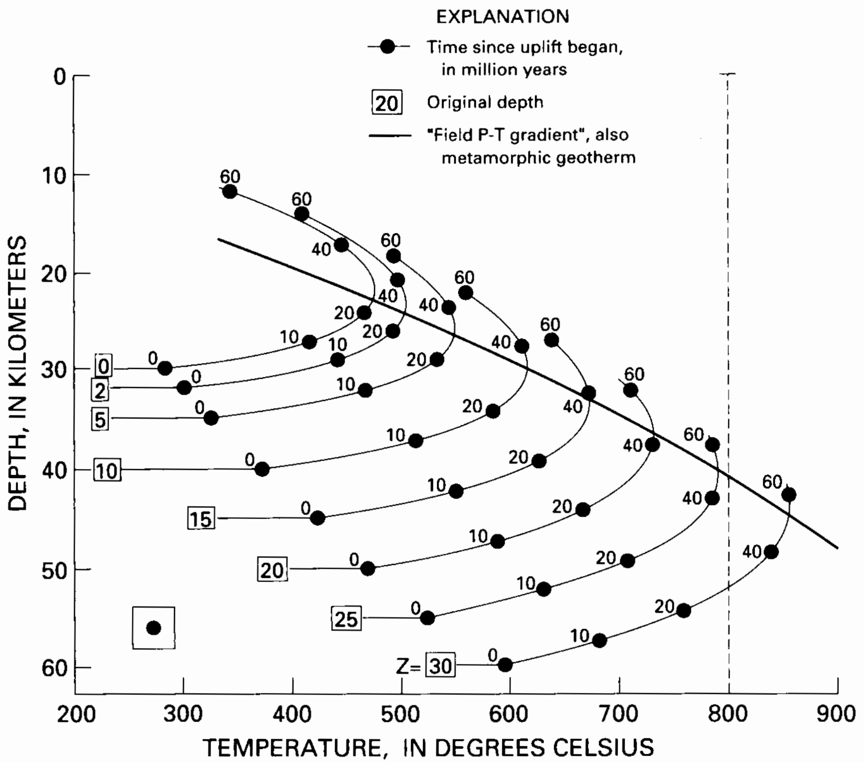

Fig. 1. Thermal model calculation of a thrust-thickened sialic crust, showing the diachronous metamorphic geotherm defined as the locus of points of maximum temperatures, and P–T–t paths for material originally located at selected levels of the crust ("stratigraphic levels" of Zen, 1988d). The original "Z" depth values are given in boxes at the left end of each uplift path. The crust originally situated at 2 km attained a maximum temperature of 500°C and therefore was the model depth to which the pluton of figure 2B had intruded and now define the erosion surface. Rocks between 2 and 15 km depth never attained a temperature of 650°C during the thermal event and thus could have produced highly leucocratic grantic melts at a later time. Box at lower left shows typical measurement uncertainties in geobarometry and geothermometry.

of the crust removed by erosion or tectonic denudation prior to plutonism at t = 100 Ma. This thickness could be appreciable because the time lapses between melt generation and the triggering tectonic event and between pluton consolidation and melt generation are unknown (Zen, 1988d). During these intervals, crustal movements might have induced considerable vertical denudation.

The permissible set of inferences, however, would be quite different for the geological conditions of model 2 (fig. 2B). The physical parameters are chosen to be similar to that of model 1, except that the country rock is not at grade with the pluton. I set the temperature at 760°C, pressure at 8 kb for the consolidation of the pluton at t = 77 Ma,

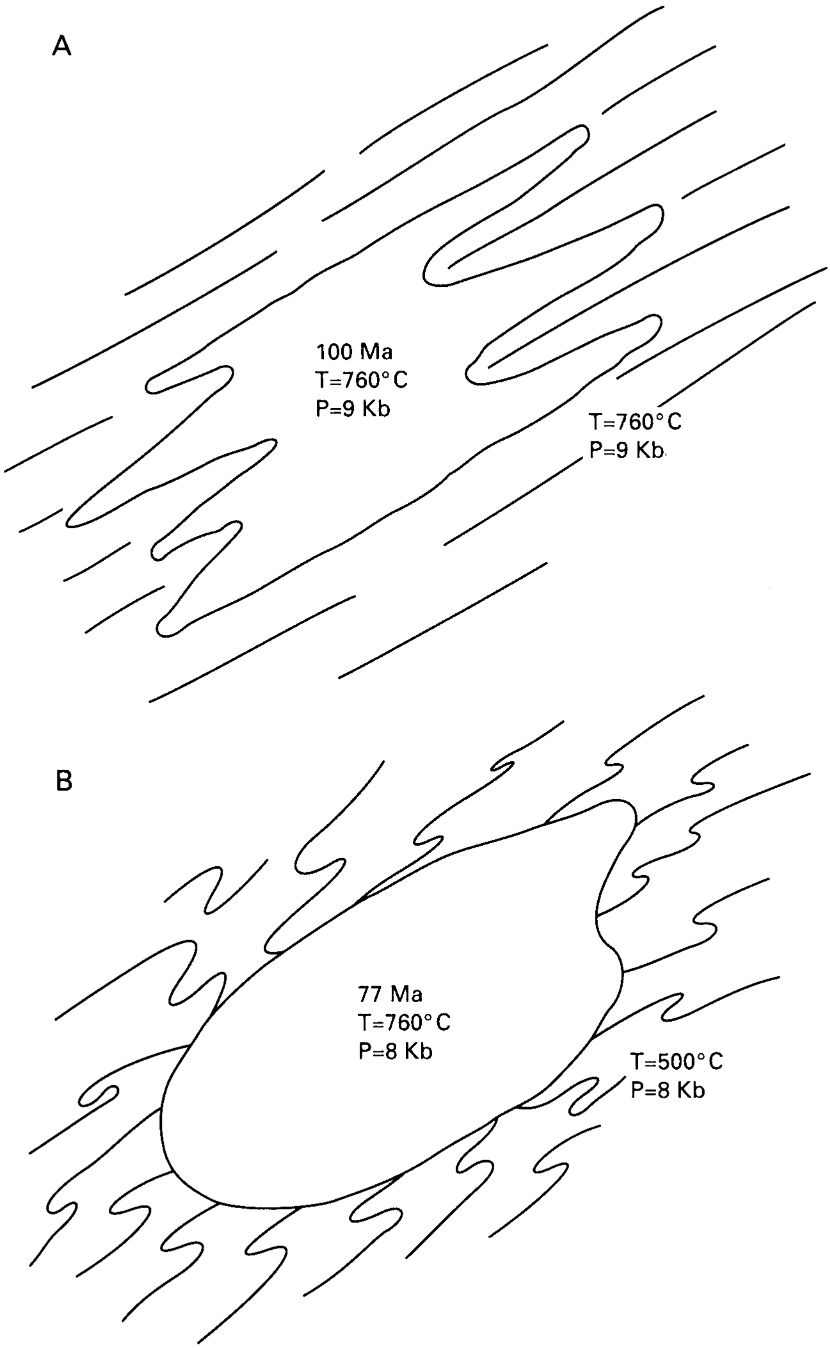

Fig. 2. Schematic examples of two tonalites showing different contact relations and temperature-pressure contrasts with the country rock. (A) Tonalite was produced by *in situ* anatexis; (B) tonalite was intrusive, and its magma was generated at greater depth.

but for the metamorphic country rock, I set T = 500°C, P = 8 kb, so that the pre-intrusion regional thermal gradient was about 20°C/km (except for the age, this example approximates the high-pressure, magmatic epidote-bearing Moth Bay pluton of southeast Alaska; Arth, Barker, and Stern, 1988; Zen and Hammarstrom, 1984a). These metamorphic conditions were attained at a time t′ that may be different from the plutonic age; however, all these parametric values can at least in theory be independently determined. The large discrepancy between the temperatures of the pluton and country rock shows that the magma was not derived locally. A thermal model for the area, then, has three constraints: First, the metamorphic geotherm must satisfy the country rock information (T = 500°C; P′ = 8 kb; t′), as well as the boundary condition (T = 0°C; P = 0 kb; all t's). Second, it must reach the temperature for tonalitic magma to form, say at 800°C, at some geologically reasonable pressure, P″ > P′. Third, melting at P″ should start at t > 77 Ma, and the affected crust must have resided in that temperature range long enough to generate sufficient melt, with allowance for the effect of latent heat (Zen, 1988d), to account for the observed size of the pluton. The difference between the depth of country rock metamorphism, P′, and the depth of anatexis, P″, then gives the *minimum* thickness of the additional continental crust that must have existed below the exposed pluton.

Figure 3 gives an example of the procedure I used to estimate P″. For any given set of model parameters, I tested three variables that sensitively affect the thermal structure of the model crust: the thermal diffusivity K, the decompression rate u, and the subcrustal ("mantle") thermal flux, Q. The criteria are that the metamorphic geotherm passed through the value given by the country rock, and that at some reasonable depth below the exposed level, the temperature becomes magmatic (taking into account latent heat and residence time). These criteria rule out many models. Each surviving candidate model provides a different estimate of the value of P″ or the depth of the continental crust present at the time of magma genesis; each model can also provide information on the thickness of the crust removed both *before* magma generation and *during* the time interval between peak metamorphism of the country rock and the intrusion. The validity of each model must be judged on other merits. Inconsistencies among uplift rate and plutonic age and depth could eliminate candidates effectively. For example, if the pluton has a measured consolidation age of 100 Ma but a model-predicted age of 77 Ma, this difference could mean either that the model is wrong or that the land surface is inherited from the Miocene. Figure 4 gives a graphical account of the disposition of crusts for a particular model and allows a summation of the minimum thickness of the crust at the model site.

If we know the depth of generation of a plutonic rock (for example, a tonalite) which is not a petrologic residual system, then if the same area is later invaded by another, lower temperature pluton (for example, a

u \ Q	30	40	50
0.1	G Y M ≈Y	G Y M Y	G Y M Y
0.3	G N M N	G Y M N	G Y M Y
0.6	G N M N	G Y M N	G Y M ≈Y
1.0	G N M N	G Y M N	G Y M N

Fig. 3. Example of screening out inappropriate model parameters to satisfy "observed" field conditions for figure 2B. Q, mantle heat flux, in mW/m^2; u, uplift rate, in km/my (mm/yr); *G*, condition that specific metamorphic geotherm passing through 500°C and 9 kb; *M*, condition that at some reasonable greater depth the temperature attained 800°C; *Y*, condition is satisfied; *N*, condition is not satisfied. Only those results where both the *G* and *M* conditions are *Y* are acceptable. The example of Q = 50 and u = 0.3 is given in figure 4 and also discussed under figure 1. Other model parameters: Rock density, 2,780 kg/m^3; heat capacity, 0.9 kJ/kg.K; thermal conductivity, 3.0 W/m.K (thermal diffusivity, $1.2 \times 10^{-6} m^2/s$). Radiogenic heat production (upper 15 km of original crust and thrust material), 1.5×10^{-6}W/m. Incubation period, 5 my; five thrust sheets, each 6 km, arriving at 0.5 my intervals. See Zen (1988d) for choice of parameters and their numerical values.

peraluminous granite rich in incompatible elements showing evidence for a hydrous vapor phase), we may make additional inferences. For model 2, part of the original crust between the highest possible level of generation of the tonalitic melt and the present level of erosion remained at a temperature below that needed for the generation of a peraluminous hydrous melt (in fig. 1, depth levels between Z = 2 km and Z = 15 km, for melt generation at 650°C), so the younger magma could be caused by a second heating event within the same crust. However, for model 1, the entire crust as far up as the present level of erosion was hot enough to generate tonalitic magma, so it could not have given rise to a low-temperature magma later. This granite would have to be either the result of remelting and fractionation of the original tonalite or, if that should prove impossible (for example, on the basis of inconsistent isotopic signatures), would require that a new sialic source was tapped, possibly from a relatively undepleted crust accreted since the first thermal event.

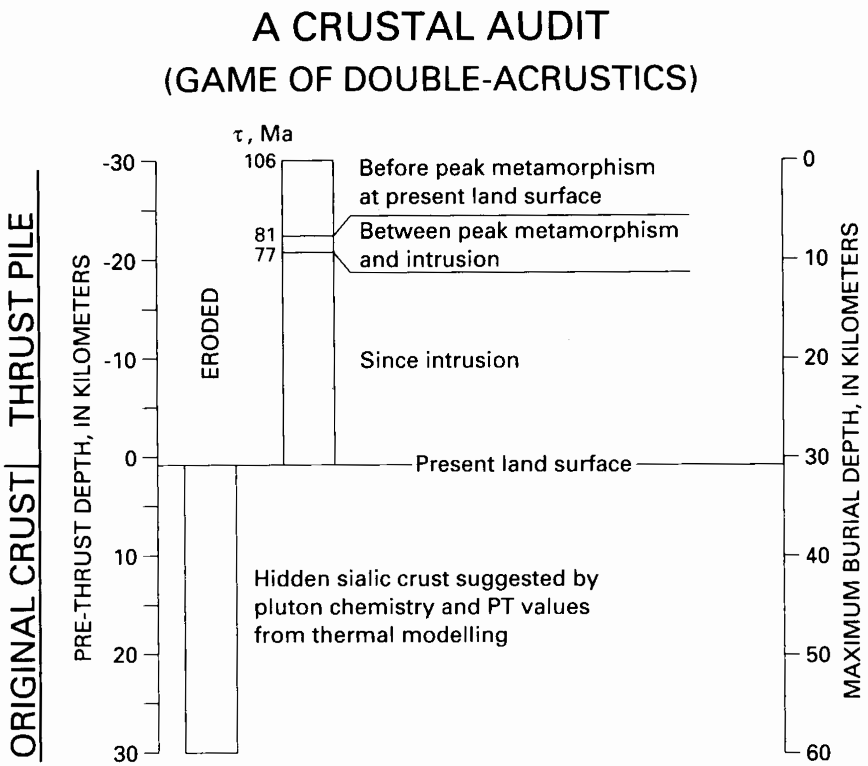

Fig. 4. A crustal audit for the situation shown in figure 2B, based on one-dimensional thermal model using specific model parameters given in figure 3, with the uplift rate, u, of 0.3 km/my and the mantle heat flux, Q, of 50 mW/m^{-2}. The time, t, at which the land surface was reduced by erosion to the specified level is indicated.

Pressure values for magma generated in the sialic component of a downgoing subducted slab obviously could not provide a meaningful measure of the thickness of the crust, but how would a geologist, looking at the resulting pluton millions of years later, distinguish a shallow subduction couple from an extremely thick hanging-wall contineta crust? Perhaps significant contamination by the mantle wedge is a key, but chemical evidence of contamination by mantle material in the pluton could indicate material supplied from below, concomitant with mantle heat transfer. These cases must be distinguished.

If we have reliable means to read the pressure history of a group of plutons from the same tectonic block, we would be better equipped to determine the crustal accretion and decompression history. The thickness and thermal state of the crust must have strongly affected the physical processes at plate margins or at incipient rifts; yet independent methods to estimate thicknesses are nearly nonexistent for crusts that have undergone later reworking. The Moho depths for both the Great Basin area (for example, Hauser and others, 1987; Allmendinger and others, 1987) and the northern Appalachians (Stewart, 1989) are now known to be very late tectonic developments, so we are not justified in assuming that the present crustal thickness has ancient relevance. Plutonic rocks might provide a way to fill these important gaps in crustal data.

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