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Two feldspar and iron-titanium oxide equilibria in silicic magmas and the depth of origin of large volume ash-flow tuffs

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Abstract

For a number of large volume ash-flow tuffs, the temperatures from two-feldspar geothermometry are consistently lower than temperatures obtained by iron-titanium oxide geothermometry by as much as 150°C (assuming low pressure). The iron-titanium oxide temperature is essentially independent of pressure, but the two-feldspar temperature has a pressure dependence of 18°C/kbar. The pressure for which the calculated temperatures correspond should indicate the depth of origin of the phenocryst assemblage. Although the present experimental calibration of these geothermometers is uncertain, other types of data suggest that useful, qualitative estimates of the pressure and depth of origin for certain ash-flow tuffs can be made. The phenocryst assemblage in the Fish Canyon Tuff (>3000 km^3, San Juan volcanic field) appears to have originated at depths of about 25 km. The phenocrysts in the Bishop Tuff (500 km^3, eastern California) appear to have originated at about 15 km depth. Even with allowance for a large error in this determination, it suggests that ash-flow magmas may develop at much deeper levels in the crust than previously thought. However, deeper sources would be consistent with some experimental work and geophysical measurements on silicic volcanic centers. This suggests that some, but not necessarily all, large volume ash-flow tuffs may have developed their chemical characteristics at mid-crustal levels. Models for the differentiation and eruption of such magmas should consider the possibility that the magmas may not have resided in shallow chambers for any significant period of time.

Introduction

In recent years there has been considerable interest in large volume ash-flow tuffs and their relationship to the origin of granitic magmas (Smith, 1979; Hildreth, 1981). Ash-flow tuffs representing magma volumes in excess of 100 km^3 are not uncommon and some exceed 3000 km^3. Magma volumes the size of batholiths have erupted, leaving calderas 20 to 50 km in diameter. These tuffs range in composition from quartz latite or rhyodacite to high-silica rhyolite. Many are very conspicuously zoned and are thought to represent the products of coherent eruption, from the top down, of compositionally stratified magma chambers (Smith, 1979). A few large ash-flow tuffs seem to represent eruption of essentially homogeneous magma (Lipman, 1975, p. 47–49; O'Leary, 1981; Whitney and Stormer, unpublished work).

The origin of these magmas and the mechanisms by which they become zoned are the subject of considerable study and debate (Hildreth, 1981; 1983; Michael, 1983a, 1983b). For a few of these tuffs, there are extensive geochemical and mineralogical data including estimates of temperature and the fugacities of volatile components. Using these data, very sophisticated quantitative models have been applied to assess the mechanisms of differentiation. However, the absolute pressure (or depth of origin) is very poorly constrained and remains a major uncertainty. The depth of origin of the magma is also of critical importance in modeling the thermal regime and the mechanics of caldera formation.

This paper presents a possible method for determining the absolute pressure and depth of origin for the magmatic mineral assemblage of plagioclase, sanidine, magnetite, and ilmenite, which is commonly found in silicic ash-flow tuffs. Currently this method suffers from the uncertain experimental calibration of two-feldspar and Fe–Ti oxide equilibria. However, data from natural mineral assemblages show a remarkable consistency which suggests

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that the method is potentially very useful and, further, that the sources of at least some large volume ash-flow tuffs are rather deep.

This work is an outgrowth of a continuing mineralogical and geochemical study of the ash flows of the central San Juan volcanic field. The ideas presented here were first developed during consideration of feldspar and Fe–Ti oxide geothermometry in the Fish Canyon Tuff. The Fish Canyon Tuff is the oldest and largest (>3000 km^3) of a series of 5 ash-flow tuffs erupted from nested calderas in the central San Juan field between 28 and 26 m.y. ago (Steven and Lipman, 1976 and Lipman et al., 1978). Lipman et al. (1978) divided these tuffs into two groups based on characteristics which suggested relatively high or low pressure origins. The Fish Canyon Tuff is in their high pressure group. It is a phenocryst-rich, quartz latite tuff with about 67% SiO_2. There appears to have been little or no compositional zoning in the magma (major elements, trace elements, mineral composition; Whitney and Stormer, unpublished work). The phenocryst assemblage consists of plagioclase, sanidine, biotite, hornblende, quartz, sphene, magnetite, apatite, and ilmenite (in order of decreasing abundance). Additional geochemical and mineralogical data for the Fish Canyon Tuff is presented by O'Leary (1981).

The mineralogy and geochemistry of the Bishop Tuff was studied in detail by Hildreth (1977, 1979). It is an excellent example of the relationship between oxide and feldspar minerals since there are abundant data available for coexisting iron-titanium oxides and two feldspars. The Bishop Tuff was erupted about 700,000 years ago from the Long Valley Caldera on the east front of the Sierra Nevada, California. This eruption may have involved as much as 600 km^3 of magma as shown in a study of the geology, structure, and geochronololgy of the tuff and the Long Valley caldera presented by Bailey et al. (1976). As a result of the potential for geothermal energy and volcanic hazards in the area, there are a number of other geological and geophysical studies available (Long Valley Symposium, 1976, *Journal of Geophysical Research*, v. 81, p. 721–860, and others discussed below). The Bishop Tuff is a high-silica rhyolite (74–76% SiO_2) which shows a strong chemical zonation that correlates with magma temperatures ranging from 720 to 790°C. This chemical and thermal zonation is thought to represent a stratification in the precaldera magma chamber. Plagioclase, sanidine, quartz, magnetite, ilmenite, and biotite are found throughout the tuff. Allanite is found only in lower temperature samples and pyroxene and pyrrhotite in higher temperature samples.

Geothermometry and the effect of pressure

The Fish Canyon Tuff contains magnetite with 10–15 mole% of ülvospinel in solid solution and ilmenite with about 15–25 mole% hematite in solid solution (O'Leary, 1981; Whitney and Stormer, unpublished work). Although the Fish Canyon Tuff is 28 m.y. old and in large part devitrified, samples can be selected which show little devitrification and contain homogeneous, unaltered oxides especially in the lower units. Oxides included in phenocrysts were often well preserved, and have essentially the same compositions as those in the groundmass. In a few cases magnetites with inhomogeneity due to incipient oxidation exsolution were analyzed by "point counting" up to 50 spot analyses per grain. Altered oxide grains could be identified by anomalous distribution of Mn and Mg between magnetite and ilmenite. The analyses used for Figure 1 have ratios of Mn(Mag./Ilm.) that are within the range of published analyses in the Bishop Tuff (Hildreth, 1977) and Paintbrush and Timber Mt. tuffs of southern Nevada (Lipman, 1971). Mg distributions are more scattered, but most also correspond to the range found in the Bishop, Paintbrush, and Timber Mt. tuffs. After recalculating the analyses as recommended by Stormer (1983), Spencer and Lindsley's (1981) geothermometer was used to obtain the temperature of equilibration. The temperatures, shown in Figure 1, average about 800°C, and range over about 30 degrees in the lower parts

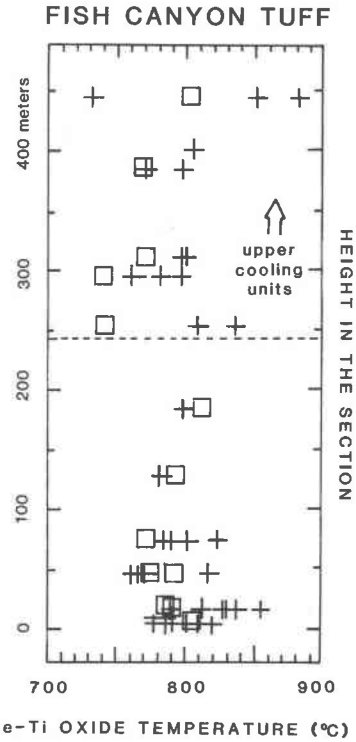

Fig. 1. Iron–titanium oxide temperatures for a representative section through the Fish Canyon Tuff plotted relative to their actual stratigraphic height above the base of the tuff. Crosses—electron microprobe analyses (Whitney and Stormer, unpublished work). Boxes from O'Leary (1981). Some of O'Leary's data use bulk chemical analysis of exsolved magnetite separates. Upper cooling units show petrographic evidence of lower-pressure partial reequilibration.

of the section. In the upper part of the section there is more scatter (probably reflecting greater alteration of the oxides in those samples) and, perhaps, a tendency toward lower temperatures. There is no obvious correlation between temperature and height in the section, however.

The feldspars in the Fish Canyon Tuff are remarkably homogeneous. The plagioclase composition averages about $Ab_{65}An_{30}Or_5$ with a total range of variation less than 5 mole% Ab. The alkali feldspar phenocrysts have compositions of $Ab_{24}Or_{74}An_1$ with 1–2% celsian ($BaAl_2Si_2O_8$) and a range of variation less than 3 mole% Ab. Carefully measured optical axial angle and X-ray peak positions (Wright, 1968) indicate that the alkali feldspar phenocrysts have an appropriate Al–Si disorder for sanidine of this composition at temperatures of about 800°C. The distribution of the albite component between the coexisting plagioclase and sanidine yields the temperatures shown in Figure 2. The potassium contents of the plagioclases are consistent with equilibrium at these temperatures, and the coexisting feldspars meet the other criteria for equilibrium suggested by Brown and Parsons (1981). However, the temperatures are almost 150 degrees lower than the Fe–Ti oxide temperatures. In addition the feldspar temperatures from the upper part of the section indicate higher temperatures, whereas the Fe–Ti oxides indicate the same or lower temperatures.

The iron-titanium oxide temperatures in the Bishop Tuff range from about 720 to 790°C (Hildreth, 1977, using Carmichael, 1967, and Buddington and Lindsley, 1964). The early-erupted parts of the tuff are lower in temperature than the later parts. The effect of recalculating Hildreth's (1977) analyses using Stormer (1983) and Spencer and Lindsley (1981) is to raise the temperature by an amount varying from 10°C for lower temperature samples to about 40°C for the highest temperature samples. On this basis the temperatures range from 730°C to 830°C.

The plagioclase compositions range from An_{14} to An_{23} and are correlated with temperature, but sanidine compositions are more constant between about Ab_{36} and Ab_{32} (Hildreth, 1979, Figs. 6 and 7). Hildreth (1979) commented that the feldspar temperatures in the Bishop Tuff were "50 to 70 degrees lower than the iron-titanium oxide temperatures" (calculated using the Stormer, 1975, geothermometer, presumably at a relatively low pressure). About the same difference is calculated using Brown and Parsons (1981, Fig. 2) at one kilobar pressure.

If the effects of pressure are considered, the discrepancies between the geothermometers can be reconciled. The Fe–Ti oxide geothermometer is known to be nearly independent of pressure up to at least 10 kilobars (Rumble, 1969). For the two-feldspar geothermometer there is a very significant pressure effect as shown by Stormer (1975). Anomalous two-feldspar temperatures from granites were explained as a result of changing pressure (Whitney and Stormer, 1977). According to Brown and Parsons (1981) the temperature increases with pressure by about 18°C per kilobar, and values from other work are similar (Stormer, 1975). This is a very substantial pressure coefficient, and the two-feldspar "geothermometer" may be useful as a "geobarometer" when combined with a pressure-independent estimate of temperature.

The *P–T* relationship between the two-feldspar and Fe–Ti oxide geothermometers is shown in Figure 3. The pressure at which the assemblage could have coexisted is given graphically by the point at which the lines representing the feldspar and the oxide equilibria intersect. If

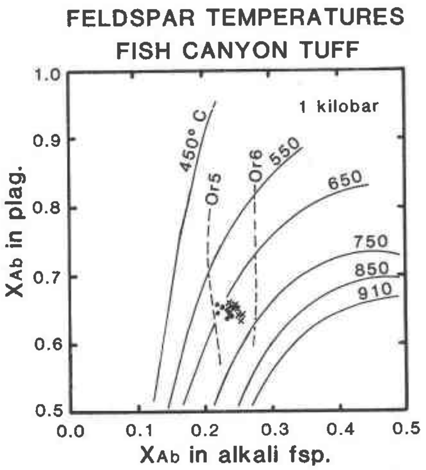

Fig. 2. Apparent 1-kilobar two-feldspar temperatures for the Fish Canyon Tuff. Circles are feldspar compositions from lower cooling units; crosses are feldspar compositions from upper cooling units. Isotherms and Or ($KAlSi_3O_8$ mole percent) contents of plagioclase from Brown and Parsons (1981, Fig. 2); feldspar data from Whitney and Stormer (unpublished work).

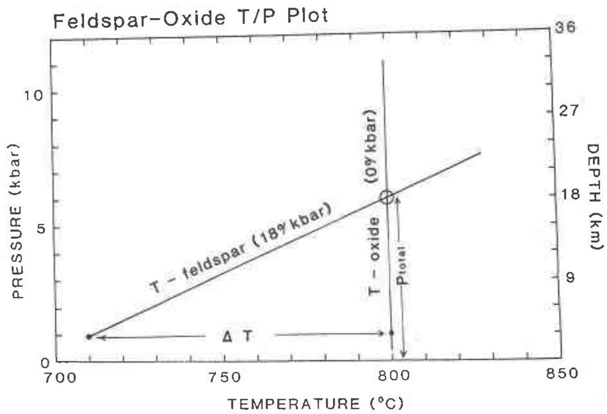

Fig. 3. The effect of pressure on the Fe-Ti oxide and two-feldspar geothermometers. The Fe-Ti oxide geothermometer is esentially independent of pressure. The two-feldspar temperature, however, has a slope of 18°C per kilobar. The intersection between the two-feldspar and Fe–Ti oxide temperature lines gives the pressure. The difference between the two temperatures determined for 1 kilobar (ΔT) is directly related to the total pressure of mutual equilibration by equation 1.

the feldspar temperature is determined for a pressure of 1 kilobar (i.e., using Brown and Parsons, 1981, Fig. 2), then the difference between the Fe–Ti oxide and 1 kilobar two-feldspar temperature is directly related to the equilibrium pressure for the assemblage by the following equation:

$$P(\text{kbar}) = 1 + \{T(\text{oxide}) - T(\text{feld., 1 kbar})\}/18 \quad (1)$$

Accuracy of the method

The accuracy of the pressure calculation is obviously dependent on the accuracy with which the apparent temperatures can be determined. With the presently available experimental data it is not possible to make a meaningful assessment of the uncertainty arising from either of the geothermometers. Stormer (1983) has shown that there are significant discrepancies between the Buddington and Lindsley (1964) and the Spencer and Lindsley (1981) Fe–Ti oxide geothermometers which could possibly be a result of problems in Spencer and Lindsley's thermodynamic model. The amount of the discrepancy varies and, fortunately, there seems to be the least amount of discrepancy at the temperatures and oxygen fugacities determined for the tuffs considered here.

The effects of minor components in the oxides are essentially unknown, but appropriate recalculation methods (Stormer, 1983) probably reduce this error to within 10 degrees, at least for these tuffs. Hildreth's (1977) analyses include every element which could reasonably be expected, and we have recalculated all his Bishop Tuff data, but recalculating using Stormer (1983) changed temperatures less than 5 degrees in almost all cases. Our quantitative analyses of the oxides in the Fish Canyon Tuff included every element except V and Nb. Energy dispersive spectra showed that there was no significant amount of Nb ($\ll$0.5% Nb_2O_5). V analyses are complicated by overlaping Ti peaks, but given the low TiO_2 contents of the magnetites, V_2O_3 contents as much as 0.5% should have been detectable. In any case, a hypothetical calculation with 1.8% V_2O_3 added to a typical Fish Canyon magnetite composition produced a shift of only 7°C (equivalent to less than 1/2 kbar pressure).

Hildreth (1979, p. 48) made an assessment of the factors affecting the precision and accuracy of his Fe–Ti oxide temperatures for the Bishop Tuff and concluded that they "may be accurate to ±10°C" (equal to a precision of about 1/2 kilobar in Fig. 4). We believe that this is a reasonable estimate of precision for selected fresh samples given careful petrography and analytical technique although the accuracy of the absolute temperature is likely to be worse.

The accuracy of feldspar temperatures is even less easily assessed. Feldspars are reasonably easy to analyze using the microprobe, and their stoichiometry affords a good check on the quality of individual analyses. For relatively homogeneous samples like the Fish Canyon and Bishop tuffs the analytical uncertainty should be less

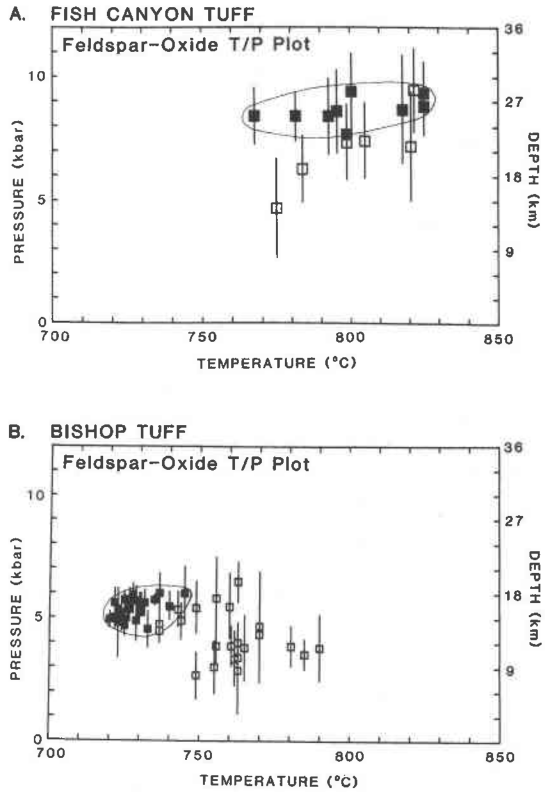

Fig. 4. Temperature and pressure plots obtained using the feldspar–oxide method. A. Samples from the Fish Canyon Tuff (analyses from Whitney and Stormer, unpublished work). Solid squares represent samples from the lower, early part of the tuff; open squares represent samples from the upper, later part of the tuff. The bars represent the maximum range of pressure calculated using the total range of variation of analyses of feldspar in the sample. The uncertainty due to variation in oxide composition is smaller. B. Samples from the Bishop Tuff (Hildreth, 1977). The solid squares represent early, nonpyroxene-bearing samples; open squares represent later, pyroxene-bearing samples. The circled area contains 45 of 60 analyses for which Hildreth (1977) has given oxide and feldspar analyses. All analyses outside the circle are from samples erupted late in the sequence. The bars have the same significance as in A. The range of variation within each sample for the oxide analyses was not given by Hildreth (1977).

than 2 mole% Ab. Depending on the particular composition this should allow temperature to be estimated to within 30–50°C. The results of our calculations for a large number of low temperature Bishop Tuff samples show that the precision can be much better and the reproducibility of the pressure estimate could be within 1–2 kilobars.

Since the Stormer (1975) geothermometer was developed there have been a number of attempts to improve a thermodynamic model for a two-feldspar geothermometer. All of these (including Haselton et al., 1983, which is the model of Powell and Powell, 1977, with new data) are fundamentally limited to dealing with data from the Ab–Or and Ab–An binary systems. The problems with these approaches were reviewed by Brown and Parsons (1981) and a graphical model based on ternary experimental data (Seck, 1971) was presented. M. Ghiorso (Univ. of Washington, unpublished manuscript) has recently developed a thermodynamic model for the feldspars including ternary parameters which were also derived by fitting Seck's (1971) data. For the compositions of feldspars considered here the derived temperatures should be similar to those from Brown and Parsons (1981).

Considering the experimental and thermodynamic data available, and applications to a wide variety of natural samples, we believe that the graphical geothermometer of Brown and Parsons (1981, Fig. 2) is the most accurate two-feldspar geothermometer available at this time. Because of the geometry of the construction and the distribution of experimental data, the accuracy of this geothermometer probably varies depending on the composition of the feldspars. It is probably best (±20–30°C?) for compositions like the Fish Canyon and Bishop feldspars, and becomes worse for feldspar pairs where the albite content of the sanidine is over 40 mole% or the albite content of the plagioclase is below 50 mole%. With the data presently available there is no way of estimating what the absolute accuracy may be.

The accuracy of the relationship of pressure to temperature as well as temperature alone is important for the feldspar geothermometer. Seck's (1971) experimental data were obtained at several pressures and the pressure coefficient (18°C/kbar) given by Brown and Parsons (1981) is generally within 10% of those calculated for the thermodynamic models. The pressure coefficient is probably much more accurate than the absolute temperature.

A factor of equal importance in the assessment of the accuracy of these pressure estimates is confirmation that the minerals coexisted in equilibrium and that the equilibrium compositions are preserved. Careful petrographic observation of the sequence of crystalization and textural interrelationships of the minerals is required, with analysis of a large number of mineral grains (checking for zonation and variation both within and between grains). Hildreth (1977) has presented evidence that the mineral compositions in the Bishop Tuff represent those coexisting in the magma immediately before eruption. In the Fish Canyon Tuff we have analyzed plagioclase and sanidine that are in contact or included in one another and oxides included in all phenocryst phases except quartz. These minerals represent a single population with little variation in composition, and we conclude that they coexisted in the magma with essentially their present composition (Whitney and Stormer, unpublished work). At magmatic temperatures feldspars tend to respond to changes in conditions by zoning or resorption whereas the oxides are more likely to react with rapid bulk compositional changes (Hammond and Taylor, 1982). In a disequilibrium situation we would expect to see the feldspars preserving a high temperature composition with the oxides reequilibrating to lower temperatures, the opposite of what is observed in these tuffs.

The scatter within samples for the Fish Canyon oxide minerals as shown in Figure 1 probably represents to some degree real variations in temperature among minerals from slightly different parts of the magma brought together by convective mixing or during eruption (Spera, 1983). The scatter may also, especially in the upper units, be a result of incipient late alteration of some grains. For some, but not all, magnetites there is an obvious alteration around the margins of the grains, apparently titanium-bearing maghemite (not oxidation exsolution to hematite). For geothermometry, obviously altered minerals were avoided, but analyses of both the obvious maghemite and apparently unaltered material show very similar compositions (especially Ti/Fe ratio) except for lower apparent totals for the maghemite. Calculated temperatures would be similar or at least not significantly higher than the actual magmatic temperatures. Reaction during eruption or postemplacement alteration would therefore be expected to produce lower apparent oxide temperatures, to reduce the difference between the oxide and feldspar temperatures, and to lower the apparent pressure.

Disequilibrium crystal growth could have displaced the phenocryst compositions from equilibrium values. A discussion of this by Dowty (1980, p. 442–446) suggests that, at least for the feldspars, the displacement may well be in the direction that would produce a higher apparent temperature. Again, it is impossible to predict what the effect of this phenomenon would be on the pressure estimate. The euhedral morphology (broken during eruption) and constant compositions of the phenocrysts and the large volumes of the tuffs indicate that these phenocrysts probably grew slowly with a small amount of undercooling in a large magma chamber. In that case the compositions should be close to equilibrium values. The eruption of silicic ash-flow tuffs apparently quenches their phenocryst compositions.

The accuracy of our pressure estimates cannot be assessed adequately on the basis of experimental or theoretical data presently available. There is some, admittedly circumstantial, evidence that the temperature–pressure relationships for the feldspars are at least roughly correct. In granulite facies metamorphic rocks there is often some independent mineralogical control on pressure. Stormer and Whitney (1977) presented the results of feldspar and pyroxene geothermometry for some granulites where the conditions of metamorphism could be independently estimated to be 750–800°C at 4–8 kilobars pressure. Recalculation of these data using Brown and Parsons (1981) gives peak feldspar temperatures of about 770°C (at 6 kbar) and, independently, 770–800°C using Lindsley and Anderson's (1983) pyroxene geothermometer at the same pressure. Bohlen and Essene (1977) and Bohlen et al., (1980) compared Fe–Ti oxide and two-feldspar geothermometry in the Adirondack granulite terrain. Using Stormer's (1975) feldspar geothermometer and Buddington and Lindsley's (1964) oxide geothermometer they found agreement of the two temperature estimates on a regional basis at an independently estimated pressure of 8 kilobars. In only two cases are analyses of both feldspar and oxide pairs in a single sample available. Our pressure calculation yields about 6 kbar for sample SR-31 (Bohlen and Essene, 1977) and about 8 kbar for sample GVR-274 (Bohlen et al., 1980). There are large analytical uncertainties in both studies, but the results suggest that the feldspar–oxide relationships we have proposed are at least very roughly correct.

At the lower end of the pressure range there are few instances of rocks containing the required mineral asemblage where a shallow depth can be assumed and the mineral compositions can be interpreted accordingly. Crecraft et al. (1981) present analyses of feldspars and oxides from rhyodacite flows at Twin Peaks, Utah which would give pressures of 0.5 and 1.5 kbar and analyses from later rhyolites which would give pressures up to 6 kbar. Other analyses of feldspars and oxides from various sources (i.e., Murat, Cantal and Isle of Eigg in Stormer, 1975, Table 1) and the epizonal granites studied by Whitney and Stormer (1976) give temperatures which coincide at pressures of about one kilobar or less. Although none of these data are of a quantity and quality to permit construction of meaningful *P–T* plots, any adjustment of the feldspar–oxide model to make the Fish Canyon Tuff and Bishop Tuff data coincide at pressures of 1–2 kbar would give physically impossible, large negative pressures for these volcanic rocks which have apparently equilibrated at shallow levels.

The evaluation of the probable accuracy of this method certainly indicates that there are large and unknown uncertainties in this pressure calculation, largely as a result of uncertainties in the experimental calibration of the mineral geothermometers. However, since there are no alternatives that are demonstrably better for pressure determination in silicic magmas, we have attempted to apply this method with mineral analyses from the Fish Canyon Tuff and Bishop Tuff. At the least, this should demonstrate the precision that could be attained, and a relative estimate of pressure.

Pressure and depth estimates

The results of applying this calculation to samples from the Fish Canyon and Bishop tuffs are shown in Figure 4. The oxide temperatures used for the Bishop Tuff samples were those calculated by Hildreth (1977) using Buddington and Lindsley (1964). Using the Spencer and Lindsley (1981) model for the Bishop Tuff plot in Figure 4 would raise the apparent pressures of the low temperature samples by about 1/2 kilobar, raise the pressure of the highest temperature samples by 2 to 3 kilobars, and raise the pressure of intermediate samples proportionately. (This would bring the pressure of all samples into closer agreement, but we have chosen to present the "worst case" considering the possible problems in the Spencer and Lindsley model for the oxides).

In comparison with the Bishop Tuff, the Fish Canyon Tuff is slightly higher in temperature and much higher in oxygen fugacity ($\log f_{O_2} = -11.3$ to -11.7, Whitney and Stormer, unpublished work). This is outside the range where Buddington and Lindsley (1964) can effectively be used and at the limit of applicability of the Spencer and Lindsley (1981) model.

All of the samples from the first-erupted, lower portion of the Fish Canyon Tuff (Fig. 4A) indicate pressures around 8 to 9 kilobars, equivalent to a depth of about 25 km. This is roughly equivalent to half the crustal thickness in this area (Prodehl and Pakiser, 1980).

For the pressures to fall in this relatively narrow range the Fe–Ti oxide temperatures and two-feldspar temperatures in each sample have to be closely correlated so that the difference in temperatures (ΔT in Fig. 3) is more consistent than T alone. In other words, relatively high oxide temperatures must occur in samples with high feldspar temperatures, and low oxide temperatures must occur with low feldspar temperatures. Since these two systems are completely independent chemically, it is likely that this consistency has a real physical basis. (If the variation in the feldspar and oxide temperatures was simply independent random analytical or sampling error, the range of pressures should be about 3 kilobars.)

We interpret the pressure values of 8 to 9 kilobars obtained from samples erupted early in this volcanic event as being representative of the actual conditions in the magma body where the phenocrysts originated. Samples from the later-erupted, upper cooling units of the tuff indicate variably lower pressures. These samples are from a petrographically distinct group of cooling units that form the upper part of one of the sections studied by O'Leary (1981) and Whitney and Stormer (unpublished work). These samples show extensive development of oxide–pyroxene rims on hornblende and biotite (low pressure dehydration reaction) and greater scatter in phenocryst compositions. These features are not seen in the earlier-erupted units and do not appear to be postemplacement effects. The lower pressures indicated for these samples in Figure 4A are interpreted as resulting from a greater opportunity for chemical adjustment toward lower-pressure conditions in the later-erupted portions of the magma, correlating with petrographic observations indicating partial reaction to a lower-pressure assemblage. Because these features are not seen in the earlier parts of the tuff and are not related to any significant temperature or compositional variation, it seems likely that they are a result of the eruption process rather than a feature of the preeruption magma body.

The apparent pressures calculated for the Bishop Tuff (Fig. 4B) are even more remarkable for their consistency. Of 60 samples for which data were available (Hildreth,

1977) 48 fall between 4.5 and 6 kilobars (equivalent to about 15 km depth). One major difference from the Fish Canyon Tuff is the apparent thermal, mineralogical, and chemical zonation in the Bishop Tuff with early erupted portions being lower in temperature than later erupted portions. The early, nonpyroxene-bearing, lower temperature samples all fall in a very tight cluster near 5 kilobars. Their phenocryst compositions are very homogeneous leading to small "error" bars in Figure 4A. We interpret this cluster of points from samples of the early part of the Bishop Tuff as representing conditions in the magma body where the phenocrysts originated.

More scattered, generally lower, apparent pressures are limited to samples from the higher temperature, pyroxene-bearing, later-erupted parts of the tuff (which also exhibit greater variation in phenocryst composition). Although there is apparent thermal and compositional continuity through the Bishop Tuff, its various lobes show differences that suggest a major shift in eruptive activity from early, low temperature flows directed toward the east and south to pyroxene-bearing, higher-temperature flows dominantly toward the north (Hildreth, 1979, p. 47). Hildreth suggested that this shift involved tapping of "deeper-level magma" at a different location along the caldera ring fracture. (Some of these higher temperature magmas do, in fact, plot at slightly higher pressures in Figure 4B.) We interpret the greater range of phenocryst composition within and between these pyroxene-bearing samples and the resulting greater range of apparent pressures as indicating a more complex eruptive history for this later, high temperature magma.

There are no other published data for two-feldspar ash-flow tuffs which are available to us and of sufficient quantity and quality to construct a meaningful *P–T* plot. In an abstract, Geissman et al. (1978) interpreted mineral data from an Oligocene dacite ash-flow tuff from Nevada as implying a mid-crustal source area. The oxide and feldspar temperatures apparently coincide at about 4 kbar. Magnetic properties, electron microscopy, and reflected light studies, show that the analysed oxides have not been altered (J. W. Geissman, Colorado School of Mines, pers. comm.).

The pressures and depths calculated for the Fish Canyon and Bishop tuffs are greater than has generally been assumed for the magmatic sources of silicic ash-flows (roughly 3–8 km; Hildreth, 1979, Fig. 16; Hildreth, 1981; Lipman, 1980; Eaton, 1980). Shallow subcaldera magma chambers could have formed immediately before or during the upward transport and eruption of the magma. However, if our pressure estimates are even approximately correct, the magmas considered here could not have developed their fundamental chemical or mineralogical characteristics in a shallow chamber. At least the early parts of the tuffs must have been transported from relatively deep (possibly mid-crustal) sources and erupted rather quickly for there to be so little evidence of phenocryst reequilibration.

Assessment of the source depths for ash-flow magmas

Although the depths calculated here are deeper than has generally been assumed for the sources of caldera-forming ash-flow eruptions, we believe that these deeper depths may well be possible, and that relatively deep sources may often be consistent with other observations. Most of the other data are equivocal or only permissive, but in total it suggests that the pressure estimates made above for the Fish Canyon and Bishop tuffs are generally correct. There are, in our opinion, no really unequivocal data on the ultimate source depths of the magma in such eruptions. Shallow plutons comagmatic with ash-flow tuffs do intrude the floors of eroded calderas. However, the intrusion of these plutons could easily have been a part of the eruptive event or have closely followed it. The shallow pluton cannot be proven to have been present before the caldera formed. Most of the thermal, structural, and geophysical manifestations of calderas do not really depend on the existence of a shallow magma chamber for any long period of time *before* the beginning of the initial ash-flow eruption. Hydrothermal convection systems indicative of shallow heat sources characteristically are established some time after the initial caldera-forming eruptions (Bethke et al., 1976: Heald-Wetlaufer et al., 1983)

The whole-rock compositions of the Fish Canyon and Bishop tuffs can be compared with experimental systems to infer something about pressure. Both the Fish Canyon and Bishop tuffs originated from silicate liquids in equilibrium with quartz, sanidine, and plagioclase and are analogous to compositions in the quarternary system Qz–Ab–Or–An (Carmichael, 1963; Winkler et al., 1975). Traditionally the compositions of silicic rocks have been compared with phase relations in the ternary system Qz–Ab–Or to infer something about pressure (for instance Hildreth 1977, Fig. 19). Although the system is not so well determined, the very significant effect of the anorthite component is better displayed as a projection of the quarternary quartz-saturated relationships onto the Ab–Or–An ternary as in Figure 5. The important features in this diagram are the cotectic lines indicating the compositions of silicate liquids in equilibrium with quartz, sanidine, plagioclase, and vapor at various pressures. In the absence of a vapor phase, these cotectic lines become surfaces dipping toward the Ab–Or sideline as shown by Whitney (1975).

Strictly speaking whole-rock analyses include the phenocrysts and, therefore, are not liquids in equilibrium with the phenocrysts. The phenocryst minerals may also have been enriched by "winnowing" during eruption and emplacement. Unfortunately, microprobe analyses of the natural glasses are also affected by hydration even where there is little or no visible sign of alteration. (Loss of Na and enrichment of K are particularly significant—Lipman et al., 1969; Whitney and Stormer, unpublished work).

Crystallization of "quench" microlites may also alter the compositions of the glasses.

In Figure 5, we have plotted the whole rock, phenocryst, and groundmass (apparent liquid) compositions for the Fish Canyon Tuff (Whitney and Stormer, unpublished work). The groundmass composition was obtained by subtracting the bulk phenocryst composition (sum of the products of the measured modal fractions and analyzed weight precents of oxides for all phenocrysts) from the whole-rock analysis. This agrees well with the analyzed glass compositions except for K_2O (which is higher in the glass because of the enrichment during hydration discussed above). Because the Fish Canyon magma was probably not vapor-saturated in its source region and the vapor-undersaturated system is not well known, we cannot determine an exact pressure. However, the position of all of these analyses indicates a pressure higher than estimated in Figure 4A.

The Bishop Tuff analyses from Hildreth's (1977) Table 21 are also plotted in Figure 5. These are apparently pumice lumps or fiamme with about 10–25% phenocrysts (Hildreth, 1977, Appendices I and III). We have taken the modal analyses of several samples with high phenocryst contents and the chemical analyses of the phenocrysts (from Hildreth, 1977) and have calculated the composition of the phenocryst assemblage. The major element composition of the phenocryst assemblage is very similar to the bulk composition of the rock (see phenocryst composition plotted in Figure 5). Because the phenocryst and bulk rock analyses are so similar the composition of the "liquid" must also be close to the bulk rock compositions (probably 1–2 mole% poorer in Or component in Figure 5). A gain or loss of phenocrysts would not significantly affect the major element composition.

For rock analyses with low CaO contents such as these (0.6–0.9 wt.%), the calculated amount of anorthite is very dependent on any cumulative errors in the alkali and aluminum analyses and is subject to greater uncertainty. The Bishop magma also was probably not vapor saturated in its source region so that it is not possible to use the plot in Figure 4 quantitatively. Interestingly, the Bishop analyses fall very roughly along a trend parallelling the Ab–Or sideline moving toward higher Ab contents as temperature decreases. Although this "trend" is hardly significant, it would represent a path of increasing H_2O activity along a vapor-undersaturated cotectic surface. The lowest-temperature samples would be near a 5 kilobar vapor saturated cotectic. This would be consistent with the results in Figure 4B.

Hildreth (1977) made calculations of total pressure for the Bishop Tuff based on the equilibration of $FeSiO_3$ component in the pyroxene with magnetite and quartz (an assemblage only found in the higher temperature, later parts of the tuff). These results ranged from over 4.5 to less than 1 kilobar, falling with temperature along a trend from 790 to 737°C. This reaction is very sensitive to the value of f_{O_2} and other temperature-dependent factors and it is unlikely that the *P–T* trend represents any real preeruption *P–T* condition.

Since the Bishop Tuff is relatively young and there is continuing volcanic and tectonic activity in the Long Valley and Mono Basin area, there have been a number of geophysical as well as geological investigations (see papers from the Long Valley Symposium, Journal of Geophysical Research, V. 81, p. 721–860, 1976 and abstracts in the symposium Calderas and Related Rocks, EOS: Transactions of the American Geophysical Union, V. 64, p. 872–891). Seismic data, which of course only reflect

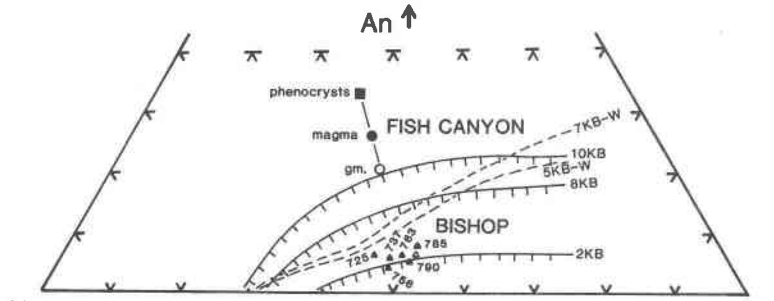

Fig. 5. The normative compositions of the Fish Canyon magma, phenocrysts, and "groundmass" (from Whitney and Stormer, unpublished work) and normative compositions of several Bishop Tuff samples (from Hildreth 1977, Table 21) are projected onto a ternary representation of a portion of the quartz saturated Ab–An–Or system (weight percent). The Bishop Tuff samples are shown as small solid triangles labled with the corresponding iron–titanium oxide temperature as determined by Hildreth (1977). The composition of the bulk phenocryst assemblage (about 20% of the sample) for the sample labled 790 is plotted as an open triangle. The phenocryst assemblage in other Bishop Tuff samples is similar but generally the phenocryst percentage is lower. The solid curves show the projected, experimentally determined composition of liquids coexisting with quartz, plagioclase, sanidine, and vapor at 2, 8, and 10 kilobars from Whitney (1972, 1975). The hachures indicate the direction of the shift in composition for vapor undersaturated liquids coexisting with the same mineral assemblage. Comparable curves for 5 and 7 kilobars from Winkler et al. (1975) are shown by dashed lines.

present conditions, suggest that there may now be magma at depths of as little as 5 kilometers under the Long Valley caldera extending to a depth of 15–20 km (Hill, 1976; Steeples and Iyer, 1976; Ryall and Ryall, 1981; Sanders, 1983; Luetgert and Mooney, 1983). The portion above 8 km seems to be two separate small stocks or cupolas (Sanders, 1983) The gravity data of Kane et al., (1976) also suggest a deep low-density mass at a depth of 8–16 km. However, Lachenbruch et al., (1976) state that "if magma had persisted at such depths (5–7 km) throughout the recent eruptive history of Long Valley, we should expect a more conspicuous heat flow anomaly . . .". All of these observations would be consistent with an original, deep source in the vicinity of 15 km, as indicated by Figure 4B, with higher level chambers or stocks developing during and after the eruption of the Bishop Tuff.

Bailey et al. (1976 and 1983) have suggested that a developing rhyolitic magma chamber analogous to the Bishop Tuff magma chamber underlies the Mono Craters area a few kilometers northwest of the Long Valley caldera. Geothermal considerations imply that, if such a magma chamber exists, its roof is deeper than 8–10 km (Lachenbruch et al., 1976). Hermance (1982) and Hermance et al. (1983) have also suggested, on the basis of magnetotelluric measurements, that no large-scale magma reservoirs exist in the upper crust in the Long Valley/Mono Basin area. An anomalous conductor is, however, present at midcrustal depths (15–20 km). The conclusion drawn by Hermance (1982) was that episodic high level intrusions occur and are then relatively rapidly cooled and solidified.

The Coso volcanic field, also in eastern California, has erupted numerous domes of high-silica rhyolite within the past 300,000 years (Duffield et al., 1980; Bacon et al., 1981). Here also, geophysical evidence suggests that, if a silicic magma chamber exists, its roof must be at depths greater than 8–10 km (Bacon et al., 1980).

Large silicic magma chambers may be preferentially formed at these midcrustal depths because of fundamental changes in the character of the crust at this level (Eaton, 1980). For instance, in the Basin and Range Province the numbers of earthquake epicenters fall off rapidly with depth between 10 and 15 km, and they are almost absent below 15 km (Eaton, 1980, Fig. 9.10), indicating a transition from brittle behavior above this depth to more ductile deformation below. Changes in the density and composition of the crust at this level could also be an important factor.

Perhaps the most accurately located, extensive crustal magma body is near Socorro, New Mexico. Although the Socorro magma body is assumed to be basaltic, intrusions of basalt are generally thought to provide the heat source for generating and maintaining silicic magma bodies in the crust (see discussion in Hildreth, 1981, p. 10182). As described by Rinehart et al. (1979), this body is apparently a horizontal sill 25 by 60 km, perhaps 1 km thick and located at 20 km depth. Rinehart et al. (1979) suggest that it is trapped at that midcrustal level by a feature corresponding to the Conrad seismic discontinuity. If mafic intrusions such as this are important in generating and maintaining large silicic magma bodies, the silicic magma chambers may develop at the same midcrustal levels.

Figure 6 displays the depths and temperatures we have calculated for the Fish Canyon and Bishop magmas in relationship to several geotherms calculated by Lachenbruch and Sass (1977) for various provinces in the western U.S. Both fall on isotherms within the range of observed heat flow in the Basin and Range province. However, the Bishop magma would (as it should) represent a thermal anomaly with temperatures 200–300°C above the geotherms characteristic of most of the Basin and Range province. It is impossible to say what the geothermal characteristics of the crust were in the San Juan field before the initiation of volcanism and later tectonic extension of the region. Even at 25 km depth it is not unreasonable to suppose that the Fish Canyon magma was produced by a thermal anomaly of 200–300°C (produced at midcrustal levels by basaltic intrusions) relative to a preexisting thick "stable" crust. Even the high geothermal gradients characteristic of most of the Basin and Range province would not result in temperatures high enough to produce magmas like the Fish Canyon and Bishop tuffs at the depths we suggest. Even if the magmas are generated at these midcrustal depths, large-volume silicic volcanic fields must require thermal input from the intrusion of basaltic magma. The location of appropriate source rock and the activity of water also would limit melting and the development of such magmas.

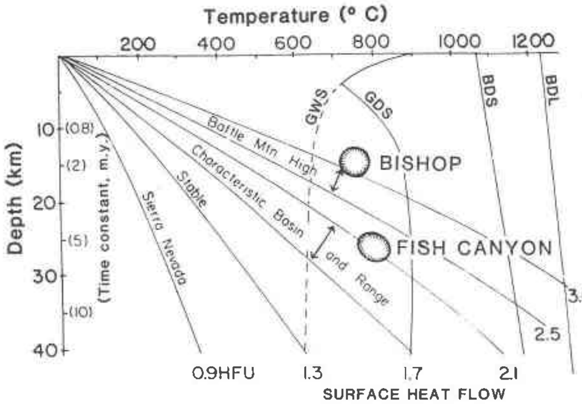

Fig. 6. The temperature and depth of origin for the Fish Canyon and Bishop magmas as compared with the generalized conductive temperature profiles calculated by Lachenbruch and Sass (1977) for several regions in the western U.S. The granodiorite solidus for "wet" (GWS) and "dry" (GDS) conditions and the basalt "dry" solidus (BDS) and liquidus (BDL) (Wyllie, 1971) are shown for reference. The time constant scale, on the left side of the diagram in parentheses, gives the time necessary to establish an equilibrium temperature profile above a higher temperature thermal source emplaced in the crust. Modified from Fig. 22, Lachenbruch and Sass, (1977).

It is interesting to note that the value of the time constant for establishment of an equilibrium thermal profile above the possible anomaly (see scale left side of Fig. 6) is roughly equivalent to the length of time between the inception of volcanism and the eruption of the ash flow for both the Fish Canyon (5 m.y.) and Bishop (2 m.y.) tuffs.

Ash-flow tuffs, calderas, and granitic ring complexes have been shown to be related parts of silicic magma systems (see Jacobson et al., 1958, and Roberts 1966, 1974). Myers (1975) and Bussell et al. (1976) describe deeper ring structures in the Coastal Batholith of Peru that feed shallow, thin, horizontal plutons within the ring fractures (Myers, 1975, Fig. 10). They suggest that roughly cylindrical masses of the roof enclosed within the ring fractures are stoped into the underlying batholith while magma is injected along the ring fractures to form these high level plutons above the stoped "plug." They also suggest that such intrusions are, at least in some cases, connected with caldera-forming eruptions at the surface. This is an attractive model for bringing large volumes of silicic magma near the surface rapidly. The near-surface chamber would serve to localize the caldera structure and would allow for the high-level storage that we apparently see in the partial equilibration of the later-erupted portions of the tuffs. Presumably if a large volume of originally vapor-undersaturated magma were rapidly emplaced above the depth at which saturation would occur (about 5–6 km in the case of the Fish Canyon Tuff) it would erupt catastrophically when a vapor phase was nucleated and rapid vesiculation occurred. In such a model the precaldera rhyolite domes and flows would represent small volumes leaked slowly from the magma at depth. Postcaldera volcanism and the intrusions responsible for resurgence would represent the residual, less volatile-rich, magma remaining after the ash-flow eruption. (In many areas where erosion has exposed the subcaldera plutons, these late intrusions would obscure the field evidence for the early structures associated with the initial ash-flow eruption.) This model represents a possible mechanism relating the results of our study to caldera structure seen in other geological studies. It could explain the anomalies in the estimates of the magma "drawdown" (Smith, 1979, p. 8 and Fig. 2) made by assuming a cylindrical magma chamber with a circular section equal to caldera area and height calculated to equal tuff volume.

None of the previous discussion is meant to imply that all ash-flow tuffs are erupted from sources at greater than 10 km depth. Those ash-flow tuffs that contain a single feldspar (Ab-rich sanidine) such as the Bandelier Tuff (Smith and Bailey, 1966), the Tala Tuff (Mahood, 1981) or the Sunshine Peak Tuff of the western San Juan field (Lipman et al., 1978) probably represent hotter and somewhat shallower magmas. In the central San Juan field some preliminary analyses (Stormer, unpublished data) suggest that ash-flow tuffs immediately following the Fish Canyon Tuff (Carpenter Ridge and Mammoth Mt.) may have been erupted from much shallower chambers. In nested caldera systems, it may be common to develop shallower chambers for successive ash-flow magmas as the system develops (The Yellowstone calderas might be a more recent example of such a system with presently existing shallow chambers. See Eaton et al., 1975 for a review of this area.) The evidence discussed earlier would suggest shallower magma bodies beneath parts of the Long Valley caldera at the present time.

Conclusions

We have shown that a technique combining Fe–Ti oxide and two-feldspar geothermometry could be very useful for determining the pressure or depth of origin for the phenocryst assemblage of certain ash-flow tuffs. The use of this method would require good analytical technique and detailed petrography on a large suite of carefully collected samples. Analyses from a few random samples, where the mutual coexistence and compositional coherence of the phenocryst assemblage are unproven, are likely to give misleading results. However, a precision within about 1 kilobar is possible as shown by the Bishop Tuff samples plotted in Figure 4B.

The experimental calibration of these geothermometers is not nearly so good as the achievable analytical precision. Further experimental work in the ternary feldspar system and new modeling in the Fe–Ti oxide system would be tremendously useful. Nevertheless, the consistent pattern we see in other data from a variety of sources suggests that the pressure and depth estimates we have made are not greatly in error. It is not possible to quantify the uncertainty in our estimates of depth, but we feel that we are probably within 20–30% of the true value.

The data and observations presented here suggest that at least some ash-flow magmas originate at midcrustal depths rather than in high-level chambers. The phenocryst assemblage must be developed at these depths, and must be transported to the surface rapidly enough to avoid major chemical modification. Intrusion into a short lived, high level, subcaldera chamber is not precluded as a part of the eruptive process, and may, in fact, be supported by the partial reequilibration seen in the later portions of these tuffs. However, these magmas developed their fundamental chemical characteristics at depths greater than 10 km. The differentiation processes controlling composition must have operated at those depths.

The depths of 15 km for the Bishop Tuff and 25 km for the Fish Canyon Tuff that we have estimated for the origin of the phenocryst assemblage do not agree with the widely held belief that large volume ash-flow magmas develop in "high level" magma chambers. Although there is no unequivocal proof, the relatively deep sources we prefer are consistent with some other data and observations. The possibility of deep magma chambers in the early stages of large-volume silicic systems should at least be considered in geophysical modeling, geothermal exploration, and volcanic hazards assessment as well as petrological studies. There are probably several interesting relationships between the depth of origin, composition, and temperature of large-volume ash-flow magmas that can be explored as more data become available. Homogeneous quartz latite magmas probably develop at the greatest depths and one-feldspar, high-silica rhyolite magmas probably develop at relatively shallow depths and high temperatures. The volume of individual ash-flow tuffs may also be related to depth, with the largest coming from the greatest depth. Smaller volume ash-flow magmas may develop in relatively "high level" magma chambers. In nested caldera complexes, such as the central San Juan calderas, there may be some trend toward a progressive shallowing of the magma source.

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Note added in proof:

Whitney and Stormer (unpublished work) has recently been accepted: Whitney, J. A., and Stormer, J. C., Jr. (1985) Mineralogy, petrology, and magmatic conditions from the Fish Canyon Tuff, central San Juan Volcanic Field, Colorado. Journal of Petrology (in press).

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