Journal of Structural Geology, Vol. 14, No. 3, pp. 291 to 300, 1992
Printed in Great Britain

0191-8141/92 $05.00+0.00

Rates of processes in magmatic arcs: implications for the timing and nature of pluton emplacement and wall rock deformation

SCOTT R. PATERSON

Department of Geological Sciences, University of Southern California, Los Angeles, CA 90089-0740, U.S.A.

and

OTHMAR T. TOBISCH

Earth Science Board, University of California, Santa Cruz, CA 95064, U.S.A.

(*Received* 18 *June* 1990; *accepted in revised form* 28 *July* 1991)

Abstract—The construction of arcs and in many cases the emplacement of plutons occur in tectonically active regions. It is critical, therefore, to evaluate the rates of structural and magmatic processes when trying to understand the evolution of arcs and the associated pluton–wall rock systems. Our best estimates of average rates or durations of processes in shallow to moderate level arc environments are the following: (1) crystal growth rates in magma of 10^{-4} cm year^{-1}; (2) growth rates of metamorphic porphyroblasts between 10^{-5} and 10^{-2} cm year^{-1}; (3) long-term magma supply rates of 10^{-1} km^3 year^{-1} and short-term rates of up to 350 km^3 year^{-1}; (4) diapiric ascent rates for mafic plutons of 1–3 m year^{-1}; (5) cooling of plutons to ambient wall rock temperatures in 10^5–10^6 years; (6) final crystallization of plutons in a small fraction of the time needed for complete cooling; (7) fault displacements of 3 cm year^{-1}; (8) development of cleavages in fault zones in less than 10^6 years at strain rates of 10^{-13} s^{-1} or higher; and (9) the development of regional cleavages in 10^6 years at strain rates of 10^{-14} s^{-1}. These rates indicate that processes operating in magmatic arcs are relatively rapid: pluton emplacement, cleavage development, etc., occur over time spans of tens of thousands to no more than a few million years. However, the rate of ascent and crystallization of plutons at shallow levels is generally shorter than that needed to get large displacements on faults or widespread cleavages developed. At deeper levels, or in zones undergoing faster strain rates, the time spans of the various processes approach one another. Thus plutons, with otherwise similar characteristics, emplaced in regions undergoing fast strain rates, or at deeper crustal levels, may appear quite different structurally from those emplaced at shallow levels or in regions undergoing slower strain rates. Comparison of these data also suggests that the rate at which wall rock deforms is the limiting factor controlling the rates of other structural processes during emplacement of plutons unless fast strain rates or multiple deformation mechanisms are considered. Thus emplacement mechanisms that rely on the transport of magma over 10^5 to 10^6 years are favored and need further consideration. Finally, we argue that the structural and other characteristics of pluton–wall rock systems will depend on rates of various processes involved and that these rates at the very least influence, and sometimes invalidate, the timing criteria previously published by us and others.

INTRODUCTION

RATES of geological processes have been notoriously difficult to obtain, largely for two reasons. First, a great deal of uncertainty exists in the extrapolation of the rates of experimental processes to natural conditions (e.g. Paterson 1987). Second, techniques for dating geological events have been imprecise (e.g. errors on the date may be larger than the duration of the event). There is now increasing evidence that geological events develop over recognizable discrete periods of time, and that the rates of individual processes can vary significantly (Price 1975, Pfiffner & Ramsay 1982). Because much of our understanding of the evolution of orogenic belts depends on the interpretation of structural features, we need to consider how different rates and durations of events influence the nature and timing of such structures.

In this paper, we investigate how the rates of certain processes influence the structure of pluton–wall rock systems in magmatic arcs, as well as affect criteria proposed previously for determining the relative timing of pluton emplacement and regional deformation (e.g. Paterson & Tobisch 1988, Paterson *et al.* 1989b). We view the construction of magmatic arcs as a combination of magma dynamics (e.g. Fedotov 1981) in which magma is generated in or near the aesthenosphere and migrates by conduits to magma chambers (some of which may rise as diapirs) and simultaneous regional deformation of the wall rocks by folding, faulting and cleavage development (e.g. Tobisch *et al.* 1986, Paterson *et al.* 1991a).

We will focus on a comparison of rates of the following processes: faulting, folding, cleavage development, magma supply and ascent, pluton emplacement, crystallization and cooling, and growth of igneous and metamorphic minerals. Obviously the rates we discuss in this paper are estimates of average rates for each process: the actual rates will vary depending on a variety of factors, the discussion of which is often beyond the scope of this paper. The main point we wish to make by this comparison, is that the final structural and textural characteristics of the pluton–wall rock systems are rate dependent.

RATES OF FAULTING

At shallow levels in the crust where 'stick–slip' faulting occurs, short-term rates of faulting (i.e. during an earthquake) can be large (Sibson 1986). However, since our immediate goal is to look at magmatic processes and the development of ductile structures, we are more interested in averages of long-term rates of creep in fault zones taking place at deeper crustal levels. Eventually, however, it would be worthwhile to consider the effect of rapid short-term rates of faulting on, say, plutons still in the crystal–liquid state.

One method for calculating long-term rates of fault motions in a magmatic arc setting is to use the plate motions to calculate the rates of possible pure thrusting or pure strike-slip motion (Fig. 1). Values calculated in this fashion assume rigid plate behavior and complete coupling between plates and thus reflect maximum rates. We have made such calculations for the Sumatra region based on the work of Hamilton (1979) (see also McCaffrey *et al.* 1990), which give values of 3 cm year^{-1}. Ekstrom & Engdahl (1989) have made similar calculations for the Aleutians, which give values of 0–7 cm year^{-1} depending on the azimuth of plate motions relative to the arc. This agrees well with the average strike-slip component along the San Andreas fault of up to 7 cm year^{-1} (e.g. Sieh & Williams 1990), and studies such as Harper *et al.* (1989) that suggest rates of thrusting in older arcs of 1–10 cm year^{-1}. Naar & Hey (1989) have noted that maximum rates of oceanic transform faults and ridge spreading are around 14–16 cm year^{-1}, which provides a likely upper limit for rates of faulting in arc settings.

In other tectonic settings, published rates of faulting tend to be slower. For example, Price (1981) noted that in foreland fold and thrust belts average rates of displacement fall between 0.2 and 0.5 cm year^{-1}. Sarawitz (1988) noted rates of up to 1.6 cm year^{-1} displacement during foreland thrusting in Argentina, and Hacker *et al.* (1990) suggest rates of 2.3 cm year^{-1} in the Ruby Mountains core complex, Nevada.

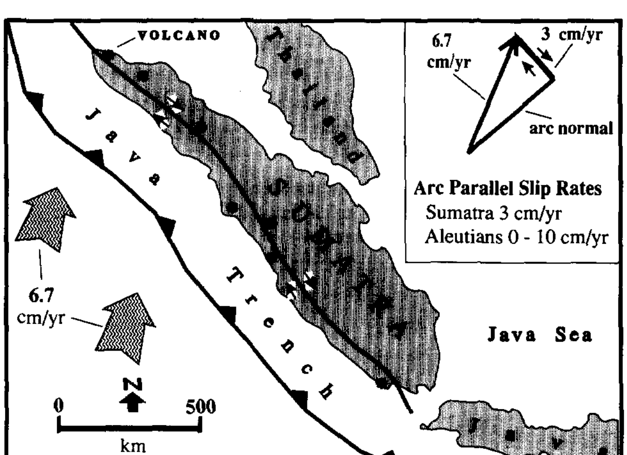

Fig. 1. Simplified tectonic outline of the magmatic arc and plate convergence near Sumatra showing method of calculating intra-arc fault slip rates. Geology after Hamilton (1979).

For the sake of comparison, we will use an average rate of faulting in arc settings of 3 cm year^{-1}. During some periods long-term rates may decrease to 0 cm year^{-1} (no faulting) or increase up to 10–16 cm year^{-1}. The crucial point here is that long-term rates of faulting will not generally exceed these limits.

RATES OF CLEAVAGE DEVELOPMENT

Rates of cleavage development have been estimated in two ways: (1) by calculating rates of contraction due to folding associated with the cleavage (e.g. Price 1975, Rockwell *et al.* 1988, Suppe 1989); and (2) by calculating the strain associated with the cleavage and determining the time span for this deformation (e.g. Kligfield *et al.* 1981, Pfiffner & Ramsay 1982). It should be emphasized that these calculations assume that the strain occurred by the shortest deformation path. If deformation was more complicated, for example involving the unfolding of folds, or several periods of non-coaxial deformation, the rates will represent minimum values.

Pfiffner & Ramsay (1982) presented equations relating total strain, strain ratios, average values of shortening and strain rates. They point out that strain (and strain ratios) accumulate in a non-linear fashion and that this rate of accumulation varies for pure shear (most efficient) and simple shear (less efficient) strain histories (Fig. 2). They also noted that typical strain rates during cleavage development are between 10^{-13} and 10^{-15} s^{-1}. Schmid (1989) noted that faster strain rates (10^{-10}–10^{-13} s^{-1}) may be associated with zones of very high strain (e.g. mylonite zones).

We have used the equations of Pfiffner & Ramsay (1982) to calculate the amount of time needed to get different amounts of total shortening and total extension given a particular strain rate. Cleavages often become visible in the field after about 20% shortening (Cloos 1949, Ramsay 1967, Tobisch 1984), which, given an average strain rate of 10^{-14} s^{-1}, would occur in 700,000 years. Many mountain belts have regionally developed cleavages and associated shortening of 50–70% (e.g. Wood 1974, Tobisch *et al.* 1977, Pfiffner & Ramsay 1982, Siddans *et al.* 1984, Paterson *et al.* 1989a). These

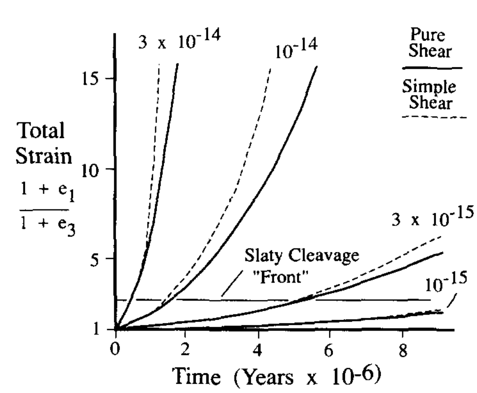

Fig. 2. Diagram showing the non-linear accumulation of strain (total strain ratio vs time) for pure and simple shear models. Diagram simplified from Pfiffner & Ramsay (1982).

cleavages would form in 2–4 Ma at a strain rate of 10^{-14} s^{-1} or as quickly as 200,000–400,000 years at a strain rate of 10^{-13} s^{-1}. If, as suggested by Schmid (1989), strain rates are even faster in narrow mylonite zones, intense cleavages could form in 20,000–40,000 years.

How does the average rate of faulting of 3 cm $year^{-1}$ noted in the previous section compare with the rates of cleavage development in ductile shear zones noted in this section? Given a 1 km wide shear zone undergoing homogeneous simple shear, a displacement rate of 3 cm $year^{-1}$ will cause a shortening (along the least principal axis of the strain ellipsoid) of 62% in 33,000 years. Clearly cleavages can develop in short time spans in ductile faults. At slower strain rates outside of fault zones, cleavages associated with comparable values of strain can still form relatively rapidly, e.g. between 300,000 years at 10^{-13} s^{-1} and 3 million years at 10^{-14} s^{-1}.

Price (1975) gave examples of folds forming at displacement rates of up to 10 cm $year^{-1}$, but he also noted that shortening rates probably vary during different phases of fold development. Rockwell *et al.* (1988) and Suppe (1989) calculated rates of displacement between 0.9 and 0.25 cm $year^{-1}$, respectively, associated with recent folding of Tertiary sediments. Vita-Finzi (1979) estimated displacement rates of 1.7–3.2 cm $year^{-1}$ associated with Holocene folding in the Zagros Mountains (Iran). These displacement rates indicate moderate to quite fast rates, depending on the scale of observation (individual fold vs whole mountain belt).

RATES OF PLUTON COOLING

Two aspects of cooling plutons need consideration: (1) how fast does the final crystallization (i.e. from 50% to less than 30% melt) of the pluton take? and (2) how fast will a crystallizing magma body cool to ambient wall rock temperatures?

How long, then, does it take a pluton to cool from emplacement temperature to ambient wall rock temperatures? This is a function of the initial temperature of the melt, wall rock temperature, thermal conductivity, size and shape of the pluton, duration of magma emplacement, and method of cooling among other factors (e.g. Jaegar 1968, Harrison & McDougall 1980, Barton *et al.* 1988). The studies listed in Table 1 suggest that the duration of cooling can vary from a few years for small dikes at shallow levels, to millions of years for moderately sized plutons in batholiths at mid-crustal levels.

Many authors have noted that as magma cools and crystallizes, it rapidly increases in viscosity over a narrow temperature range (Fig. 3) (e.g. Shaw 1965, Arzi 1978, Van der Molen & Paterson 1979, Wickham 1987, Cruden 1990). This rapid change in viscosity occurs between about 55% (Marsh 1981) and 75% crystals (e.g. Arzi 1978, Van der Molen & Paterson 1979) and takes place over a temperature range of a few hundred degrees centigrade (Cruden 1990). Paterson *et al.* (1989b) noted that foliations and microstructures form in plutons during a continuum of conditions from magmatic flow to low-temperature, solid-state flow (Fig. 3). They divide this continuum into four categories: magmatic flow, submagmatic flow, high temperature solid-state flow, and low-temperature solid-state flow (not shown on Fig. 3). We suggest that which type of flow occurs largely reflects the changing viscosity (crystal %) of the magma: magmatic flow occurs in crystal–melt systems with

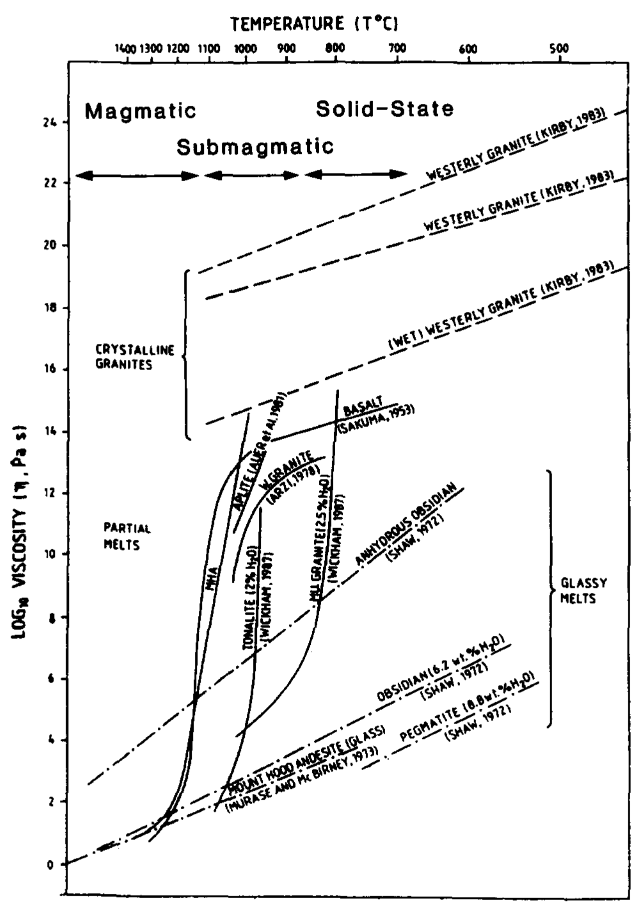

Fig. 3. Diagram showing a generalized relationship between temperature and viscosity of a crystal–melt system. Redrafted from Cruden (1990).

Table 1. Cooling durations of magma bodies

Object	Duration	Reference
30 m dike (shallow)	7 years	Savage (1974)
Magma bodies	Years to Ma	Jaeger (1968)
Pluton (1000°C, 20 km)	2–4 Ma	Barton *et al.* (1988)
Quottoon Pluton (700–450°C)	2 Ma	Harrison & Clarke (1979)
Separation Point Pluton (700–450°C)	2 Ma	Harrison & McDougall (1980)
Sierra Nevada Batholith	>10 Ma	Stern *et al.* (1981)

greater than 50% melt, submagmatic flow occurs between 20 and 50% melt, and high-temperature solid-state flow will occur when less than 20% melt is present (Fig. 3).

Whether or not a pluton undergoes magmatic, submagmatic or solid-state deformation (or preserves evidence of these processes) depends on the rate of change in crystal content or magma viscosity. Because this change will occur over a restricted temperature range, we argue that *the length of time needed to change from magmatic to solid-state behavior in a pluton will be much shorter than the overall length of time of cooling of the pluton.* Thus it is likely that the rapid viscosity change in magmas will vary from a year to less than 1 Ma depending on the factors noted above.

CRYSTAL GROWTH RATES

Another means of estimating crystallization rates of plutons is to look at the rates of crystal growth and compare these rates to the average crystal size in plutons. Swanson (1977) suggested that crystals of quartz, K-feldspar and plagioclase grew at rates of 3 mm day^{-1} to 1 mm $year^{-1}$ at 8 kb and 400–900°C. Cashman & Ferry (1988) noted crystal growth rates in basaltic lava lakes of around 0.003 mm $year^{-1}$. Cashman (1990) has summarized many of the factors affecting crystallization rates as well as summarized available data on crystallization rates. Most studies cited by Cashman (1990) suggest rates of 10^{-2}–10^{-3} mm $year^{-1}$.

Even the largest crystals in plutons on average do not usually exceed 10 cm in length. If we consider a growth rate of 3×10^{-3} mm $year^{-1}$, these large phenocrysts could grow in 33,000 years. We might assume that this crystal growth rate is an order of magnitude too fast since crystals may become resorbed or grow at different rates during the crystallization process (e.g. Cashman 1990). Even so, all of the crystals in plutons could easily grow to observed sizes in a few hundred thousand years.

Crystal growth rates are also important when considering the relationships between metamorphic porphyroblasts and structures. At present there is little agreement on how fast metamorphic porphyroblasts grow. Some studies suggest that porphyroblasts take up to 10 million years to grow (Christensen *et al.* 1989), or possibly grow episodically throughout the evolution of an orogenic belt (Bell & Johnson 1989). Other studies suggest that growth rates are geologically very rapid and that many metamorphic minerals grow in hundreds to hundreds of thousands of years (e.g. Walther & Wood 1984, Ridley 1986, Cashman & Ferry 1988, Joesten & Fischer 1988). Detailed studies that consider the kinetics of metamorphic mineral growth generally estimate growth rates of about 2×10^{-5} cm $year^{-1}$ (Ridley 1986, personal communication 1990) for regional metamorphism, to rates as high as 1.3×10^{-2} cm $year^{-1}$ in contact aureoles (e.g. Joesten & Fischer 1988). If such rates are correct, a porphyroblast 5 cm long could grow in 300–300,000 years.

RATES OF MAGMA SUPPLY AND PLUTON EMPLACEMENT

Movement of magma from its source region to higher crustal levels involves several processes any of which can be the rate limiting process. These steps include the supply of magma to the network of fractures and magma chambers, the transport of magma through the system, and the final emplacement (if considering plutons) or extrusion (if considering volcanoes). It should be noted here that most direct measurement data available (and much of the theoretical calculations) are for basaltic or andesitic systems, which will differ somewhat from the lower temperature, more viscous felsic magma systems, especially when it comes to estimating ascent velocities.

Crisp (1984) and Shaw (1985) noted that the volumetric ratio of intrusive to extrusive materials in arcs is about 10 to 1. Crisp summarized data in support of long-term rates of magma addition in arc settings of about 10^{-1} km^3 $year^{-1}$, whereas Shaw (1985) suggested that long-term averages of the addition of intrusive material was around 10^{-2} km^3 $year^{-1}$. Using the slowest long-term rate (10^{-2} km^3 $year^{-1}$), the magma necessary to make a moderate-sized pluton (1500 km^3) would accumulate in 150,000 years. Sufficient magma to form even some of the large intrusive suites in the Sierra Nevada (30,000 km^3) could accumulate in 3 Ma.

Because these long-term rates generally reflect averages of episodic magma transport (Fedotov 1981, Wadge 1981), short-term rates are more likely to be important when considering rates of volcanic eruptions or the filling of subvolcanic chambers (i.e. plutons). Short-term rates on a scale of days to years range from 1 to 0.1 km^3 $year^{-1}$ (Swanson 1972, Crisp 1984, Decker *et al.* 1987). If we consider an average short-term rate of 1 km^3 $year^{-1}$, sufficient magma could accumulate at a spot to form moderate-sized plutons (1500 km^3) to large plutons (30,000 km^3) in 30,000 years or less.

Table 2 lists ascent rates of magma or diapirs based on both theoretical studies and field measurements. Field measurements are based on analogies with the rise of salt domes (Seni & Jackson 1984, Talbot & Jarvis 1984), the rate of uplift of strata overlying laccoliths (Corry 1988) or buried plutons (Relinger *et al.* 1980) and seismic

Table 2. Emplacement rates of magma and diapirs

Observation	Rate (m $year^{-1}$)	Reference
Salt diapirs	0.005–0.17	Seni & Jackson (1984) Talbot & Jarvis (1984) Jackson & Talbot (1986)
Magma body	0.005	Relinger *et al.* (1980)
Dikes	10^6–10^7	Delaney & Pollard (1981)
Laccolith	266	Corry (1988)
Magma	3×10^7	Spera *et al.* (1985)
Plutons	2	Mahon *et al.* (1988)
Plutons	0.4 to m $year^{-1}$	Anderson (1981)
Basalts	Average = 3	Gill (1981) Decker *et al.* (1987)
Mafic magma	1.5–4	Marsh (1982)
Andesites	0.1–1	Gill (1981)

studies of the rise of magma beneath volcanic centers (e.g. Eaton 1962, references in Gill 1981, Decker *et al.* 1987). Theoretical studies are based on the rise of hot convecting 'magma spheres' (Marsh 1978, 1982, Mahon *et al.* 1988) or the ascent of magma by crack propagation or 'magma fracturing' (Marsh 1978, Shaw 1980, Spera 1980, Delaney & Pollard 1982).

These studies indicate that magma can be emplaced much more rapidly by crack propagation methods than by diapiric ascent (Marsh 1978, argued for rates of magma transport 10^6 times faster by crack propagation than diapiric rise, but noted that magma *must* travel at least 10^4 times faster by crack propagation to prevent cooling) and that rates of ascent show a wide range (Table 2). Ascent rates are the most poorly constrained and most variable parameter of the processes we have examined.

However, several types of studies suggest that diapiric ascent rates of andesites and basalts fall between 1 and 3 m year^{-1}. These studies include theoretical calculations by Marsh (1982), seismic studies beneath active volcanoes (Decker *et al.* 1987) and an evaluation of rates of ascent needed for andesitic magma to transport mafic magma globules or mafic inclusions (e.g. Gill 1981). For the purposes of this study, we will use an average rate of diapiric ascent for andesites and basalts of 2 m year^{-1}.

In summary, magma supply rates indicate that, given a suitable conduit system, sufficient volumes of magma can be transported to a locality to form most plutons in hundreds to hundreds of thousands of years. Some magma may rise as diapirs instead of along conduits, in which case ascent rates of the diapirs may range up to 3 m year^{-1}. Diapiric ascent rates and the rates of final pluton emplacement, in part, depend on the rates of wall rock deformation. We will evaluate these rates in a later section of this paper.

COMPARISON OF RATES OF PROCESSES

We have suggested that the length of time over which different processes operate in magmatic arcs is quite variable (Tables 1 and 2). One means of comparing these different processes in a meaningful context is to constrain the duration of one process, and then consider the extent to which other processes would operate during that time. With this in mind, we will assume that the rapid change in viscosity of magmas will take place on average in 250,000 years, a reasonable estimate for mid- to shallow-crustal levels given current data. This implies that the crystal–melt system has cooled a few hundred degrees centigrade. During that time (Table 3), igneous crystals could grow to 75 cm, metamorphic crystals could grow to 5 cm during regional metamorphism and many tens of centimeters during contact metamorphism, 2500–250,000 km^3 of magma could be supplied to a chamber by flow in conduits, magma diapirs could ascend in considerable excess of hundreds of kilometers, slip along faults would be 7.5–15 km, and at strain rates of 10^{-14} and 10^{-13} s^{-1}, respectively,

Table 3. Extent of various processes in 250,000 years

Process	Average rate	250,000 years
Igneous crystal growth	3×10^{-4} cm year^{-1}	75 cm
Metamorphic crystal growth		
Regional	2×10^{-5} cm year^{-1}	5 cm
Aureoles	1.3×10^{-2} cm year^{-1}	3250 cm
Magma supply		
Short-term	365 km^3 year^{-1}	9×10^7 km^3
Long-term	0.01 km^3 year^{-1}	2500 km^3
Diapiric ascent	2 m year^{-1}	500 km
Fault slip	3 cm year^{-1}	7.5 km
Cleavages	10^{-14} s^{-1}	−4%
	10^{-13} s^{-1}	−33%

regional shortening of −3% to −33% would occur. Obviously other factors (such as reversals of processes, supply of heat or of appropriate reactants) may retard, accelerate or otherwise influence each process. But this comparison does allow us to gain some feeling of the timespans needed for different structural or textural features to develop.

One of the immediate conclusions of this study is that many processes operating in magmatic arcs are relatively rapid: crystal growth, pluton emplacement, cleavage development, etc., occur over time spans of tens of thousands to no more than a few million years. Because these time spans are shorter than, or similar to, most errors on isotopic ages, the exact duration and timing of different processes will remain unclear until dating methods are further refined.

However, we maintain that the durations needed to complete different processes vary significantly: (1) short durations are needed to grow crystals to observed sizes and to supply magma by conduit flow; (2) intermediate durations are needed for the diapiric ascent and crystallization of magmas; and (3) long durations are needed to get large displacements in fault zones, for plutons to cool to ambient country rock temperatures, and particularly to form well-developed cleavages at strain rates of 10^{-14} s^{-1}. These time spans will approach one another if strain rates or fault displacements are faster, or if rates of pluton ascent and crystallization are slower.

Rate-dependent characteristics of pluton–wall rock systems

Paterson *et al.* (1989b) summarized criteria for recognizing pre-, syn- and post-tectonic plutons. The effect of variable rates of processes, however, was not considered. For example, one criterion suggested in support of syn-tectonic emplacement is evidence that melt was present during regional deformation. These features would only be preserved if the rates of the regional processes (cleavage development, folding, faulting) were comparable to rates of magma emplacement and crystallization. At shallow crustal levels, plutons may crystallize so quickly that evidence of concurrent regional deformation and migration of a melt phase might not be preserved. At deeper levels the opposite might occur. Magma could be emplaced prior to regional deformation and then crystallize so slowly that it preserves evidence of melt present during a later regional event.

Hobson & Dellinger (1989) have noted that many of the characteristics of plutons, including apparent timing relationships, change with depth. We have noted that similar changes occur on a batholithic scale: deeper plutons in the Sierra Nevada Batholith are more likely to be deformed and/or appear syn-tectonic (Tobisch *et al.* 1989a). Some of these changes with depth are rate dependent.

The often noted relationship between syn-tectonic plutons and fault zones (e.g. Brun & Pons 1981, Hutton 1982, 1988a,b, Guineberteau *et al.* 1987, Vernon *et al.* 1989) is another case in point. These plutons may appear syn-tectonic because the faster strain rates in fault zones are closer to crystallization rates and would more likely leave an imprint on a rapidly crystallizing pluton. Syn-tectonic plutons emplaced in regions deforming slowly, may crystallize and cool so rapidly that no evidence of the syn-emplacement regional strain is preserved in or near the pluton.

Structural patterns are another feature used to infer timing relationships (e.g. Brun & Pons 1981, Castro 1987, Paterson & Tobisch 1988). Foliation patterns in the aureoles of plutons reflect both the nature of emplacement and regional strain fields and the relative rates of pluton emplacement and regional deformation. These foliation patterns will show the greatest differences at the 'ends' of plutons, that is, the areas where the pluton contact is at high angles to the strike of regional structures (e.g. Paterson & Tobisch 1988). Pre-tectonic plutons may show preserved strain shadows, or emplacement-related foliations that are folded by later regional shortening. Post-tectonic plutons may show aureoles with hornfelsic microstructures that overprint regional structures, or a regional foliation that is folded by emplacement-related deformation. Syn-tectonic plutons may show all three types of relationships, hornfelsic aureoles, regional structures deforming emplacement-related structures, or emplacement-related structures deforming regional structures, depending on the relative rates of different processes. We argue that the cases where a single foliation and cleavage triple points are formed (e.g. Brun & Pons 1981) may be the exception rather than the rule for syn-tectonic plutons, and that this situation implies similar time spans of emplacement-related and regional shortening and relatively rapid rates of pluton cooling and recrystallization in plutons and wall rocks.

Porphyroblast–matrix relationships are another rate-dependent type of timing criteria (Zwart 1962, Bell *et al.* 1986, Vernon 1989). We suggest that the nature of inclusion trials preserved within porphyroblasts, and of prophyroblast–matrix relationships will depend on two factors: (1) the relative rates of porphyroblast growth and matrix deformation; and (2) whether or not the porphyroblasts rotate with respect to geographic coordinates (e.g. Bell 1985). For example, traditional interpretations of spiral inclusion trails in garnet require that a very specific relationship exist between the rate of crystal growth, crystal rotation and matrix deformation over long time spans (e.g. compare Christensen *et al.* 1989 with Bell & Johnson 1989).

A wide range of dating techniques using a variety of minerals with different closure temperatures are increasingly used to constrain timing relationships (e.g. Parrish 1989). Cooling rates affect both the closure temperature and thus the calculated date (Harrison *et al.* 1985, Harrison & Fitzgerald 1986) as well as the interpretation of these dates. For example, U–Pb zircon dates are often used to infer emplacement ages. However, at slow cooling and crystallization rates, continued magmatic and submagmatic flow could continue for millions of years after the magma passed through the U–Pb closure temperature of zircon (>750°C). Caution also must be used in the interpretation of hornblende and biotite ^{40}Ar–^{39}Ar and K–Ar dates from multiple deformed terranes, particularly if those terranes cooled slowly (e.g. Harrison *et al.* 1985, Kligfield *et al.* 1986).

Emplacement mechanisms

Recent studies of pluton emplacement mechanisms point out the need for significant wall rock deformation to make space for the pluton (e.g. Hutton 1988a, Miller *et al.* 1988). It is therefore important to consider the relative rates of deformation in the wall rocks and rates of magma ascent, crystallization and cooling. A consideration of ascent rates of magma (Table 2), or the rates at which a certain volume of magma can be transported to a magma chamber (Crisp 1984, Shaw 1985), indicate that the transport of magma will not be the rate limiting process in pluton emplacement; instead, one can speculate whether or not present models of pluton emplacement require unreasonably rapid rates of wall rock deformation during emplacement?

For example, Hutton (1982, 1988b) has suggested that plutons like the Strontian and Main Donegal are emplaced in dilational jogs in fault zones. One means of evaluating this hypothesis is to determine if fault motion can make space fast enough to accommodate a pluton of a particular size before the pluton would completely crystallize. If one assumes a displacement of 5 cm $year^{-1}$ by homogeneous slip in a fault zone as wide as the presently exposed pluton, then enough space could be made to emplace the Strontian granite in 480,000 years and the Main Donegal granite in 900,000 years. If the fault were not as wide, then the time or the displacement rate needed to emplace the plutons would increase, or other types of deformation in the wall rocks (e.g. simple shear plus flattening) would have to be called upon. Both of these large plutons were emplaced at moderate levels, so rates of cooling would be in the range of millions of years and final crystallization in the range of less than a million years. If the assumptions made above continue to be supported by data collected near these plutons, then this type of emplacement mechanism appears feasible from a standpoint of the rates of processes. This emplacement model does imply, however, that small batches of magma were continuously added, for hundreds of thousands of years at the present level of exposure, as the dilational jog expanded (e.g. Hardee 1982, Lagarde *et al.* 1990).

Another popular emplacement model is that of ballooning or in situ expansion (e.g. Holder 1979, Bateman 1985, Ramsay 1989). Although, there are geometrical and mechanical aspects of this model that need further consideration (e.g. Paterson *et al.* 1989b, 1991), we can evaluate the model purely from a standpoint of rates. One example of such a pluton is the Papoose Flat pluton (Sylvester *et al.* 1978). These authors suggest that the change in thickness of stratigraphy around the pluton records −90% shortening. At a strain rate of 10^{-14} s^{-1}, −90% shortening would take 7.3 million years. This is too great a time given that this pluton was emplaced at shallow levels (Sylvester *et al.* 1978) and was likely emplaced and crystallized in less than 1 Ma (e.g. Barton *et al.* 1988). However, if a strain rate of 10^{-13} s^{-1} is used, −90% shortening would occur in 700,000 years, an appropriate length of time to emplace and cool this pluton. We should note, however, that even −90% shortening, given the width of the aureole, is not sufficient shortening to make the necessary space for this pluton. Our calculations suggest that only about 21% of the needed space can be accommodated by this strain. Additional shortening, and thus faster strain rates, or other mechanisms of deformation also must occur.

RATES OF PROCESSES IN FOOTHILLS TERRANE, SIERRA NEVADA

In previous papers, we have described detailed work on the geology and geochronology of a suite of Late Jurassic plutons and their wall rocks located in the Foothills Terrane, central Sierra Nevada, California (Paterson *et al.* 1989a, Saleeby *et al.* 1989, Tobisch *et al.* 1989b). These plutons are thought to represent the roots of a Late Jurassic arc and provide an interesting case study of rates of processes in an arc setting. Table 4 lists the approximate duration of events (or processes) in the Foothills Terrane.

The Late Jurassic plutons generally lack magmatic foliations, have narrow to no contact aureoles, and are texturally and sometimes spacially associated with volcanic rocks. Elsewhere we have argued that these plutons were emplaced at shallow levels during active deformation of the wall rocks (Paterson *et al.* 1991b). These features also suggest that the plutons were emplaced as magma with >50% melt and crystallized and cooled relatively rapidly. In fact, one pluton has a concordant U–Pb zircon age of 147 +2/−1 Ma and a ^{40}Ar–^{39}Ar hornblende age of 146.6 ± 0.03 Ma (Saleeby *et al.* 1989) indicating the the pluton probably was emplaced, crystallized, and cooled below 500°C in 1 Ma or less.

A comparison of nearby U–Pb zircon dates and ^{40}Ar–^{39}Ar hornblende and biotite dates throughout the region suggest that plutons generally cooled rapidly. Our calculations (based on comparisons of U–Pb and ^{40}Ar–^{39}Ar dates) show that cooling rates of at least 75–250°C per million years occurred during and immediately after pluton emplacement.

Strains associated with a widespread cleavage indicate that the wall rocks underwent about −50% shortening (Paterson *et al.* 1989a) after emplacement of the Late Jurassic plutons and before emplacement of a suite of Early Cretaceous plutons. If the cleavage formed continuously from about 151 to 123 Ma (the oldest post-cleavage pluton), then a strain rate of 10^{-15} s^{-1} could account for the measured strains. Our structural and geochronological studies indicate, however, that much of the strain occurred between 137 and 145 Ma, suggesting faster strain rates. If the strains formed at a strain rate of 10^{-14} s^{-1}, only 2–3 million years was needed to form the cleavage. At a strain rate of 10^{-13} s^{-1}, the strain and cleavage could form in less than 500,000 years.

CONCLUSIONS

It is now well known that arc magmatism often occurs in tectonically active regions (e.g. see abstracts in Structure and Tectonics of Volcanic Arcs, *Eos* **70**, 1299–1300, 1307–1309). Hutton (1988a) among others also has argued that most plutons are emplaced during active deformation of the wall rocks. We contend, therefore, that it is crucial to evaluate the rates of different magmatic and structural processes when trying to understand the structural evolution of arcs and associated pluton–wall rock systems.

In this paper, we have made an initial attempt to compare the rates of faulting, cleavage development, pluton ascent, magma supply and crystallization, and pluton cooling in an active arc environment. Our best estimates

Table 4. Summary of the duration of structural events in Foothills Terrane, Sierra Nevada, California

Process	Approximate duration
Rigid rotation of beds and initial large-scale folding	160–145 Ma
Emplacement of Jurassic plutons	151–137 Ma
Movement on Bear Mountains fault zone	Before 155 Ma to 123 Ma
Development of regional cleavage and metamorphism	146–123 Ma
Emplacement of Cretaceous plutons and deformation in aureoles	123–110 Ma
Development of spaced cleavages and kinks	Cretaceous or younger

of average rates or durations of processes, listed in the abstract, indicate that many processes operating in magmatic arcs are relatively rapid: pluton emplacement and cleavage development can occur over time spans of tens of thousands to no more than a few million years. In this context, however, the ascent and crystallization of plutons at shallow to moderate crustal levels generally occurs over shorter time spans than that needed to get large displacements on faults or widespread cleavages developed during strain rates of 10^{-14} s^{-1} or slower.

A comparison of these rates also suggests that the flow of wall rocks is limiting process during emplacement of plutons, unless fast strain rates of multiple deformation mechanisms are considered. Emplacement mechanisms that rely on the transport of magma along fractures and the addition of numerous small pulses of magma over hundreds of thousands to millions of years, however, avoid this limitation and need further consideration. The obvious disadvantage of using this mechanism for magma emplacement is that the rate of cooling and crystallization would be faster.

Another problem emphasized by these rates is that older structures can be rapidly overprinted, and if deformation is intense enough, it may wipe out evidence of earlier regional or emplacement-related deformation. Our calculations suggest that new mylonitic foliations often form in less than 1 million years in fault zones and sometimes form in much shorter time spans. A natural example where this occurs is the Alpine fault zone, New Zealand (Allis 1981, Johnston & White 1983, Rattenbury *et al.* 1988). Shifting plate motions have recently caused the Alpine fault to change from a strike-slip fault to an oblique-slip fault. Even at the surface, structures have changed within the last few million years to reflect this oblique-slip motion (Allis 1981, Johnston & White 1983, Rattenbury *et al.* 1988). At depth where temperatures are higher, this change may occur over an even shorter length of time.

Lastly, we are struck by the often lengthy time attributed to deformation in arcs by geochronology (e.g. Table 4), in comparison to the much shorter time it takes structures to theoretically develop at any one point (Fig. 2 and Table 3). This disparity raises many questions, among them being the relationship between ongoing subduction and structures produced in the upper plate. For example, we believe from U–Pb and ^{40}Ar–^{39}Ar data that the total duration of regional cleavage formation and metamorphism in the Foothills Terrane spanned about 23 Ma (Table 4), while the cleavage could theoretically form in 2 Ma (at a strain rate of 10^{-14} s^{-1}). What structures, then, if any, developed during the remaining 21 Ma as subduction presumably continued? One possibility is that the rates of strain varied dramatically in different domains through time. In this scenario, the pervasively foliated rocks we now see developed over large tracts may have sequentially developed cleavage in numerous small domains relatively quickly (2 Ma), with the domains gradually coalescing to form a regional terrain of foliated rock over a longer time span (23 Ma). Such diachronous partitioning of deformation is probably the norm in orogenic belts (e.g. Bell 1985, Tobisch & Paterson 1988), and more detailed isotopic and structural studies are needed to test the relationships of timing and rates of processes.

Acknowledgements—We thank Eli Silver for calculating the spreading rate shown in Fig. 1, Mark Brandon for discussions about strain rates, and Jim Gill, Maria Jackson, and an anonymous reviewer for suggestions about the manuscript. This research was supported by NSF grants EAR 860717, EAR 8805063 and INT 8611452.

REFERENCES

Allis, R. G. 1981. Continental underthrusting beneath the Southern Alps of New Zealand. *Geology* **9**, 303–307.

Anderson, D. L. 1981. Rise of deep diapirs. *Geology* **9**, 7–9.

Arzi, A. A. 1978. Critical phenomena in the rheology of partially melted rocks. *Tectonophysics* **44**, 173–184.

Barton, M. D., Battles, D. A., Debout, G. E. Capo, R. C. Christensen, J. N., Davis, S. R., Hanson, R. B., Michelsen, C. J. & Trim, H. E. 1988. Mesozoic contact metamorphism in the western United States. In: *Metamorphism and Crustal Evolution of the Western United States, Rubey Volume* 7 (edited by Ernst, W. D.). Prentice-Hall, Englewood Cliffs, New Jersey, 110–178.

Bateman, R. 1985. Aureole deformation by flattening around a diapir during *in situ* ballooning: the Cannibal Creek granite. *J. Geol.* **93**, 293–310.

Bell, T. H. 1985. Deformation partitioning and porphyroblast rotation in metamorphic rocks: a radical reinterpretation. *J. metamorph. Geol.* **3**, 109–118.

Bell, T. H., Fleming, P. D. & Rubensch, M. J. 1986. Porphyroblast nucleation, growth, and dissolution in regional metamorphic rocks as a function of deformation partitioning during foliation development. *J. metamorph. Geol.* **4**, 37–67.

Bell, T. H. & Johnson, S. E. 1989. Porphyroblast inclusion trails: the key to orogenesis. *J. metamorph. Geol.* **7**, 279–310.

Brun, J. P & Pons, J. 1981. Strain patterns of pluton emplacement in crust undergoing non-coaxial deformation, Sierra Morena, southern Spain. *J. Struct. Geol.* **8**, 219–230.

Cashman, K. V. 1990. Textural constraints on the kinetics of crystallization of igneous rocks. In: *Modern Methods of Igneous Petrology: Understanding Magmatic Processes* (edited by Nicholls, J. & Russell, ?). *Miner. Soc. Am., Rev. Miner.* **24**, 259–314.

Cashman, K. V. & Ferry, J. M. 1988. Crystal size distribution (CSD) in rocks and the kinetics and dynamics of crystallization, III. Metamorphic crystallization. *Contr. Miner. Petrol.* **99**, 405–415.

Castro, A. 1987. On granitoid emplacement and related structures. A review. *Geol. Rdsch.* **76**, 101–124.

Christensen, J. N., Rosenfeld, J. L. & Depaolo, D. J. 1989. Rates of tectonometamorphic processes from matrix 87Sr/86Sr evolution recorded radially by garnet. *Geol. Soc. Am. Abs. w. Prog.* **21**, A27.

Cloos, E. 1949. Oolite deformation in the South Mountain fold, Maryland. *Bull. geol. Soc. Am.* **58**, 843–918.

Corry, C. E. 1988. Laccoliths; mechanics of emplacement and growth. *Spec. Pap. geol. Soc. Am.* **220**.

Crisp, J. A. 1984. Rates of magma emplacement and volcanic output. *J. Volcan. Geotherm. Res.* **20**, 177–211.

Cruden, A. 1990. Flow and fabric development during the diapiric rise of magma. *J. Geol.* **98**, 681–698.

Dallmeyer, R. D. & Van Breeman, O. 1981. Rb–Sr whole-rock and ^{40}Ar/^{39}Ar mineral ages of the Togus and Hallowell Quartz Monzonite and Three Mile Pond Granodiorite plutons, south-central Maine: their bearing on post-Acadian cooling history. *Contr. Miner. Petrol.* **78**, 61–73.

Decker, R. W., Wright, T. L. & Stauffer, P. H. 1987. Volcanism in Hawaii: Volume 2: Structure, dynamics, and history of investigations of Hawaiian volcanism. *Prof. Pap. U.S. geol. Surv.* **1350**.

Delaney, P. T. & Pollard, D. D. 1981. Deformation of host rocks and flow of magma during growth of minette dikes and breccia-bearing intrusions near Ship Rock, New Mexico. *Prof. Pap. U.S. geol. Surv.* **1202**.

Delaney, P. T. & Pollard, D. D. 1982. Solidification of basaltic magma during flow in a dike. *Am. J. Sci.* **282**, 856–885.

Eaton, J. P. 1962. Crustal structure volcanism in Hawaii. In: *The Crust of the Pacific Basin* (edited by MacDonald, G. A. & Kuno, H.). *Am. Geophys. Un. Geophys. Monogr.* **6**, 13–29.

Ekstrom, G. & Engdahl, E. R. 1989. Earthquake source parameters and stress distribution in the Adak Island region of central Aleutian islands, Alaska. *J. geophys. Res.* **94**, 15499–15519.

Fedotov, S. A. 1981. Magma rates in feeding conduits of different volcanic centres. *J. Volcan. Geotherm Res.* **9**, 379–394.

Gill, J. B. 1981. *Orogenic Andesites and Plate Tectonics*. Springer, New York.

Guineberteau, B., Bouchez, J. L. & Vigneresse, J. L. 1987. The Mortagne granite pluton (France) emplaced by pull-apart along a shear zone: structural and gravimetric arguments and regional implications. *Bull. geol. Soc. Am.* **99**, 763–770.

Hacker, B. R., Yin, An, Christie, J. M. & Snoke, A. W. 1990. Differential stress, strain rate, and temperature of mylonitization in the Ruby Mountains, Nevada: implications for the rate and duration of uplift. *J. geophys. Res.* **95**, 8569–8580.

Hamilton, W. 1979. Tectonics of the Indonesian region. *Prof. Pap. U.S. geol. Surv.* **1078**.

Hardee, H. C. 1982. Incipient magma chamber formation as a result of repetitive intrusions. *Bull. Volcanol.* **45**, 41–49.

Harper, G. D., Saleeby, J. B., Pessagno, E. A., Jr. & Heizler, M. 1989. The Josephine ophiolite: generation, translation, and emplacement of Late Jurassic paraauthochthonous back-arc lithosphere, Klamath Mountains, California–Oregon. *Geol. Soc. Abs. w. Prog.* **21**, A28.

Harrison, T. M. & Clarke, G. K. C. 1979. A model of the thermal effects of igneous intrusion and uplift as applied to Quottoon pluton, British Columbia. *Can. J. Earth. Sci.* **16**, 411–420.

Harrison, T. M., Duncan, I. J. & McDougall, I. 1985. Diffusion of ^{40}Ar in biotite: Temperature, pressure and compositional effects. *Geochim. cosmochim. Acta* **49**, 2461–2468.

Harrison, T. M. & Fitzgerald, J. D. 1986. Exsolution in hornblende and its consequences for ^{40}Ar/^{39}Ar age spectra and closure temperature. *Geochim. cosmochim. Acta* **50**, 247–253.

Harrison, T. M. & McDougall, I. 1980. Investigations of an intrusive contact, northwest Nelson, New Zealand—I: Thermal, chronological and isotopic constraints. *Geochim. cosmochim. Acta* **44**, 1985–2004.

Hibbard, M. J. 1987. Deformation of incompletely crystallized magma systems: granitic gneisses and their tectonic implications. *J. Geol.* **95**, 543–561.

Hobson, C. A. & Dellinger, D. A. 1989. Depth dependency of granite types, illustrated by tilted plutons in the North Cascades, Washington. *Geol. Soc. Am. Abs. w. Prog.* **21**, 95.

Holder, M. T. 1979. An emplacement mechanism for post-tectonic granites and its implications for their geochemical features. In: *Origin of Granite Batholiths, Geochemical Evidence* (edited by Atherton, M. P. & Tarney, J.). Shiva, Nantwich, 116–128.

Hutton, D. H. W. 1982. A tectonic model for the emplacement of the Main Donegal Granite, NW Ireland. *J. geol. Soc. Lond.* **139**, 287–294.

Hutton, D. H. W. 1988a. Granite emplacement mechanisms and tectonic controls: inferences from deformation studies. *Trans. R. Soc. Edinb.* **79**, 245–255.

Hutton, D. H. W. 1988b. Igneous emplacement in a shear zone termination: the biotite granite at Strontian, Scotland. *Bull. geol. Soc. Am.* **100**, 1392–1399.

Jackson, M. P. A. & Talbot, C. J. 1986. External shapes, strain rates, and dynamics of salt structures. *Bull. geol. Soc. Am.* **97**, 305–323.

Jaeger, J. C. 1968. Cooling and solidification of igneous rocks. In: *Basalts: the Poldervaart Treatise on Rocks of Basaltic Composition, Vol.* 2 (edited by Hess, H. H. & Poldervaart, A.). Interscience, New York, 504–536.

Joesten, R. & Fisher, G. 1988. Kinetics of diffusion-controlled mineral growth in the Christmas Mountains (Texas) contact aureole. *Bull. geol. Soc. Am.* **100**, 714–732.

Johnston, D. C. & White, S. H. 1983. Shear heating associated with movement along the Alpine Fault, New Zealand. *Tectonophysics* **92**, 241–252.

Kligfield, R., Carmignani, L. & Owens, W. H. 1981. Strain analysis of a Northern Apennine shear zone using deformed marble breccias. *J. Struct. Geol.* **3**, 421–436.

Kligfield, R., Hunziker, J., Dallmeyer, R. D. and Schamel, S. 1986. Dating of deformation phases using K–Ar and ^{40}Ar/^{39}Ar techniques: results from the Northern Apennines. *J. Struct. Geol.* **8**, 781–798.

Lagarde, J.-L., Brun, J.-P. & Gapais, D. 1990. Formation of epizonal granitic plutons by in situ assemblage of laterally expanding magma. *C. r. Acad. Sci., Paris Ser. II,* **310**, 1109–1114.

Mahon, K. I., Harrison, T. M. & Drew, D. A. 1988. Ascent of granitoid diapir in a temperature varying medium. *J. geophys. Res.* **93**, 1174–1188.

Marsh, B. D. 1978. On the cooling of ascending andesitic magma. *Phil. Trans. R. Soc. Lond.* **A288**, 611–625.

Marsh, B. D. 1981. On the crystallinity, probability of occurrence, and rheology of lava and magma. *Contr. Miner. Petrol.* **78**, 85–98.

Marsh, B. D. 1982. On the mechanics of igneous diapirism, stoping, and zone melting. *Am. J. Sci.* **282**, 808–855.

McCaffrey, R., Bock, Y. & Rais, J. 1990. Crustal deformation and oblique plate convergence in Sumatra. *Trans. Am. Geophys. Un.* **71**, 37.

Miller, C. F., Watson, E. B. & Harrison, T. M. 1988. Perspectives on the source segregation and transport of granitoid magmas. *Trans. R. Soc. Edinb., Earth Sci.* **79**, 135–156.

Naar, D. F. & Hey, R. N. 1989. Speed limit for oceanic transform faults. *Geology* **17**, 420–423.

Parrish, R. R. 1989. Rates of tectonic processes from U–Pb geochronology. *Geol. Soc. Am. Abs. w. Prog.* **21**, A27.

Paterson, M. S. 1987. Problems in the extrapolation of laboratory rheological data. *Tectonophysics* **133**, 33–43.

Paterson, S. R. 1988. Creek granite: Post-tectonic "ballooning" intrusion or pre-tectonic piercement diapir? *J. Geol.* **96**, 730–736.

Paterson, S. R. & Tobisch, O. T. 1988. Using Pluton ages to date regional deformations: problems with commonly used criteria. *Geology* **16**, 1108–1111.

Paterson, S. R., Tobisch, O. T. & Bhattacharyya, T. 1989a. Regional structural and strain analyses of terranes in the Western Metamorphic belt, central Sierra Nevada, California. *J. Struct. Geol.* **11**, 255–273.

Paterson, S. R., Tobisch, O. T. & Vernon, R. H. 1991a. Deformation and pluton emplacement during volcanic arc construction in the Foothills Terrane, central Sierra Nevada, California. *Tectonophysics* **191**, 89–110.

Paterson, S. R., Vernon, R. H., Brudos, T. C., Fowler, K., Carlson, C. & Bishop, K. 1991b. The Papoose Flat pluton: forceful expansion or post-emplacement deformation? *Geology* **19**, 324–327.

Paterson, S. R., Vernon, R. H. & Tobisch, O. T. 1989b. A review of criteria for the identification of magmatic and tectonic foliations in granitoids. *J. Struct. Geol.* **11**, 349–363.

Pfiffner, O. A. & Ramsay, J. G. 1982. Constraints on geologic strain rates: arguments from finite strain states of naturally deformed rocks. *J. geophys. Res.* **87**, 311–321.

Price, N. J. 1975. Rates of deformation. *J. geol. Soc. Lond.* **131**, 553–575.

Price, R. A. 1981. The Cordilleran foreland thrust and fold belt in the Southern Canadian Rocky Mountains. In: *Thrust and Nappe Tectonics* (edited by McClay, K. & Price, N. J.). *Spec. Publs geol. Soc. Lond.* **9**, 427–448.

Ramsay, J. G. 1967. *Folding and Fracturing of Rocks*. McGraw-Hill, New York.

Ramsay, J. G. 1989. Emplacement kinematics of a granite diapir: the Chindamora batholith, Zimbabwe. *J. Struct. Geol.* **11**, 191–210.

Rattenbury, M. S., Cooper, A. F. & Norris, R. J. 1988. Recent Alpine Fault thrusting at Kaka Creek, central Westland, New Zealand. *N.Z. J. Geol. Geophys.* **31**, 117–120.

Relinger, R., Oliver, J. Brown, L., Sanford, A. & Balazs, E. 1980. New measurements of crustal doming over the Socorro magma body, New Mexico. *Geology* **8**, 291–295.

Ridley, J. 1986. Modeling of the relations between reaction enthalpy and the buffering of reaction progress in metamorphism. *Mineralog. Mag.* **50**, 375–384.

Rockwell, T. K., Keller, E. A. & Dembroff, G. R. 1988. Quarternary rate of folding of the Ventura Avenue anticline, western Transverse Ranges, southern California. *Bull. geol. Soc. Am.* **100**, 850–858.

Saleeby, J. B., Geary, E. E., Paterson, S. R. & Tobisch, O. T. 1989. Isotopic systematics of U/Pb (zircon) and 40Ar/39Ar (biotite/hornblende) from rocks of the central Foothills Terrane, Sierra Nevada, California. *Bull. geol. Soc. Am.* **101**, 1481–1492.

Sarawitz, D. 1988. High rates of late Cenozoic crustal shortening in the Andean foreland, Mendoza Province, Argentina. *Geology* **18**, 1138–1142.

Savage, W. Z. 1974. Stress and displacement fields in stably folded rock layers. Unpublished Ph.D. thesis, College Station, Texas A&M University.

Schmid, S. M. 1989. Episodes in Alpine orogeny. *Geol Soc. Am. Abs. w. Prog.* **21**, A28.

Seni, S. J. & Jackson, M. P. A. 1984. Sedimentary record of Cretaceous and Tertiary salt movement. East Texas Basin: Times, rates, and volumes of flow and their implications to nuclear waste isolation

and petroleum exploration. *Univ. of Texas of Austin Bur. Econ. Geol. Rep. Invest.* **139**.
Shaw, H. R. 1965. Comments on viscosity, crystal settling, and convection in granitic magmas. *Am. J. Sci.* **263**, 120–152.
Shaw, H. R. 1980. The fracture mechanisms of magma transport from the mantle to the surface. In: *Physics of Magmatic Processes,* 1980 (edited by Hargraves, R. B.). Princeton University Press, Princeton, New Jersey.
Shaw, H. R. 1985. Links between magma–tectonic rate balances, plutonism, and volcanism. *J. geophys. Res.* **90**, 11,275–11,288.
Sibson, R. H. 1986. Earthquakes and rock deformation in crustal fault zones. *Annu. Rev. Earth & Planet. Sci.* **14**, 149–175.
Siddans, A. W. B., Henry, B., Kligfield, R., Lowrie, W., Hirt, A. & Percevault, M. N. 1984. Finite strain patterns and their significance in Permian rocks of the Alps Maritimes (France). *J. Struct. Geol.* **6**, 399–368.
Sieh, K. E. & Williams, P. L. 1990. Behavior of the southernmost San Andreas fault during the past 300 years. *J. geophys. Res.* **95**, 6629–6646.
Spera, F. J. 1980. Aspects of magma transport. In: *Physics of Magmatic Processes.* 1980 (edited by Hargraves, R. B.). Princeton University Press, Princeton, New Jersey.
Spera, F. J., Yucn, D. A., Greer, J. & Sewell, G. 1985. Magma withdrawal from compositionally zoned magma chambers (Abs.). *Eos* **66**, 397.
Stern, T. W., Bateman, P. C., Morgan, B. A., Newell, M. F. & Peck, D. L. 1981. Isotopic U–Pb ages from zircon from the granitoids of the central Sierra Nevada, California. *Prof. Pap. U.S. geol. Surv.* **1185**.
Suppe, J. 1989. Rates of fault slip and fault growth. *Geol. Soc. Am. Abs. w. Prog.* **21**, A28.
Swanson, D. A. 1972. Magma supply rate at Kilauea Volcano, 1952–1971. *Science* **175**, 169–170.
Swanson, S. E. 1977. Relation of nucleation and crystal-growth rate to the development of granitic textures. *Am. Miner.* **62**, 966–978.
Sylvester, A. G., Oertel, G. Nelson, C. A. & Christie, J. M. 1978. Papoose Flat pluton: a granitic blister in the Inyo Mountains, eastern California. *Bull. geol. Soc. Am.* **89**, 1205–1219.
Talbot, C. J. & Jarvis, A. 1984. Age, budget and dynamics of an active salt extrusion in Iran. *J. Struct. Geol.* **6**, 521–533.
Tobisch, O. T. 1984. Development of cleavage in lapilli-bearing volcaniclastic rock. *Tectonophysics* **109**, 309–335.
Tobisch, O. T., Fiske, R. S., Sacks, S. & Taniguchi, D. 1977. Strain in metamorphosed volcaniclastic rocks and its bearing on the evolution of orogenic belts. *Bull. geol. Soc. Am.* **88**, 23–40.
Tobisch, O. T. & Paterson, S. R. 1988. Analysis and interpretation of composite foliations in areas of progressive deformation. *J. Struct. Geol.* **10**, 745–754.
Tobisch, O. T., Paterson, S. R. & Saleeby, J. B. 1989a. Magmatic fabrics in Late Cretaceous granitoids, central Sierra Nevada, (CSN) California: implications for Late Mesozoic tectonic models. *Geol. Soc. Am. Abs. w. Prog.* **21**, 150.
Tobisch, O. T., Paterson, S. R., Saleeby, J. B. & Geary, E. E. 1989b. Nature and timing of deformation in the Foothills terrane, central Sierra Nevada, California. *Bull. geol. Soc. Am.* **101**, 401–413.
Tobisch, O. T., Saleeby, J. B. & Fiske, R. S. 1986. Structural history of continental volcanic arc rocks, eastern Sierra Nevada, California: a case for extensional tectonics. *Tectonics* **5**, 64–94.
Van der Molen, I. & Paterson, M. S. 1979. Experimental deformation of partially melted granite. *Contr. Miner. Petrol.* **70**, 299–318.
Vernon, R. H. 1978. Porphyroblast–matrix microstructural relationships in deformed metamorphic rocks. *Geol. Rdsch.* **67**, 288–305.
Vernon, R. H. 1989. Porphyroblast–matrix microstructural relationships—recent approaches and problems. In: *Evolution of Metamorphic Belts* (edited by Daly, J. S., Cliff, R. A. & Yardley, B. W. D.). *Spec. Publs geol. Soc. Lond.* **42**.
Vernon, R. H., Paterson, S. R. & Geary, E. E. 1989. Evidence for the syntectonic intrusion of plutons in the Bear Mountains fault zone, California. *Geology* **17**, 723–726.
Vita-Finzi, C. 1979. Rates of Holocene folding in the coastal Zagros near Bandar Abbas, Iran. *Nature* **278**.
Wadge, G. 1981. The variation of magma discharge during basaltic eruptions. *J. Volcan. Geotherm. Res.* **11**, 139–168.
Walther, J. V. and Wood, B. J. 1984. Rate and mechanism in prograde metamorphism. *Contr. Miner. Petrol.* **88**, 246–259.
Wickham, S. M. 1987. The segregation and emplacement of granitic magmas. *J. geol. Soc. Lond.* **144**, 281–297.
Wood, D. S. 1974. Current views on the development of slaty cleavage. *Annu. Rev. Earth & Planet. Sci.* **2**, 369–401.
Zwart, H. J. 1962. On the determination of polymetamorphic mineral associations and its application to the Bosost area (Central Pyrenees). *Geol. Rdsch.* **52**, 38–65.