PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Physical Constraints on Magma Contamination in the Continental Crust: An Example, the Adamello Complex [and Discussion]

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Physical constraints on magma contamination in the continental crust: an example, the Adamello complex

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Igneous intrusions may move upwards through the crust by zone melting, by penetrative intrusion, by stoping or by some combination of these three mechanisms. Each mechanism offers different opportunities for contamination of the magma by country rock. Both zone melting and stoping offer the greatest possibility of assimilation, but in most natural situations the maximum amount of country rock assimilable by a magma is considerably less than its own original volume. Thermal and other constraints limit the amount of ascent that a magma body may accomplish by either zone melting or stoping: once, and twice to three times the original height of the magma body respectively.

The assimilability of a xenolith sinking in a magma (i.e. the possibility of reaching the bottom without fully melting) depends on the fourth power of the radius r of the xenolith because the time required for assimilation increases as r^2 and the time available decreases as r^2 (because of increased sinking velocity). This can explain the observed size distribution of xenoliths in some intrusions.

Applied to the Tertiary Adamello igneous complex of northern Italy, these considerations suggest that the intrusion may have been initiated by the emplacement of a mafic magma body in the lower crust. The body remained gravitationally stable until its composition had been modified and its density so lowered by zone melting of its roof that it began to ascend through the crust by either penetrative or stoping processes. The intrusion finally solidified at a depth of between six and ten kilometres and the last stages of emplacement occurred by stoping. The complex comprises a number of separate intrusions and this process was repeated seven or more times over a 10 Ma period (between 30 and 40 Ma). Each episode of intrusion, however, lasted less than 1 Ma.

Introduction

This paper is primarily concerned with the physical constraints on the chemical interaction between magma and the continental crust through which it ascends until it comes to rest, and the extent to which its composition may be modified by these processes. We do not consider the process of magma generation and segregation.

We assume a coherent and large (say many tens of cubic kilometres) body of magma situated in the lower part of the crust. The magma is taken to be hotter and less dense than the country rock although, as discussed later, the latter assumption will not hold true in every case.

The ascent of the magma body is driven by the density difference between it and the country rock, and is resisted by the viscosity of the magma and the strength of the country rock. During the ascent the viscosity of the magma is likely to increase as the temperature and pressure fall (Shaw 1965; Kushiro *et al.* 1976; Kushiro 1978) and the intrusion must come to rest at a height where the viscosity of the magma and the strength of the country rock overcome the buoyancy forces. The rate of heat loss during the ascent – and thus the rate of increase of viscosity – depends both on the rate of ascent and on the temperature and temperature gradient in the country rock. In the limit, if the ascent is fast enough there is effectively no time for conductive heat transfer between the magma and the country rock. It is therefore to be expected that, other things being equal, the faster moving intrusions travel farthest.

These and related problems have been considered by Pitcher (1978, 1979) and Marsh (1982).

Chemical interaction between the magma and its surroundings during emplacement

In general magmas will have an initial composition different from that of the country rocks with which they are in contact. It will often, but not always, be the case that the magma is at a temperature close to or higher than that at which the country rock begins to melt.

Chemical transfer between the country rock and the magma may take place either by diffusion of components across the rock–magma interface or by partial melting of the country rock. The available observational and experimental evidence suggests that solid state diffusion is unlikely to be very important (Hofmann 1980). We therefore concentrate here on melting of new material as the main means of modification of the composition of the magma. Melting of country rock may occur either round the walls of the magma body or within the magma body where stoped blocks of country rock have been incorporated within it.

From a thermal, as distinct from mechanical, point of view, we distinguish three main classes of intrusive process and assess the extent to which they may be expected to modify magma composition. The three processes are *penetrative intrusion*, *stoping* and *zone melting*. In the first, the only mass transfer is the movement of magma with respect to the country rocks; in the second there is mass transfer of the magma upwards in the crust and a compensating downward movement of the crust by continuous collapse of the magma chamber roof and the sinking of the collapsed blocks through the magma. In the third the magma body makes its way upwards through the crust by melting the roof and simultaneously precipitating refractory phases. Probably all three processes contribute to some degree to the migration of every intrusion. However, they are competing processes and, as discussed below, it is to be expected that in any particular case circumstances will favour one process over another and it will dominate. Equally, one process may take over from another at different stages in the crustal ascent.

Zone melting

This process was first proposed by Harris (1957) as one by which a magma might ascend through the crust and modify its composition as it did so. A body of magma having a temperature higher than that at which the country rock melts is thought to melt the roof rocks above and to crystallize in its interior and at its base, possibly also precipitating phases that settle to the magma chamber floor. In this way both the top and bottom of the magma body advance upward through the crust (figure 1). In part, the latent heat of melting at the top is balanced by the latent heat of crystallization at the base. This assumes that magma body remains well mixed.

As Harris pointed out, this process is analogous to the zone-refining of metals, in that the magma becomes progressively more enriched in those elements that tend to be partitioned preferentially into the liquid phase and the composition changes continuously as the intrusion advances.

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The thermal aspects of this process were examined quantitatively by Ahern *et al.* (1981). They showed that the effectiveness of zone melting largely depends on the initial height of the magma chamber, and the amount by which the initial magma temperature exceeds the melting temperature of the country rock. The process is also affected by the thermal gradient and the solidus gradient in the country rock. Ahern *et al.* showed, however, that even in improbably favourable circumstances (i.e. no lateral loss of heat) this process was at best likely to allow the top of a magma body to migrate upwards by an amount equal to the initial height of the magma body. These same conditions would lead to vigorous thermal convection within the magma.

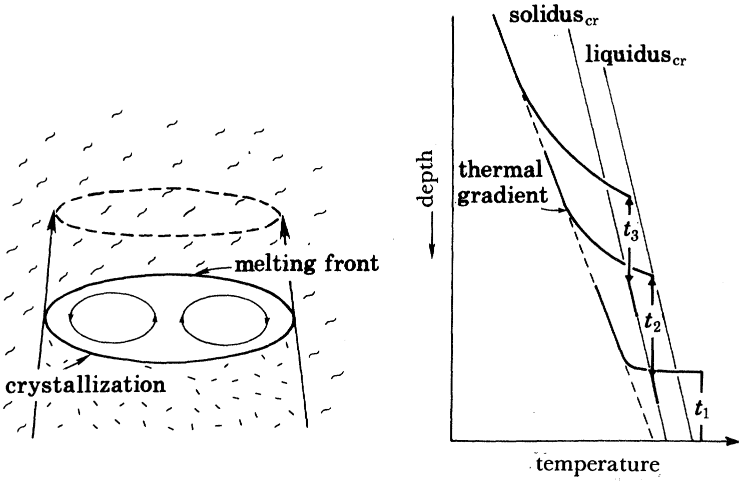

FIGURE 1. Zone melting. A body of magma undergoing internal convection migrates upwards in the crust by melting and assimilating the rock ahead of it and crystallizing at its base. Left side shows schematically initial and final position of such a body. Right side shows crustal temperature at different stages (t_1, t_2, t_3) during zone melting; the thermal gradient is the initial country rock gradient; solidus$_{cr}$ and liquidus$_{cr}$ refer to the country rock. Note that the top and bottom of the magma are maintained at nearly the same temperature by convection. At t_1 the magma body has just been emplaced and is significantly hotter than its surroundings. At t_2 the height of the body has increased because the amount of material assimilated is greater than that crystallized; the temperature at the base of the intrusion (lower arrow) must lie on the solidus$_{cr}$ and that at the top close to liquidus$_{cr}$. At t_3 the body is slowing in its ascent because the temperature at the top has dropped below liquidus. After the molten zone has passed, the temperature below it is, for simplicity, shown to be close to solidus$_{cr}$, but will in fact be an unknown amount above it because the material crystallizing at the base is relatively refractory.

Zone melting clearly has enormous potential for the modification of magma composition; by the time it finally solidifies, the magma may have assimilated a mass of country rock not very different from its own original mass. It should be emphasized that the range of magma compositions that are generated by the zone melting process is different from those that may be produced by simple assimilation of country rock, in so far as zone melting involves the continuous precipitation of refractory phases and enrichment of the liquid in the incompatible elements.

Penetrative intrusion

Under this heading we consider the range of processes that have been termed 'diapirism', 'permissive intrusion', 'forceful injection', etc. These terms carry very different tectonic and mechanical implications; they have in common, however, the fact that magma is thought to penetrate cooler country rocks to which it must lose heat by conduction (figure 2).

The distance a magma may ascend by this mechanism must depend on the balance between the driving forces – buoyancy forces or externally applied surface forces – and the retarding effects of the magma viscosity, and the strength (or 'viscosity') of the country rock.

Mechanically the situation is rather different, depending on whether the penetrative intrusion takes place by the continuous flow of magma along a conduit or whether as one or more discrete pods that displace material above them by some mechanism of macroscopic ductile deformation.

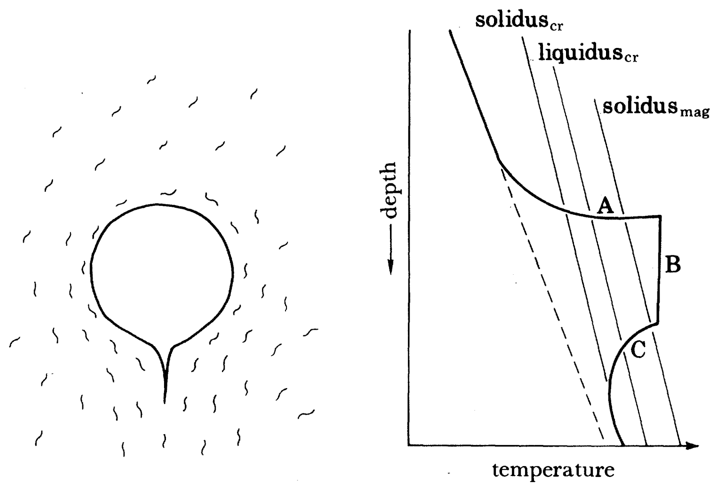

FIGURE 2. Penetrative ascent. Left side: a tear-shaped body of magma ascends through ductile country rock. Immediately round the locus of penetration the country rock must initially be compressed as body approaches, and subsequently be stretched after it has passed. Right side: temperature profile during penetrative intrusion; there is inadequate time for equilibration between magma body and country rock; cr: country rock, mag: magma; A = conductive boundary layer ahead of diapir; B = convecting interior of diapir; C = warmed zone behind diapir.

In either case it is expected that the larger a body happens to be, the farther and faster it is able to move, both because the ratio of buoyant body force to surface drag forces is greater, and because larger bodies cool more slowly and retain a lower viscosity longer.

In general this mechanism of ascent offers the least possibility of contamination of the magma. Chemical interchange will occur only by diffusion across the contacts or by small amounts of melting. In cases where successive bodies of magma ascend by the same path, there is still less opportunity for contamination, as the later batches of magma may ascend by a conduit already coated and pre-heated by preceding intrusions (but see Patchett 1980).

In detail it is rather difficult to know the rate of cooling of a body of magma that penetrates the crust in this way; a variety of local conditions that are peculiar to each particular intrusion is likely to be important. Marsh (1982) has carried out a detailed analysis of some of the possibilities.

A different and simpler approach is sufficient for our purposes here. If pre-heating of the country rock by the passage of a preceding intrusion is ignored, we can consider the conductive cooling of a spherical body of magma as a limiting case. A sphere has the lowest ratio of surface area to volume of any geometrical body and will for that reason cool more slowly than any other body of similar volume; for a given volume, any departure from spherical shape, or cooling by any process other than conduction, will only speed the process.

As the body ascends through the crust it moves down the geothermal gradient to cooler surroundings. The effect of this is to maintain a relatively high rate of cooling during ascent (by comparison with cooling in a static situation where the cooling rate would fall with an inverse root of time dependence): the relation of cooling to solidification would depend on the form of the solidus and liquidus curves. However, in the upper half of the crust both curves should be less steep than the country rock thermal gradient (the solidus even having a negative slope at shallow depth).

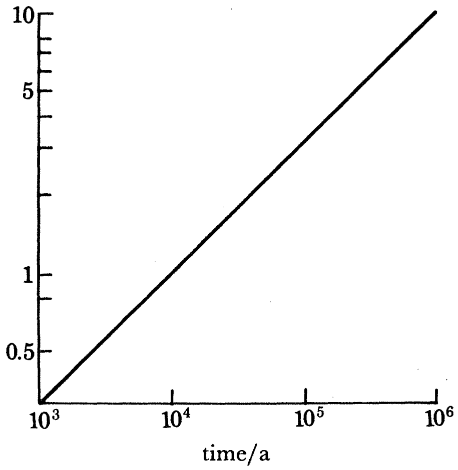

FIGURE 3

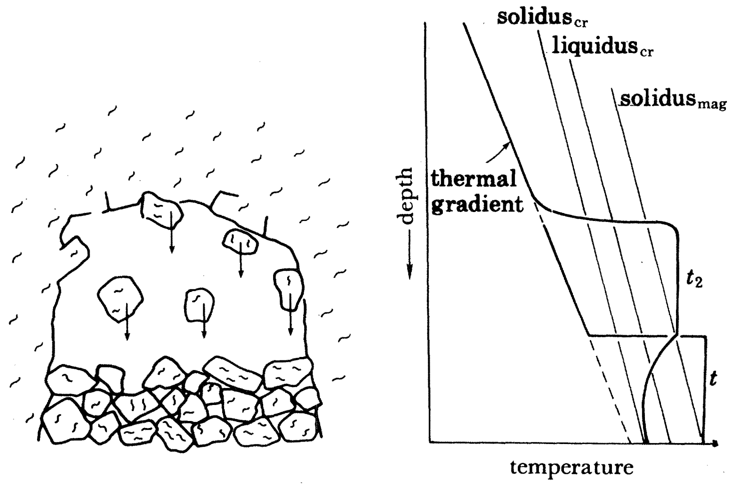

FIGURE 4

FIGURE 3. Maximum times of conductive cooling of a hot spherical body as a function of radius. Times are those required for the centre of the intrusion to cool to 0.8 of the initial temperature. Latent heat of crystallization is accommodated by the approximation of Jaeger (1968). The solution used is that appropriate for a body that does not move with respect to its surroundings; upward movement of the body to cooler regions will speed both cooling and, to a lesser extent, crystallization, as the magma solidus temperature decreases with pressure. The effect of any cooling process other than conduction would be to make cooling faster. Thermal properties as in Jaeger (1968). For full discussion see text.

FIGURE 4. Stoping. Left side: the magma body is thought to move up through the country rock through its inability to support its roof which collapses in a series of subsiding blocks. Right side: abbreviation as in figure 2; the magma cools by conduction and by assimilation as it ascends from t_1 to t_2; note that the temperature below a stoped intrusion must be buffered close to the country rock solidus.

Figure 3 shows the times required for crystallization of spherical igneous bodies of various sizes. The initial conditions assumed are stated in the caption. It is evident that igneous bodies of significant volume, e.g. 4000 km³ (i.e. sphere $r \approx 10$ km) must crystallize in less than one million years. This presumably provides an upper limit to the duration of their ascent through the crust, and to that of all bodies of similar volume unless there is significant pre-heating.

Stoping

In the process of stoping (figure 4) the roof of the magma chamber is thought to collapse into the magma and sink through it, displacing magma upwards as it does so (Daly 1903). The extreme end member of the range of possible stoping processes would be the development of cauldron subsidence and associated ring dyke features, in which the whole roof to the magma chamber sank as a single coherent block, while the magma rose up along its sides. At the other extreme it is possible to imagine shattered country rock sinking as relatively small blocks moving independently of each other. There is abundant observational evidence that stoping is important, at any rate locally, in the upper part of the crust. It clearly requires that the density of the magma be less than that of the roof rocks; in the case of acid and intermediate magmas this is likely to be so in virtually every case, but mafic magmas may have densities close to the solid densities of many crustal rocks.

A magma that ascends by stoping loses heat in two ways. It loses heat by conduction through its roof and walls as do all bodies of magma, but in addition it loses heat to the stoped blocks that sink through it. Having come from higher in the crust, they are invariably cooler than the magma and must extract heat as they pass through it; on the other hand they will to some

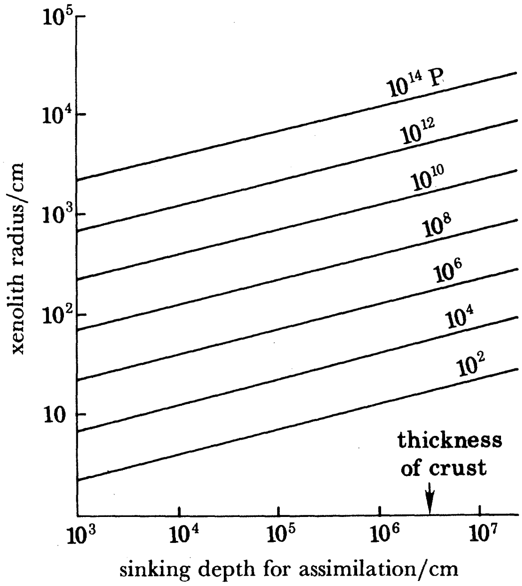

FIGURE 5. Assimilation of xenoliths: the vertical depth that a xenolith of a given radius must sink in magmas of different viscosities (10^{14}–10^2 P) to be fully melted. See text for discussion.

extent have been pre-heated by upward conductive heat loss from the magma chamber. This problem is examined in more detail below in the discussion of the assimilation of xenoliths; however, the cooling effect of sinking blocks largely depends on the ratio of surface area to volume. For equivalent masses of stoped material a large single-stoped block will have a much smaller ratio than an equivalent mass comprising a number of smaller blocks, and will thus extract heat from the magma more slowly (the surface area for conductive heat transfer being smaller). In the limit there is no difference between the two cases, but in magma chambers of finite height, large blocks will tend to reach the bottom without having been fully heated to magma temperature while the small ones will not. The stoping of small blocks therefore is a very effective means of cooling and solidifying an ascending magma.

These same considerations may be applied to the problem of assimilation. A stoped block is available for assimilation only while it is passing through the magma; when it reaches the bottom of the magma chamber it is liable to be buried by succeeding blocks and is relatively unlikely to contribute further to the magma. In that case it is appropriate to compare the time taken for a block of any particular size to sink a particular distance, with the time taken for it to be heated to the temperature of the surrounding magma. The time taken for sinking will in general have an inverse dependence on r^2, where r is the radius of the block, while the time taken for heating will have a direct dependence on r^2.

This means that for a magma chamber of any particular height, the degree of assimilation of a sinking stoped block depends on r^4. This effect is illustrated in figure 5 and means that for magmas of any particular composition (or rather any particular temperature and viscosity) stoped blocks that may be observed today as assimilated xenoliths will have a sharp upper size limit, those that exceeded that size having sunk through the chamber before becoming melted. This phenomenon is discussed further below.

There is a sense in which, from a thermal point of view, there is no sharp distinction between the stoping mechanism and the other two mechanisms previously described. At one extreme, cauldron subsidence represents the stoping of the roof in a single, large, plug-like block; the magma may be considered to make a penetrative ascent along the ring fracture with very little opportunity for contamination. At the other, the stoping of the roof in very small blocks that are capable of rapid assimilation is very close to zone melting of the roof, although the stoping roofs might be expected to migrate upwards more rapidly.

There is also a sense in which the stoping process could be viewed as a relatively inefficient way of transporting magma upwards; depending somewhat on the size and shape of the stoped blocks, they might be expected to accumulate at the base of the magma chamber with a significant amount of 'pore space' between them. This space would necessarily be occupied by magma. It is easy to show that a magma body may migrate upwards by stoping a distance no greater than n times its initial height where $n = (100-\phi)/\phi$ and ϕ is the percentage 'pore space'. If ϕ is 30%, n is 2.3. Ideally the end result of this process might be visualized as a vertical, cylindrical zone in the crust, choked with foundered blocks, the interstices between which were occupied by the original magma.

Convection processes

If thermal convection occurs within the magma body all three of the thermal processes discussed above are affected to some degree. Indeed thermal convection is required if zone melting is to be important. The main consequences of convection are likely to be homogenization both of temperature and composition within the region affected by the flow. This means that unless immiscible liquids are formed by the melting of country rock, the effects of assimilation should become rapidly and uniformly distributed through the magma chamber.

Convection within the magma will in all cases increase the rate of cooling and crystallization by the transfer of heat from the interior of the magma chamber to the margins, where conductive heat losses to the country rock occur.

It is not our main concern here to establish whether or not convection occurs in any particular case. However, we note that this depends on the balance between those factors that promote convection and those that inhibit it, as expressed in the dimensionless ratio known as the Rayleigh number, R.

$$R = g\alpha l^3 \,\Delta T/\kappa\nu,$$

where g is the acceleration due to gravity, α is the coefficient of thermal expansion, l is the height of the magma body, ΔT is the temperature difference between top and bottom, κ is the thermal diffusivity and ν is the kinematic viscosity. In those situations for which $R > 10^3$ natural convection is expected. Evidently R is very sensitive to variation in the height of the magma layer, l. Using reasonable values for all the other quantities suggests that over a wide range of viscosities (Shaw 1965; Kushiro *et al.* 1976; Kushiro 1978) most magma chambers more than a few kilometres high should be affected by convection. In nature convective processes within magma chambers are certainly more complex than assumed above;

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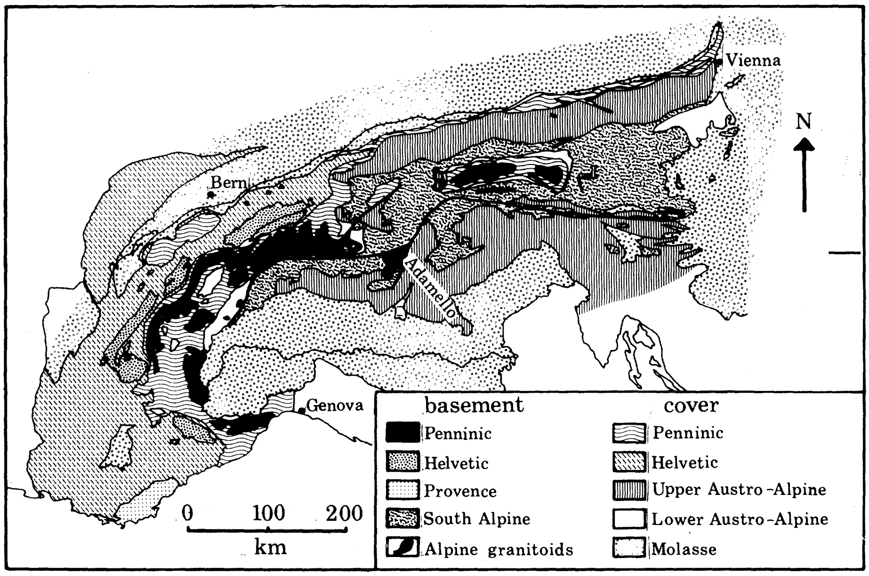

FIGURE 6. Tectonic map of the Alps. The regional setting of the Adamello complex, shown as the largest Alpine granitoid at the junction of two major faults (heavy lines), the east–west trending Insubric–Tonale line and the northeast–southwest trending Giudicarian line. The terms *basement* and *cover* in the key should be read as indicating structurally lower or structurally higher units in a purely local sense.

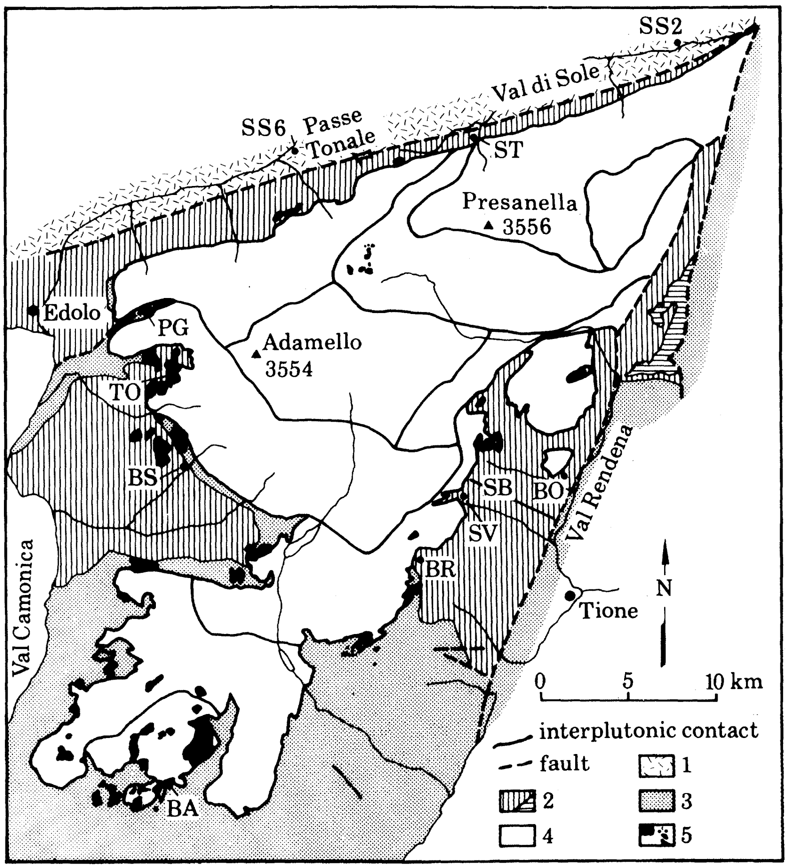

FIGURE 7. A geological sketch map of the Adamello area (after Bianchi *et al.* 1970): 1, metamorphic rocks of the upper east Alpine nappe; 2, palaeozoic schists and phyllites forming the 'basement' of the southern Alps; 3, permo-mesozoic sedimentary cover of the southern Alps, unconformable on 2, and Hercynian intrusions (horizontal ornament); 4, tonalites and granodiorites of the Adamello complex: solid lines show internal boundaries between separate intrusions; 5, small mafic intrusions, largely hornblende gabbros. Pairs of letters indicate localities at which McRae (1983) determined contact aureole pressures and temperatures.

local crystallization can lead to density differences by depletion of the liquid in certain components. Locally such differences may be greater than thermal density differences which they may reinforce or counteract (see, for example, Sparks *et al.* 1984).

A natural example: the Adamello Intrusive Complex

We now apply some of the considerations discussed above to an excellently exposed igneous complex in the southern Alps.

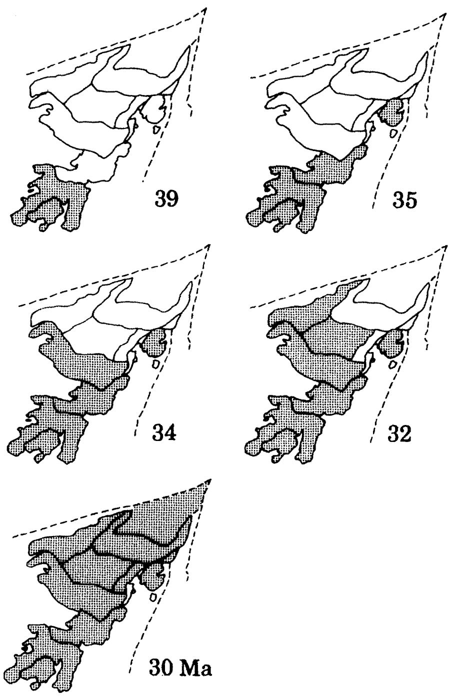

FIGURE 8. Progressive intrusion of the Adamello complex. The tone shows the area of the intrusive complex emplaced at different times; note the southwest–northeast progression. Data from the geochronology laboratory of the University of Pisa (McRae 1983).

Regional setting

The Adamello complex is a calc-alkaline, multiple intrusion that penetrates the south Alpine metamorphic basement and its unmetamorphozed Mesozoic cover (figure 6). It is of late Eocene–Oligocene age (see below) and is elongated in a northeast–southwest direction with a maximum length of about 80 km and a maximum width of about 30 km (figure 7). The complex is bounded to the north by a major fault, the Tonale line: the easterly continuation of the Insubric line of the central Alps. This line today separates a region to the north within which there is clear evidence of Alpine (i.e. mid-late Tertiary) regional metamorphism and deformation, from a region to the south that is unaffected by post-Palaeozoic regional metamorphism and where Mesozoic rocks are much less intensely deformed. The line also separates a region of generally northward vergence to the north, from a region of southward-facing structures to the south.

To the east the complex is bounded by the Giudicarian line, a major fault that has clear topographic and geological expression but is of controversial origin (McRae 1983).

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To both north and east thin slivers of country rock intervene between the fault and the igneous rocks of the complex. McRae (1983) has shown that metamorphic mineral assemblage in the country rocks, exposed at present levels of erosion, crystallized at depths ranging from about five to ten kilometres, the depth generally increasing from southwest to northeast.

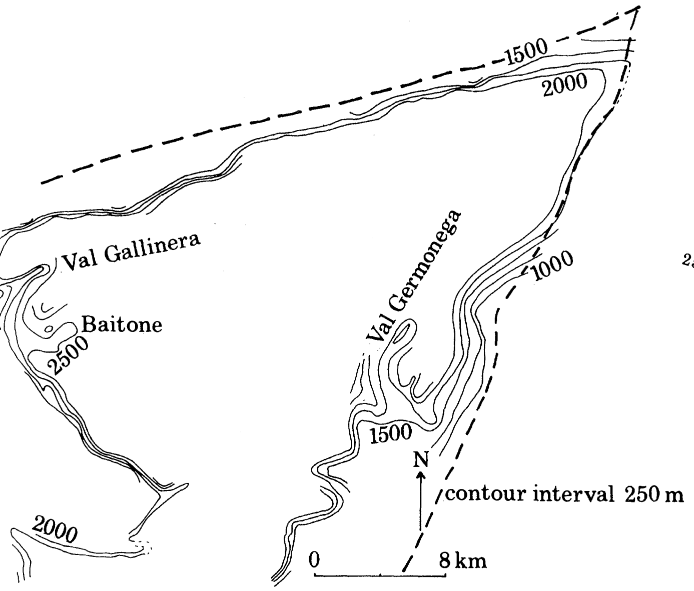

FIGURE 9

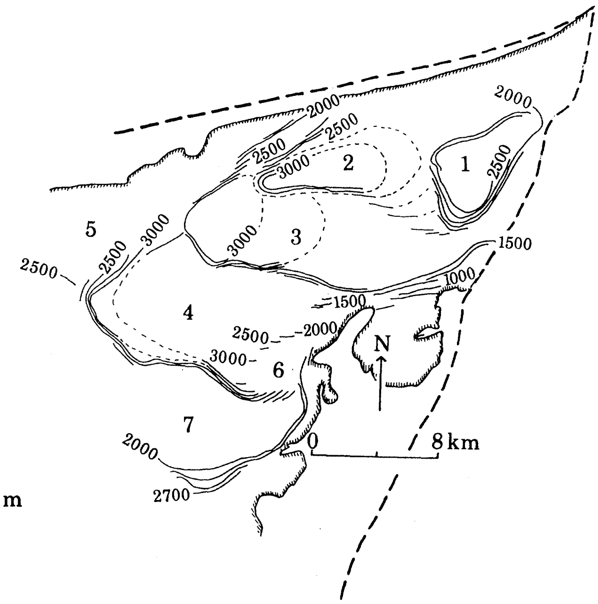

FIGURE 10

FIGURE 9. The external form of the Adamello complex. Structural contours have been constructed on the contact with the country rock (contour values in metres). Note that the contact dips outwards on all sides and that it shallows inwards.

Heavy dashed lines are major faults: the Insubric–Tonale line on the north and the Giudicarian line on the east. Compare with figure 7.

FIGURE 10. Shapes of the component intrusions of the Adamello complex; structural contours shown as in figure 9: 1, Upper Nambrone Valley quartz diorite; 2, NE Presanella tonalite; 3, Central Presanella tonalite; 4, Adamello Central Peaks quartz diorite; 5, Avio Valley quartz diorite; 6, Lower Genova Valley quartz diorite; 7, Corno Alto leucoquartz diorite.

The different intrusive bodies that make up the complex are distinguished on textural and compositional grounds. They are generally of tonalitic to granodioritic composition and commonly contain quartz, plagioclase, biotite and amphibole. Except locally, other components tend to be of minor importance. The relative proportions of all the main components can vary markedly within and between individual intrusions.

Several features of the complex deserve particular attention. Figure 8 shows how the complex grew with time. The figure is based on extensive Rb–Sr and K–Ar studies carried out at the geochronology laboratory of the University of Pisa (Ferrara 1962; Borsi *et al.* 1966, 1977; McRae 1983) and on relative age relations between different bodies determined by field observation. The oldest ages measured are close to 42 Ma and are in the extreme southwest. Although the pattern will no doubt be modified with further field and laboratory studies, a general progression from south west to north east is clear over a time-span of about 10 Ma. Even the largest component intrusions have maximum horizontal dimensions (i.e. $2r$) of less than 20 km. Reference to figure 4 shows that, for any reasonable assumption about their vertical dimensions, they would have solidified in less than one million years. Thus the time taken for

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the emplacement of each individual body was small by comparison with the duration of the formation of the complex as a whole. Figure 8 shows that pre-heating could not have been very important.

The exposure in and around the Adamello complex is so good and the relief so extreme, that it is possible to construct structural contours on both the external contacts of the complex with the country rock (figure 9) and the internal contacts between the individual intrusive bodies

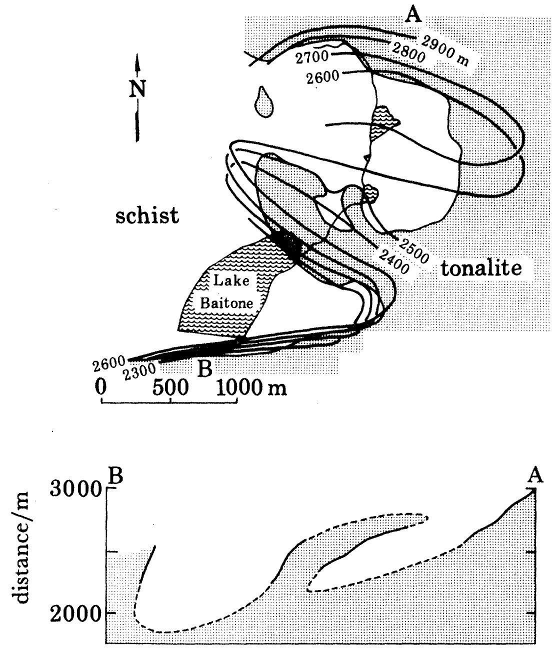

FIGURE 11. Roof stoping at Baitone: structural contours on the contact between tonalite and schist and, below, cross section; kilometre-scale blocks of the roof appear to have been sinking into the top of the magma chamber at the time the magma finally froze.

figure (10). The external contacts dip uniformly outwards and in several places, e.g. Baitone and Val Germonega, it is clear that part of the roof of the complex is preserved as the dip of the contact shallows inwards. In a number of places it is possible to map very large masses of country rock (i.e. dimensions of hundreds to thousands of metres) that are surrounded by igneous rock. In many cases there is no way of telling whether these blocks had become detached from the roof or not. At Baitone, however, detailed mapping and density measurements suggest (figure 11) that a large mass of country rock was in the process of separating from the roof and was about to sink into the magma chamber at the time when the presently exposed levels crystallized.

Although there is some deformation of the country rock surrounding the intrusion associated with its emplacement, it is of a relatively minor nature and amounts to small-scale folding and jointing (McRae 1983). Where the intrusive contacts are exposed they are commonly sharp, with the tonalite occasionally containing angular blocks of wall rock of clearly local origin. In some cases the country rocks near the contact seem to have tilted inwards, towards the intrusion, on a scale of hundreds of metres.

Examination of the contacts that are internal to the body indicates the same process. In general these contacts are less readily identified because of the similarity of intruded and intruding lithologies. Nevertheless they may be satisfactorily mapped (see Callegari & Dal Piaz 1973 for discussion). The younger bodies appear to truncate the older bodies, and the contacts of the younger bodies dip outwards into the older bodies as if the truncation had occurred by stoping (figure 10).

These characteristics suggest that the final stages of emplacement of the intrusions of the Adamello complex were accomplished by the stoping of the roof and walls, and that the stoped blocks may in some cases have been on the kilometre scale. The absence of strong deformation of the surrounding country rock on the one hand, and of conspicuous chemical interaction with the country rock on the other, make it very difficult to attribute any significant role either to zone melting or to penetrative intrusion (compare the Peruvian Batholith; Pitcher 1978).

Xenoliths

In some parts of the Adamello complex xenoliths are abundant. They are of two kinds: those that may be clearly related to the present-day country rock and those that are simply recognized as mafic patches within the igneous host. The former may be of almost any size from centimetres to kilometres, tend to have sharp margins, angular corners, seem to have separated from the country rock very shortly before being frozen into their present position and are restricted to the marginal zone. The latter are mostly less than 30 cm in diameter, although occasional examples may reach one metre. In shape they are very rarely angular, and are commonly spherical to prolate ellipsoidal, with clusters of xenoliths in the same area showing the same orientation and relation to the foliation; occasionally they are drawn out into long streaks and may even be folded. They are much more abundant than the first type and may occur anywhere within the complex.

Xenoliths of the second type display the same mineralogy as their hosts (quartz, plagioclase, amphibole ± biotite) and are distinguished simply by a greater proportion of mafics or by a difference in texture. Although mafic compositions are more prominent, felsic varieties are also present although much less conspicuous. The origin of these xenoliths is not clear and requires further study; they could be completely melted country rock; they could be cognate, or they could be derived from earlier intrusions.

The size limitation of the latter group of xenoliths may be interpreted in different ways; it may simply reflect the size distribution of blocks acquired by the magmas during the earlier stages of their emplacement. In that case the range of sizes was much more restricted than that which was available in later stages of emplacement (i.e. xenoliths of the first type).

Alternatively, the size distribution could result from the dependence of the assimilation process on the fourth power of the xenolith radius discussed earlier. Figure 5 shows that a block with a radius of 15 cm would need to sink about 1 km in a magma with a viscosity of *ca.* 10^4 P† in order to be fully melted. The sinking distance (i.e. the height of the intrusion) could be rather less if the viscosity were greater; blocks of larger radius would tend to sink rapidly to the bottom of the magma chamber. The effect of convection would be to modify this pattern a little. Blocks entrained in a descending flow would simply reach the bottom more rapidly than otherwise. Those caught in ascending flows would sink more slowly, and this would result in the complete melting and incorporation of a few xenoliths larger than otherwise expected.

The simple r^4 dependence discussed above would in nature be modified in another way; as

† 1 P = 10^{-1} Pa s.

the outer parts of a sinking xenolith were warmed and melted the mean density difference between it and the magma would be reduced. This in itself would have a relatively small effect, but if the outer melted zone were continuously left behind as the xenolith sank, much larger xenoliths could be assimilated. Effectively the value of r becomes a function of time, and xenoliths that started their descent rapidly would slow down rapidly as they lost material by 'melting ablation'. The tracks of such xenoliths through the magma should be marked by steep elongate zones enriched in mafic material, perhaps even giving a crude layering, depending on the size and shape of the blocks.

Although the values that may be read from figure 5 may be regarded as no more than order-of-magnitude estimates, they show clearly that there must be a size limit to assimilated blocks; the limit will change during the ascent and cooling of any particular body.

Amounts of assimilation

We now consider the overall limits on assimilation (i.e. melting) of country rock by a magma. The problem involves the interaction of a number of independent variables and in detail is rather complicated. The approach is to consider what requirements must be met if, for example, a magma is to assimilate its own mass of country rock if it undergoes no other gain or loss of heat in the process. If the magma and the country rock have similar melting temperatures and the country rock is initially close to its melting temperature, the magma must have an initial temperature three or four hundred degrees Celsius above its melting temperature, in order to provide the latent heat of fusion for the country rock and not to crystallize itself,

$$\text{i.e.}\quad c_p \Delta T \geqslant \Delta H_f,$$

where c_p is the heat capacity of the magma, ΔT is the amount by which the magma temperature exceeds the melting temperature, and ΔH_f is the latent heat of fusion of the country rock, typical values of c_p and ΔH being 0.25 cal g^{-1} and 80 cal g^{-1}† respectively.

If the initial temperature of the country rock is lower than assumed, ΔT must be greater; if the country rock melts at a lower temperature than the magma or has a lower value of ΔH_f, ΔT may be correspondingly less. In all natural situations, however, the magma body will continuously lose heat by conduction through its roof and walls; such losses would have to be compensated by a somewhat greater value of ΔT. These qualitative considerations suggest that for plausible values of ΔT a magma is unlikely to be able to completely assimilate more than its own mass of country rock.

It was assumed above that no concurrent crystallization of the magma occurred during melting of the country rock. If this condition is relaxed crystallization of refractory phases from the magma may provide heat for greater amounts of assimilation. Such a balance between crystallization and melting is effectively what occurs in the zone melting process discussed earlier. There it was shown that in the most favourable circumstances a body of magma might melt or 'process' an amount up to its own volume of country rock. In so far as the analysis of zone melting takes account of upward, but not lateral, conductive heat losses, this amounts to saying that if the conclusion of the previous paragraph is modified to take account of *both* vertical conductive heat losses *and* crystallization effects, they roughly cancel each other, and the overall upper limit to assimilation remains the same. The lateral heat losses that have been ignored must mean that under very favourable circumstances (e.g. a basaltic magma in granite

† 1 cal = 4.184 J.

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or quartz pelitic country rock) a realistic upper limit to assimilation by a magma body must be significantly less than its own volume.

Discussion

McRae (1983) has reviewed the isotopic and non-isotopic chemical information (see, for example, Bianchi *et al.* 1970; Cortecci *et al.* 1979; Taylor 1980; Dupuy *et al.* 1982) presently available for the Adamello complex. The rather limited information is compatible with the formation of magmas of granodioritic to tonalitic composition by the assimilation of variable amounts of country rock by a high-alumina basaltic magma.

Although their significance is unclear, it is noteworthy that numerous small mafic intrusive bodies occur round the margin of the Adamello complex. These are predominantly hornblende gabbros and are of variable age in so far as there is some suggestion that a phase of localized minor mafic intrusion preceded the emplacement of each major calc-alkaline body within the complex (figure 7).

One highly speculative means of generating calc-alkaline bodies such as that at Adamello would be by the emplacement of a large volume of mafic magma at the base of the crust where it could be gravitationally stable, having a lower density than the mantle and density similar to that of the lower crust (this is plausible but uncertain, because of lack of knowledge of the compressibility of magmas at high pressures).

Unless tectonically injected to higher levels such a magma would presumably remain at the base of the crust possibly 300–400° hotter than its surroundings. Initially it would be too dense for there to be any tendency to ascend either by diapirism or stoping. On the other hand it presumably would begin to melt its roof and to crystallize at its base. As incorporation of country rock progressed, the density of the magma would fall as its gross composition changed. This process might continue until a sufficient density difference existed that either diapiric penetrative intrusion or stoping began and the magma started a relatively rapid ascent through the crust.

In the case of the Adamello complex only the last phase of this ascent is seen, as the present-day erosion level is close to the top of the intrusions. It is interesting to note, but not easy to explain, why the present erosion level corresponds roughly to the tops of all the intrusions, although they seem on the evidence of the contact aureole mineral assemblages to have been intruded under different thicknesses of cover (McRae 1983). It is fairly clear that this last phase of ascent was accomplished by stoping, each successive intrusion stoping both country rock and the marginal parts of adjacent earlier bodies. The larger blocks of both sank rapidly through the magma and affected it relatively little; smaller blocks were completely melted but, at least in some cases, retained their integrity and were not dispersed in the magma.

If this model is correct the middle crust under Adamello should today comprise a jumble of blocks of different sizes and different lithologies partially melted to different degrees. This would probably be mapped as a migmatite complex if exposed at the surface.

It is similarly intriguing to wonder what field evidence should be sought in crust through which an intrusion has ascended by penetrative intrusion or by zone melting. On the one hand dyke-like penetrative intrusion would presumably leave a narrow sheet-like body reflecting some degree of fissure closure after the main volume of the magma had passed; on the other hand diapiric penetrative intrusion of tear-shaped bodies should leave behind it quite distinctive concentric, or at any rate axiosymmetric structural patterns, in which the rocks showed evidence first of radial shortening and then radial stretching, as the body passed through them.

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Conclusions

The only magmas that are able to pass through the crust without substantial modification of their composition by crustal contamination are those that ascend by penetrative intrusion. Magmas ascending by this means must either have densities lower than the mean crustal density or be pumped to high crustal levels by tectonic processes.

Both zone melting and stoping offer the possibility of modifying magma composition by assimilation, but in either case the maximum volume of material assimilable is rather less than the original volume of the magma.

On the assumption that most igneous intrusions undergo relatively little further vertical movement after they have crystallized, most igneous bodies of geological interest are sufficiently small that they must rise through the crust relatively rapidly to their final emplacement position (i.e. less than 10^6 years).

There are physical limits on the amount of crustal ascent that can be accomplished either by zone melting or by stoping: a distance equal to the original height of the magma body in the first case, and several times the height in the latter.

The assimilability of any xenolith depends on the fourth power of its radius, and this will impose a relatively sharp upper limit on the size of assimilated xenoliths observed in intrusive igneous bodies.

The separate intrusives of the Adamello complex were probably formed by the zone melting of the lower crust by high-alumina basaltic magmas. The resultant intermediate composition magmas penetrated to within 10 km of the surface before solidifying and did so at least in part by stoping processes. The role of penetrative intrusion is unknown.

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Discussion

M. J. O'Hara, F.R.S. (*Department of Geology, U.C.W. Aberystwyth, Dyfed, U.K.*). The thermal constraints that limit the extent to which an ascending magma body could assimilate material from its walls and roof in either the zone refining or the stoping models of ascent, are greatly relaxed if the advancing magma-filled chamber has been the site of extensive passage of magma, or input of one (parental) magma and output of another (derivative) magma. Have the authors any evidence which bears on the existence of

(*a*) extensive acid eruptives associated with the uprise of the Adamello plutons, and

(*b*) magma mixing within the granitoids or

(*c*) more basic magmas within the same magma chambers but beneath the currently exposed granitoids?

Alternatively, do the field observations permit them to quantify the extent of assimilation in the later stages of emplacement, and hence to constrain the permissible extent of magma throughput by less direct means?

E. R. Oxburgh. The authors agree with the comment, and this question is considered in some detail in the written paper. In answer to the specific questions: there is no sign in the Tertiary sedimentary record of Northern Italy of acid eruptives contemporary with the emplacement of the Adamello complex.

There is no clear evidence of magma mixing within the granitoids but in some areas mafic schlieren are common. Their significance is unclear, nor is there evidence of basic magmas beneath the presently exposed granitoids. The relatively smooth gravity field around and across the complex suggests that any underlying mafic material is not shallow.

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