

# Effects of volatiles on mixing in calc-alkaline magma systems

Herbert E. Huppert\*, R. Stephen J. Sparks† & J. Stewart Turner‡

\* Department of Applied Mathematics and Theoretical Physics, University of Cambridge, Silver Street, Cambridge CB3 9EW, UK

† Department of Earth Sciences, University of Cambridge, Downing Site, Cambridge CB2 3EQ, UK

‡ Research School of Earth Sciences, Australian National University, Canberra, Australia

*The intimate mixing between different magmas of disparate densities, characteristic of some calc-alkaline magma systems, is explained by a new mechanism involving the emplacement of a layer of wet undersaturated mafic magma at the base of a magma chamber containing more differentiated magma. Turbulent transfer of heat between the mafic magma and the overlying magma leads to crystallization and exsolution of volatiles in the lower layer. For an initial water content of a few per cent the bulk density of the mafic magma can become equal to that of the overlying magma causing overturning and intimate mixing of the magmas. The mechanism is most probable in the low pressure environment of a high level magma chamber.*

THE replenishment of a magma chamber by an influx of new magma from depth has been recognized as important in determining petrological and geochemical characteristics of volcanic and plutonic rocks<sup>1</sup> and in triggering volcanic eruptions<sup>2-5</sup>. Previous investigations of the fluid dynamics of replenishment<sup>3,6-9</sup> have been directed mainly towards understanding systems involving dry basic magmas and have not been concerned with effects resulting from the release of gas. In this article we investigate some of the fluid dynamical phenomena due to the presence and release of water vapour which can occur when a magma chamber is replenished by hydrous magma, and consider specifically magmas of the calc-alkaline association. We note that the term 'calc-alkaline' is a common, but imprecisely defined way of describing magmas which are generally associated with orogenic belts and some island arcs. Such magmas are usually, but not ubiquitously, characterized by low amounts of Fe enrichment and relatively high amounts of volatiles.

There is increasing evidence that intimate mixing of magmas is a common phenomenon. In basaltic magmas mixing can be explained in many circumstances because differentiated basaltic magmas can have the same density as higher temperature, more primitive basaltic magmas<sup>3,6,9</sup>. Also the relatively low viscosity of many basaltic magmas allows rapid mixing between them. However, evidence for magma mixing in volcanic rocks of the calc-alkaline association is also common. The belief that mixing has an important role in the evolution of many basaltic andesites, andesites and dacites is based on observations of disequilibrium phenocryst assemblages, inhomogeneous glass compositions, wide ranges of compositions of glass inclusions in neighbouring phenocrysts, banded pumice fragments and mafic clots or micropillows incorporated in more silicic pumice or lava which show distinctive quench textures<sup>5,10-14</sup>. Possibly the most convincing petrographical evidence for intimate mechanical mixing comes from the occurrence of bimodal phenocryst assemblages. Many basaltic to andesitic lavas, for example, contain both calcium-rich and calcium-poor plagioclase phenocrysts and orthopyroxene and augite with high Fe/Mg ratio together with pyroxene with low Fe/Mg ratio<sup>11-14</sup>. Petrological arguments often indicate that mixing between two or more compositionally distinct magma batches must have occurred only shortly before eruption, leading to the suggestion that mixing may even trigger an eruption<sup>2,4,5,10</sup> or alternatively that the mixing was a consequence of the eruption.

If magma mixing is accepted as a widespread process in many calc-alkaline magma systems, there is a major physical problem to solve. Dry magmas of the calc-alkaline association show a steady decrease in density with differentiation along with the

decline in temperature<sup>15</sup>. In a conventional model of a volcano, the plumbing system or chamber is filled with magma which gradually differentiates during repose periods. If new undersaturated magma replenishes a chamber from depth (perhaps resulting in an eruption) and forms a lower layer it will, in general, be more primitive, hotter and denser. But the decrease in density of the new magma due to crystallization and thermal effects alone is not sufficient to make the densities of the new and resident magmas equal. Thus there does not at first sight appear to be a driving force in such a situation to cause mixing. This argument, however, assumes that vapour is absent (no free volatiles). Calc-alkaline magmas are frequently rich in water, and Anderson<sup>16</sup> has summarized the evidence which suggests that the water content of many mafic to intermediate calc-alkaline magmas is often in the range 1-4% by weight. Eichelberger<sup>5</sup> has shown that the presence of either dissolved or exsolved volatiles in calc-alkaline magmas has large effects on bulk density. We continue along this line to demonstrate that even a small percentage of volatiles can have profound effects on the fluid dynamics of convection in replenished magma chambers.

We consider in this article the influence of water exsolution on the densities of oversaturated hydrous mafic magmas and show that the induced changes in density can be substantial at low to moderate pressures. This idea follows on from the study of Eichelberger<sup>5</sup> who, influenced by observed density differences of mafic inclusions within rhyolitic lavas, proposed that vesiculation of hydrous mafic magma at the interface between two layers can cause density changes large enough to allow mafic magma to become buoyant in more silicic magma. However, only the relatively thin interfacial region would be destabilized by the process suggested by Eichelberger, and not the entire lower layer. We aim to demonstrate that density changes in the whole layer of saturated hydrous basaltic or andesitic magma can be considerable and that, at low to moderate pressures, intimate mixing can take place between magmas of the calc-alkaline association.

## Effects of volatile release on magma density

It follows from basic principles that at equilibrium the bulk density of a hydrous saturated magma in the presence of excess water is given by

$$\sigma^{-1} = RT(N - n)/P + (1 - N + n)/\rho \quad (1)$$

where  $\sigma$  is the bulk density of the magma,  $R$  is the gas constant,  $T$  is the temperature in Kelvin,  $P$  is the pressure,  $N$  is the total weight fraction of water,  $n$  is the saturated weight fraction of water dissolved in the melt phase of the magma, and  $\rho$  is the

weighted density of crystals plus melt. The first term in equation (1) represents the contribution due to the gas and the second term that due to the crystals and melt. Only if  $N > n$  is there any gas; otherwise  $\sigma = \rho$ . The perfect gas law can be used in equation (1) because at the temperatures and pressures under consideration the departures from this law are less than a few per cent.

The weight fraction  $n$  can be estimated from data on the saturation concentration of water in silicate melts<sup>17</sup>. The solubility of water can be calculated from the empirical formula

$$n = s(1-x)P^{1/2} \quad (2)$$

where  $s$  is the solubility constant and  $x$  is the weight fraction of the crystalline phases in the magma.

As an example, we consider the case of hydrous basaltic magma injected into the base of a magma chamber containing more fractionated, cooler and lower density magma (basaltic andesite, andesite or dacite). We make the simplifying assumptions that the crystallization of the basaltic magma involves only anhydrous phases. The bulk density,  $\rho$ , of the residual melt and crystals has been calculated by assuming that the crystals have a constant density of  $2.8 \text{ g cm}^{-3}$  and by evaluating the decrease in density of the residual melt as water is concentrated by crystallization. The density of the melt was calculated using the method of Bottinga and Weill<sup>18</sup>. Variations of  $\rho$  with  $T$  alone due to thermal expansion are neglected because they are negligibly small in comparison with changes in bulk density of the magma due to volatile exsolution. The solubility constant  $s \approx 0.0014 \text{ bar}^{-1/2}$  is taken from solubility data of water in Mount Hood andesite<sup>17</sup>.

Finally, we require a relationship between temperature and crystal content. The fractional crystallization of calc-alkaline magmas is quite complex and varies from volcano to volcano. Thus a unique relationship cannot be accurately formulated. However, inspection of estimates on the eruption temperatures and on fractional crystallization models of calc-alkaline magmas suggests that

$$T = 1,373 - 200x \quad (3)$$

is a plausible expression<sup>14,19,20</sup>. The results of our density calculations are not in fact strongly dependent on temperature and thus an approximate relationship such as equation (3) seems acceptable.

Figure 1 presents the results of bulk density calculations for a hydrous basalt crystallizing at 0.5, 1.5 and 3.0 kbar pressure with an initial temperature of  $1,100^\circ\text{C}$  and zero initial crystal content. The specific calculations presented are representative of the general behaviour of gas-rich magmas and are similar to the results of Eichelberger<sup>5</sup>. From equation (2), the maximum total weight fraction of dissolved water in a silicate melt is represented by  $N = SP^{1/2} \equiv N_0$ . For  $N < N_0$ , the magma is initially undersaturated. It continues to be undersaturated until the weight fraction of crystals  $x = 1 - (N/N_0)$ , when saturation first occurs. During this stage the bulk density increases slightly as the proportion of crystals in the system increases. Further crystallization involves exsolution of small water vapour bubbles and, assuming these remain suspended in the magma, leads to a continuously decreasing bulk density. In general, if the total water content is decreased, keeping all other variables fixed, the density increases. If the pressure is increased, keeping all other variables fixed, the density of the gas bubbles increases and thus so does the bulk density.

The specific calculations demonstrate that the bulk density of hydrous basaltic magma can vary as a consequence of total water content and substantially as a result of volatile exsolution. The approximate densities of dry degassed magmas of the calc-alkaline association are indicated on the right-hand side of Fig. 1. It follows that at 1.5 kbar, for example, oversaturated basaltic magma can reach the same density as dry, or even volatile undersaturated, andesitic, dacitic or rhyolitic magma. This comparison of hydrous basalt densities with 'dry' densities

is merely presented to illustrate the very substantial variations that can occur in the bulk density of a fixed magma composition if excess volatiles are present in the lower layer.

Even at pressures  $> 1.5$  kbar the changes can still be large enough to cause density inversions to occur in some magma chambers, although the total water contents required are at the high end of the range reported by Anderson<sup>16</sup>. If water contents between 1 and 4% are typical of calc-alkaline mafic to intermediate magmas, the density changes necessary for convective mixing to take place can occur in shallow-level magma chambers.

### Geological applications

When mafic magma is injected into the base of a chamber containing less dense, more differentiated magma, any mixing

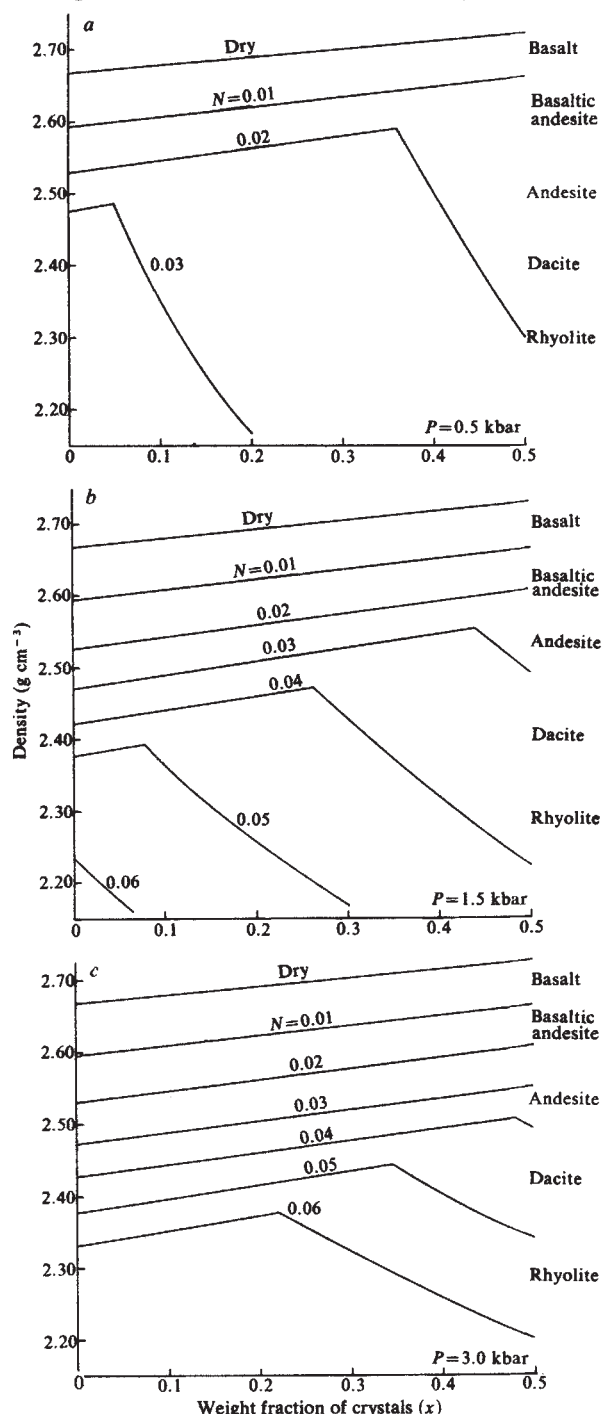
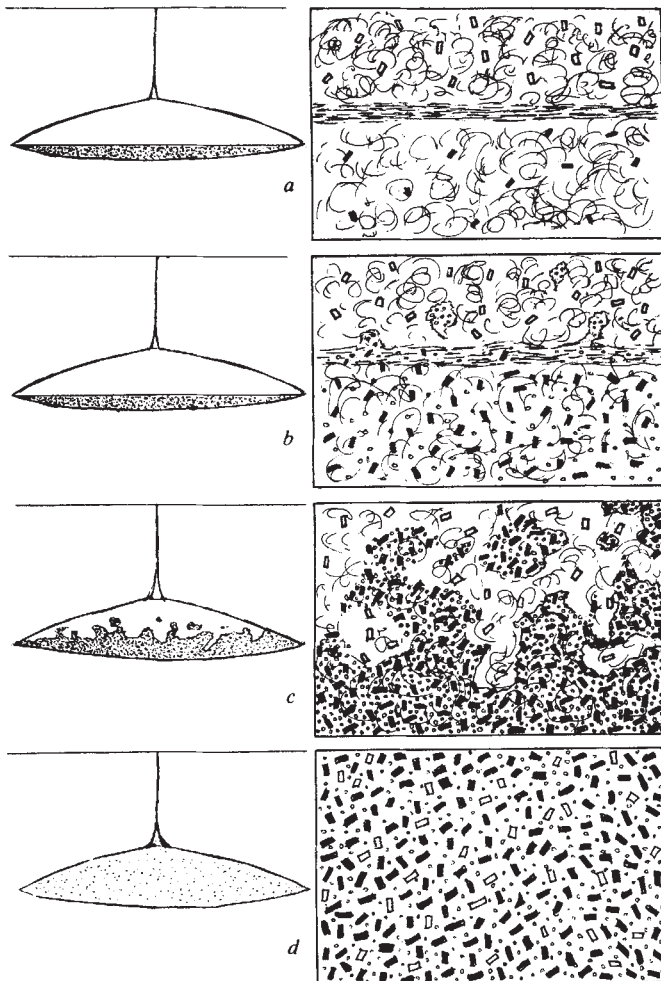


Fig. 1 The bulk density of wet basaltic magma as a function of the weight fraction of crystals for various total weight fractions of water,  $N$ , at: a, 0.5 kbar; b, 1.5 kbar; and c, 3.0 kbar.



**Fig. 2** Schematic model for a magma chamber replenished by an influx of hot, wet basaltic magma. The four diagrams on the right represent close-up views at the height of the interface of the corresponding diagrams on the left. *a*, The heavier magma of the lower layer loses heat across the interface and crystallizes. *b*, The magma of the lower layer becomes saturated and further crystallization produces exsolution of water vapour bubbles. *c*, The bulk density of the magma of the lower layer becomes equal to that of the overlying magma and the lower-layer magma rises into the upper layer. *d*, Intimate mixing between the magmas results.

will initially be strongly inhibited by the density contrast. A two-layered or multi-layered system, with a horizontal interface separating the recently injected hot, denser, mafic magma from the cooler, less dense, silicic magma above, therefore seems a likely precursor to any well-mixed state. Subsequent crystallization is one process for producing a decrease in density of the lower layer. In the case of picritic basalt magmas, crystallization of dense components, notably olivine, leads to a decrease in the density of the residual liquid which can be sufficient to cause overturning and mixing with overlying differentiated basalt<sup>6,7</sup>. In the case of calc-alkaline magmas, changes in the density of the liquid composition due to crystallization alone may not be sufficient to result in overturn. We suggest that it is the release of water vapour once the lower layer has become saturated, which is essential.

The overall process we envisage has a qualitative resemblance to one proposed<sup>6,7</sup> for basaltic magma chambers. We suppose that hydrous mafic magma enters a chamber containing more differentiated magma, as depicted in Fig. 2*a*. This influx may be a consequence of tectonic activity and could possibly trigger an eruption<sup>2,4</sup>. We assume that the volume of new magma forming the lower layer is much less than that of the upper layer and so the evolution of the lower layer proceeds with

little change in the upper layer. This is depicted quantitatively for the case of a dry basaltic magma chamber in Fig. 7 of ref. 7. As in our previous analyses of dry basaltic systems, the hydrous mafic magma transfers heat through a sharp interface and crystallizes (Fig. 2*a*). As crystallization proceeds, the weight fraction of dissolved water increases in the residual liquid until the magma is saturated. It can be shown<sup>7,21,22</sup> that the Rayleigh number of layers, even as thin as a few metres, of viscous andesitic magma at a temperature excess of 100 °C often exceeds 10<sup>6</sup>. Thus the convective motions will be turbulent, as depicted in Fig. 2.

After the magma saturates, small bubbles form and, along with the crystals, are held in suspension by the vigorous turbulent convective velocities in the lower layer (Fig. 2*b*). The bubbles remain in suspension because the velocity of rise of small gas bubbles relative to quiescent fluid is insignificant in comparison with the r.m.s. turbulent vertical velocities in the lower layer. Similar reasoning applies to the crystals; quantitative calculations for the velocities in this case are given in ref. 7. The exsolution of water vapour causes the bulk density of the lower layer to decrease continually in accord with the results presented in Fig. 1. Eventually, the bulk density may become equal to that in the overlying magma (which has changed by only an insignificant amount), whereupon overturning and hybridization of the magmas occur (Fig. 2*c, d*). Further reactions involving the gas phase may be possible during the mixing. We emphasize that we chose the simple and arbitrary chamber geometry depicted in Fig. 2 for schematic purposes only. The processes we have described are fundamental and would also occur in more complex and geometrically irregular plumbing systems. Note that the magma in the upper layer need not be compositionally homogeneous; the same processes would occur if the upper layer were compositionally zoned, though the mixing would be confined to the lower parts of the system<sup>9</sup>.

Eichelberger<sup>5</sup> proposes a different mechanism for mixing in which the interface between the mafic and silicic layers vesiculates and becomes unstable. This mechanism seems possible and is incorporated in Fig. 2*b*. However, the interface thickness can be calculated<sup>7</sup> to be quite small, of the order of a few centimetres for temperature differences between layers of ~100 °C, and will thicken to only a few metres as the layers exchange heat. Thus, while instability of a vesiculating interface may account for the occurrence of some mafic inclusions in silicic volcanic rocks, it cannot account for the mixing of large volumes of magma.

There are several further predictions which follow from the general model when applied directly to calc-alkaline systems. For a given pressure the mafic magma must have a water content within a restricted range for the mechanism to be viable. For example, as illustrated in Fig. 1*b*, at 1.5 kbar the total water content for the mafic magma to be initially undersaturated must be <5.4% and, assuming that a fluid model is applicable only up to a crystal content of 50%, the total water content for saturation to occur must be >2.8%. Overturning into an upper rhyolitic layer of density 2.3 g cm<sup>-3</sup> will not occur unless the total water content exceeds 3.8%. If the upper layer of andesitic magma has degassed, however, the total water content of the mafic magma needs only to exceed 3.0%. Thus there will be situations in the calc-alkaline association where overturning cannot occur and a final compositional stratification will result. Due to the variation of these limits on water content with chamber pressure, with the composition of the more differentiated magma and with the water content of the input mafic magma, both compositional zonation and intimate mixing could occur in the same magmatic system, but at different depths. The data in Fig. 1 suggest that the mechanism we propose is most easily achieved in high-level conduits and near surface chambers beneath volcanoes.

There are considerable viscosity variations amongst calc-alkaline magmas, which will result in large variations in the time scale of the evolution of the layers. A quantitative analysis can be made for the heat transfer in a two-layer system<sup>7</sup> and,

from equation (11a) of ref. 7, the temperature and crystallization history of the lower layer can be predicted. For basaltic systems the calculations suggest that the time taken to overturn can range from weeks to years depending on the initial thickness of the lower layer. In calc-alkaline magmas, silica enrichment can result in increases in viscosity from two to more than four orders of magnitude over the viscosity of tholeiitic basalts. Application of the two-layer system model<sup>6,7</sup> indicates that the time for overturn varies directly as the cube root of the viscosity, and hence overturn times may be increased by factors of order 10 or more for calc-alkaline magmas. Incubation times are then typically years or decades. For more complex layered systems, a similar overturn mechanism can occur<sup>9</sup>, but the thermal history of the system cannot yet be quantified. However, overturn times are likely to be considerably longer.

In connection with volcanic activity, we note that in a rigid chamber of fixed volume, exsolution of gas in the lower layer must result in increasing chamber pressure. Blake<sup>4</sup> has shown that an increase of chamber pressure sufficient to exceed the chamber wall fracture strength can occur by influx of new magma of volume typically of the order of  $10^{-3}$  of the chamber volume. Exsolution of a gas phase in the lower layer will result in increasing chamber pressure, and thus the generation of gas due to crystallization could trigger an eruption. For example, if the lower layer constitutes 5% of the total chamber volume and a density decrease of  $0.1 \text{ g cm}^{-3}$  occurs due to vesiculation in this layer, the volume of gas will be  $2 \times 10^{-3}$  of the chamber volume. Thus many situations can be envisaged where eruption could be triggered by vesiculation in the lower layer.

We also note that the mixing during overturn may produce reactions between the magmas and volatiles which could be significant in driving or triggering volcanic eruptions. Vesiculated mafic magma from the lower layer will be transported during the mixing process to environments which, on average, will be lower in temperature and pressure. Further crystallization of the mafic magma during mixing will yield additional gas release. The bubbles will have to expand, or alternatively increase their internal pressure, which will generate extra chamber pressure to drive magma to the surface.

Most of the convincing petrographical evidence for hybrid

magma in the calc-alkaline association comes from stratovolcanoes erupting small-to-medium volumes of mafic to intermediate volcanic products<sup>10,12,14</sup>. In such systems stagnant magma occupies the high-level conduits and reservoir systems beneath the summit during quiescent periods. Such magma will have undergone partial or complete degassing during previous periods of activity and periods of quiescence. We envisage that this magma will become either just saturated in volatiles, or in some cases undersaturated if magma degassed at the surface sinks back into the shallow-level reservoir. Many eruptions from such volcanoes are envisaged to occur during ascent of fresh magma from depth. We suggest that as new magma arrives in the shallow-level system, oversaturation can occur and that mixing will ensue between the old degassed magma and the new volatile-bearing magma.

There are other mechanisms which can produce magma-mixing phenomena. For example, as shown in Fig. 1, a saturated mafic magma containing excess volatiles can be lower in density than the more differentiated resident magma. In this case the new magma is immediately buoyant and ascends as either a laminar or an entraining turbulent plume<sup>3</sup>. Mixing can also occur by sidewall crystallization of a mafic magma forming a turbulent layer of light silicic melt<sup>23,24</sup>. Mixing between the boundary-layer fluid and chamber fluid can form hybrid magma, although there is no evidence that this mechanism can produce large volumes of hybrid magma. There are no doubt other situations which also lead to local mixing within the complex, near-surface plumbing systems of volcanoes. We have tried to document a plausible mechanism which is fluid-dynamically possible, and also consistent with conventional ideas on the plumbing systems of volcanoes. The mechanism is capable of explaining the common occurrence of hybrid calc-alkaline volcanic rocks. In some circumstances basaltic magma can become lighter than even dry rhyolitic magma and thus mixing of mafic magmas derived from the mantle and silicic magmas derived from the crust could occur.

We thank A. T. Anderson, R. J. Arculus, J. C. Eichelberger and B. D. Marsh for helpful comments on a previous draft of the manuscript. The paper is contribution 228 from the Department of Earth Sciences.

Received 9 February; accepted 30 April 1982.

- O'Hara, M. J. & Matthews, R. E. *J. geol. Soc.* **138**, 237-277 (1981).
- Sparks, R. S. J. *et al. Nature* **267**, 315-318 (1977).
- Sparks, R. S. J. *et al. Earth planet. Sci. Lett.* **46**, 419-430 (1980).
- Blake, S. *Nature* **289**, 783-785 (1981).
- Eichelberger, J. C. *Nature* **288**, 446-450 (1980).
- Huppert, H. E. & Sparks, R. S. J. *Nature* **286**, 46-48 (1980).
- Huppert, H. E. & Sparks, R. S. J. *Contr. Miner. Petrol.* **75**, 279-289 (1980).
- Huppert, H. E. & Turner, J. S. *Earth planet. Sci. Lett.* **54**, 144-152 (1981).
- Huppert, H. E. *et al. Earth planet. Sci. Lett.* **57**, 345-357 (1982).
- Anderson, A. T. *J. Volcan. geotherm. Res.* **1**, 1-33 (1976).
- Eichelberger, J. C. *Nature* **275**, 21-27 (1978).

- Luhr, J. F. & Carmichael, I. S. E. *Contr. Miner. Petrol.* **71**, 343-372 (1980).
- Sakuyama, M. *J. Volcan. geotherm. Res.* **5**, 179-208 (1979).
- Sakuyama, M. *J. Petrol.* **22**, 553-583 (1981).
- Bottinga, Y. & Weill, D. F. *Am. J. Sci.* **269**, 169-182 (1970).
- Anderson, A. T. *J. Geol.* **87**, 509-531 (1979).
- Burnham, C. W. *Geochim. cosmochim. Acta* **39**, 1077-1084 (1975).
- Bottinga, Y. & Weill, D. F. *Am. J. Sci.* **269**, 169-182 (1970).
- Rose, W. I. *et al. J. Geol.* **85**, 63-87 (1977).
- Arculus, R. J. & Wills, K. J. A. *J. Petrol.* **21**, 743-799 (1980).
- Bartlett, R. W. *Am. J. Sci.* **267**, 1067-1082 (1969).
- Rice, A. *J. geophys. Res.* **86**, 405-417 (1981).
- McBirney, A. R. *J. Volcan. geotherm. Res.* **7**, 357-371 (1980).
- Turner, J. S. & Gustafson, L. B. *J. Volcan. geotherm. Res.* **11**, 93-125 (1981).

## Population dynamics of human helminth infections: control by chemotherapy

Roy M. Anderson\* & Robert M. May†

\* Department of Pure and Applied Biology, Imperial College, London University, London SW7 2BB, UK

† Department of Biology, Princeton University, Princeton, New Jersey 08544, USA

*An analysis is presented of the population dynamics of the major helminth parasites of man with the aim of understanding observed patterns in the age-specific prevalence and intensity of infection. Mathematical models are used to investigate the possibility of controlling helminth diseases by mass chemotherapy, and to explore the advantages of selective treatment of the most heavily infected individuals in a community.*

OVER 30 years ago Stoll<sup>1</sup> noted that the major helminth infections of man were among the most prevalent of all human infectious diseases. Today the global picture is similar, and we still live in a very 'wormy world': there are roughly 1,000 mil-

lion cases each of roundworm (*Ascaris lumbricoides*), pinworm (*Enterobius vermicularis*) and whipworm (*Trichuris trichuria*); over 600 million cases of hookworm (*Necator americanus* and *Ancylostoma duodenale*); 300 million cases of filarial infections.