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A Continuum Model of Crustal Generation in Iceland; Kinematic Aspects

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Abstract. A steady-state plate-tectonic kinematic model of crustal accretion in Iceland is presented. It describes quantitatively the overall time-averaged movements of solid crustal elements during the accretion process, and correlates accretion parameters in the axial zone (width of lava deposition zone, total lava production rate, width of horizontal strain zone, spreading velocity, normal faulting) with structural properties in the Tertiary lava pile (lava dips, lava deposition rate, dyke fraction). The model is used, firstly, to predict the accretion parameters of the Tertiary volcanic zone on the basis of observed structural properties in the Tertiary lava pile; secondly, to predict possible structures of the lower crust in terms of a lava/intrusion ratio; thirdly, the model may be used to calculate the crustal temperature field caused by intrusions, but this application is outside the scope of the present paper. The model is essentially a further development of a previous one presented earlier by the author. The analysis, in terms of the model, of various published structural observations indicates that the width of lava deposition and the spreading rate in the Tertiary volcanic zone were consistent with the corresponding properties in the present-day volcanic zone. This may suggest a certain uniformity in the volcanic processes during the last 10–15 Ma. The visible Tertiary lava pile was, according to the model, deposited outside the innermost 50-km-wide central part of the volcanic zone, which may explain the difference in appearance between the two main volcanic regions of Iceland, i.e., the active volcanic zone and the Tertiary flood basalts. Furthermore, an analysis of possible structures of the lower crust, consistent with various surface observations, indicates a gradual rather than a sharp transition from an upper lava-dominated crust to a lower intrusion-dominated crust.

Key words: Iceland – Mid-ocean ridges – Crustal accretion model – Lava dip – Dykes – Normal faults.

Introduction

In recent years a considerable amount of radiometric age data from the Icelandic basalt pile has been published (e.g. McDougall et al., 1977; Watkins and Walker, 1977). They show a remarkably constant rate of deposition of lavas which may be taken to indicate a certain uniformity in the intensity of the volcanic processes during the time interval since Iceland began to be formed. Other kinds of data on the regional structure of the Tertiary lava pile, such as dips of lavas and dyke volume fraction, also testify to the processes that originally formed the lava pile. The lavas generally dip gently towards the zone of rifting and volcanism, with the dip increasing with depth in the lava pile. The dyke fraction generally increases downwards in the crust, but laterally there are large variations in the dyke fraction. The first systematic study of these relationships was made by Walker (1959, 1960) in eastern Iceland. He interpreted his observations in terms of a sagging process associated with lava deposition in an active volcanic zone. This process was further elaborated by Bodvarsson and Walker (1964).

The present day volcanic zone of Iceland differs in appearance from the Tertiary flood basalt areas. In part this difference is due to the wet environment of the Pleistocene volcanism, but apart from that, the sheer number of visible volcanic vents and faults and fissures is much greater in the axial zone than can be observed in the Tertiary flood basalt pile. The inference is not unnatural that the present day volcanic zone represents a separate episode of volcanism, distinct from that which formed the Tertiary lava pile.

To clarify the relationships between the present day volcanic and tectonic processes in the axial zone and the structure of the Tertiary lava pile, a model is needed which describes the volcanic and tectonic processes taking place. In the present paper such a model will be discussed. It is based on plate tectonics concepts, and is in agreement with the processes envisaged by Bodvarsson and Walker (1964). It is essentially a further development of a model previously presented by the author (Pálmason, 1973). The main modification is that the processes of lava deposition and strain by dyke injection are assumed to have a normal distribution across the volcanic zone instead of a truncated one.

As is well known the active volcanic zone forms a rather complicated pattern through Iceland, with parallel branches and a possible shifting of the zone between two or more locations. Under these circumstances a simple steady-state model can at best only be a rough approximation. When only regional properties are considered, such a model may still be of help in clarifying relationships between various observable properties.

In this paper only the kinematic aspects of the model will be considered, while the thermal aspects will be considered in another paper. It should be noted that the model is intended to describe only the solid or semi-solid crust, but not the underlying two phase flow at temperatures in the melting range. The lower boundary of the model will thus be outlined by isotherms, deduced by calculations of the thermal state of the crust.

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The Model

The assumptions involved in the model were discussed in a previous paper (Pálmason, 1973) and will not be repeated here in detail. The model is essentially a kinematic one, describing the overall time-averaged movement of solid crustal elements during the accretion process. A flow field is constructed in such a way that it fits certain boundary conditions. It is characterized by several parameters that may be varied to fit observations of various structural properties of the lava pile in Iceland. The divergence of the flow velocity field is equal to the intensity of the emplacement of magma in the crust. Specifying the flow field thus makes possible calculation of the thermal state of the crust.

In the model the discontinuous processes of lava extrusion, dyke intrusion and tectonic movements are treated as continuous processes representing the average behaviour over a long time. A disadvantage of this point of view is that the details of the volcanic and tectonic processes, such as individual central volcanic complexes, dykes, and faults, are lost in the model. An advantage is, on the other hand, that analytical descriptions of the regional movement of crustal material are easily made, as well as calculations of the thermal state of the crust. Another approach to the modeling where the volcanic and tectonic processes were treated statistically, was made by Daigniéres et al. (1975).

The Flow Field

The model is a two-dimensional one. The boundary conditions for the flow field are the following:

(a) At the surface of the crust lava is deposited at a certain average rate which varies with the distance from the axis. For a steady-state process the downward movement of the solid crustal elements must be equal to the deposition rate.

(b) At the axis the horizontal component of the flow must be zero for reasons of symmetry.

(c) The distant lithosphere moves horizontally as a 'rigid' body with a constant velocity. The movement is assumed to be perpendicular to the axis.

A flow field satisfying these boundary conditions can be set up in many ways, but perhaps the simplest way is to assume that both velocity components are independent of depth, but vary with the horizontal coordinate only. If furthermore the horizontal strain rate associated with the increase in horizontal velocity from zero at the axis to a final value V_d in the distant lithosphere, is assumed to have a normal distribution with standard deviation σ_1, and the lava deposition rate is also assumed to have a normal distribution with a standard deviation σ_2, the two velocity components may be written, assuming the density to be constant:

$$v_x=\int_0^x \frac{dv_x}{dx}\cdot dx=V_d\frac{\sqrt{2}}{\sqrt{\pi}\cdot\sigma_1}\cdot\int_0^x \exp\left(-\frac{x^2}{2\sigma_1^2}\right)dx$$

$$=V_d\cdot\operatorname{erf}\left(\frac{x}{\sqrt{2}\cdot\sigma_1}\right) \tag{1a}$$

$$v_z=\frac{q}{\sqrt{2\pi}\cdot\sigma_2}\cdot\exp\left(-\frac{x^2}{2\sigma_2^2}\right) \tag{1b}$$

where q is the rate of lava deposition per unit length of the axial zone, but integrated across the width of the zone. The four parameters V_d, q, σ_1, and σ_2 characterize the simple model. Later on, when discussing the effects of normal faulting, two additional parameters will be introduced. In that case the velocity components do not depend only on the horizontal coordinate, but on the depth as well.

The flow field described by these velocities does not conserve material:

$$\operatorname{div}\bar{v}=\frac{\sqrt{2}\cdot V_d}{\sqrt{\pi}\cdot\sigma_1}\cdot\exp\left(-\frac{x^2}{2\sigma_1^2}\right). \tag{2}$$

This is the horizontal strain rate associated with the spreading process and gives the volume rate of emplacement of magma intrusions per unit volume in the crust which in this case depends only on the horizontal coordinate. It is assumed that the magma is emplaced instantaneously from the underlying source zone of partial melting, which is equivalent to saying that the time taken for the movement of the magma to its destination in the crust is short on the time scale of movements of the crust. Some of the magma reaches the surface where it solidifies as lava.

In the flow field above the axial topography is assumed to be a plane surface. This is not a critical assumption, and an irregular but steady-state topography, e.g., a rift valley, would do just as well, but would make the calculations more cumbersome.

For a given flow field it is possible to compute various properties of interest in the solid crust by methods that were described in the previous paper (Pálmason, 1973). Trajectories and isochrons of lava elements, regional dip of lavas, and the lava fraction in the crust can all be computed. In this paper the lava fraction L will be used as much as the dyke fraction D which was used in the previous paper ($L=1-D$). The thermal state of the crust can also be computed, but this is outside the scope of the present paper.

The isochrons are the loci of lavas of constant age in the crust. As long as the lavas deform by bending in the axial zone during the accretion process, the dip of the isochrons is the same as the regional dip of the lavas. This will probably be the case for the uppermost part of the frozen crust. At greater depth the lavas will be more broken by faults and dyke intrusions. In that case the dip of the isochrons will not necessarily be the same as the dips of the lavas.

It is convenient when discussing the flow field to use dimensionless variables for the horizontal and vertical coordinates and the time (age of lavas). They will be defined as follows:

$$\xi=\frac{x}{\sqrt{2}\sigma_2};\quad \eta=\frac{2V_d}{q}\cdot z;\quad \tau=\frac{V_d}{\sqrt{2}\sigma_2}\cdot t;$$

The trajectories of lava elements from their origin of deposition at the surface in the axial zone may in the present case be written as follows (Pálmason, 1973):

$$\eta=\frac{2}{\sqrt{\pi}}\cdot\int_{\xi(0)}^{\xi}\frac{\exp(-s^2)\,ds}{\operatorname{erf}\left(\frac{\sigma_2}{\sigma_1}\cdot s\right)} \tag{3}$$

where $\xi(0)$ is the horizontal coordinate of the point of origin at the surface, and s is a variable of integration. The ratio σ_1/σ_2 crustal strain/lava deposition standard deviations can be expected to lie in the interval 0–1. The two limiting values give the following trajectories:

$$\begin{array}{lll} \text{I.} & \sigma_1/\sigma_2=0 & \eta=\operatorname{erf}(\xi)-\operatorname{erf}[\xi(0)] \\ \text{II.} & \sigma_1/\sigma_2=1 & \eta=\ln\left[\dfrac{\operatorname{erf}(\xi)}{\operatorname{erf}(\xi(0))}\right] \end{array} \tag{4}$$

The difference between the trajectories in the two limiting cases is small at shallow depths in the crust, but increases with increasing depth.

The time parameter, or age of the lavas from the time of their deposition at the surface, may be calculated from the following formulas (Pálmason, 1973):

$$\tau=\int_{\xi(0)}^{\xi}\frac{ds}{\operatorname{erf}\left(\frac{\sigma_2}{\sigma_1}\cdot s\right)} \tag{5}$$

where the integral is along a trajectory originating at the surface position $\xi(0)$.

The lava fraction L may be calculated from the equation of continuity for lavas within the crust.

$$\frac{\partial L}{\partial t}+\operatorname{div}(L\cdot\bar{v})=0. \tag{6}$$

This may also be written:

$$\frac{dL}{L}=-\frac{\operatorname{div}\bar{v}}{v}\cdot ds=-\frac{\operatorname{div}\bar{v}}{v_x}\cdot dx$$

or

$$\ln L=-\int_{x(0)}^{x}\frac{\operatorname{div}\bar{v}}{v_x}dx \tag{7}$$

where the integral is along a trajectory originating at the surface position $x(0)$.

In the special case where $\bar{v}$ depends only on the horizontal coordinate, the expression for the lava fraction reduces to:

$$L=\frac{v_x(x(0))}{v_x(x)}. \tag{8}$$

In general the isochrons and the lava fraction may be calculated numerically from the above equations. In the limiting case $\sigma_1/\sigma_2=0$ (zero width of crustal strain zone) the trajectory $\eta=\operatorname{erf}(\xi)$ forms the sharp boundary between 100% lavas (above) and 100% intrusions (below). The isochrons in the lava pile can in this case be expressed as follows:

$$\eta=\operatorname{erf}(\xi)-\operatorname{erf}(\xi-\tau). \tag{9}$$

Although this expression applies strictly only for $\sigma_1/\sigma_2=0$, it can still be considered a good approximation to the isochrons in the uppermost part of the crust, for other values of σ_1/σ_2. This is because the structure of the uppermost crust, where the dyke fraction is small, is largely independent of σ_1/σ_2. Thus the above relationship can be expected to be a good approximation to the isochrons in the visible Tertiary lava pile of Iceland. The lava trajectories and isochrons for this case are shown in Fig. 1.

The case $\sigma_1/\sigma_2=0$ was discussed by Cann (1974) using graphical methods. As shown by Kidd (1977) this is the only case which with the present assumptions can approximate the relatively sharp transition between pillow lavas and sheeted dyke complexes, which is observed in ophiolite complexes, e.g., in the Troodos Massif in Cyprus. Later in this paper other possibilities will be discussed which can also lead to a relatively sharp transition between lavas and intrusions.

For the other limiting case $\sigma_1/\sigma_2=1$ (crustal strain and lava deposition standard deviations equal) the trajectories and isochrons are shown in Fig. 2. The lava fraction L in the crust is here dependent only on the η coordinate, and is given by

$$L=e^{-\eta} \quad (\text{or } D=1-e^{-\eta}). \tag{10}$$

The Standard Deviations σ_1 of Crustal Strain Rate and σ_2 of Lava Deposition Rate

It is of some importance for the usefulness of the model that the standard deviations σ_1 and σ_2, and in particular their ratio, can be estimated from field data.

If the distribution of lava deposition rate from a single linear vent is normal with a standard deviation σ_3, and the distribu-

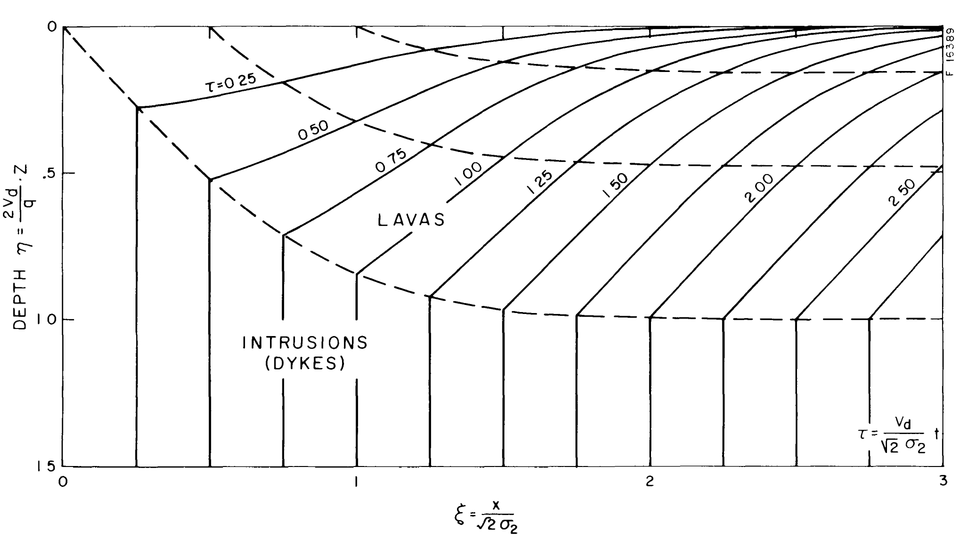

Fig. 1. Dimensionless lava isochrons *(solid lines)* and trajectories *(dashed)* in the model crust for $\sigma_1/\sigma_2=0$ (crustal strain/lava deposition standard deviations)

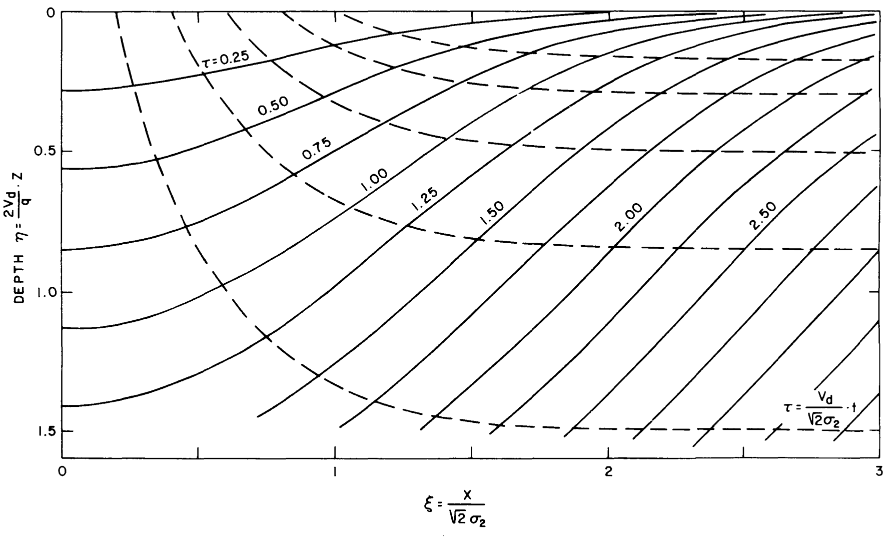

Fig. 2. Dimensionless lava isochrons *(solid lines)* and trajectories *(dashed)* in the model crust for $\sigma_1/\sigma_2 = 1$ (crustal strain/lava deposition standard deviations)

tion of the linear vents around the axis is also normal with a standard deviation σ_1, it may be shown that

$$\sigma_2^2 = \sigma_1^2 + \sigma_3^2 \tag{11}$$

provided the two distributions are independent of each other.

For the mid-ocean ridges, estimates based on the magnetic anomaly pattern indicate strain zone values of a few km for σ_1 (Matthews and Bath, 1967; Harrison, 1968), while no estimates seem to exist for σ_2 of deposition. From the above relation it seems natural to assume that σ_2 is greater than σ_1. The ridge crest topography may, however, play a decisive role in controlling the ratio σ_1/σ_2.

For the Iceland segment of the Mid-Atlantic Ridge the volcanic activity occurs over a considerably wider zone than usually postulated for the mid-ocean ridges. On the basis of the areal distribution of lavas erupted in Iceland during the last 10,000 years, a rough estimate may be obtained for the present-day value of σ_2. An effective width W_e, that is comparable to the width of the zone of Postglacial lavas, will be defined in such a way that at its edge the deposition rate is 1 lava in 10,000 years. The width W_e is then related to σ_2 by Eq. (1 b).

$$v_z = \frac{q}{\sqrt{2\pi} \cdot \sigma_2} \cdot \exp\left(-\frac{W_e^2}{8\sigma_2^2}\right)$$

For two values of v_z, corresponding to lava thicknesses of 2 and 5 m, the following relationships are obtained between W_e and σ_2 (with $q = 4/3 \times 10^{-4}\,\text{km}^2/\text{a}$.):

Table 1

	$v_z = 2 \times 10^{-7}$ km/a	5×10^{-7} km/a
σ_2 km	W_e km	W_e km
5	28.2	24.7
10	51.2	43.5
15	71.9	59.4
20	91.0	73.1
25	108.7	85.1
30	125.3	95.5

The width of the zone of Postglacial lavas is quite variable but mostly in the range 50–100 km. From the table we may then infer that the corresponding σ_2 values are likely to be in the range 10–30 km. It should be noted that the lava thicknesses (or v_z) chosen have a relatively small effect on the relationship between σ_2 and W_e. There is a second set of σ_2 values consistent with Eq. (1 b) for given values of v_z and W_e, but this gives a much wider distribution of lavas with $\sigma_2 > W_e$, and is not likely to be consistent with the conditions in the volcanic zone of Iceland.

A considerably lower value is indicated for the strain value, σ_1, probably in the range of 5–15 km. The standard deviations probably reach maximum values near central Iceland, where according to Jakobsson (1972) the lava deposition rate is highest, and decrease towards the north and southwest along the ridge.

In the following, a few estimates will be made of σ_2 for lava deposition from the Tertiary lava pile, and compared with the above estimate from the present day volcanic zone.

Model Parameters and the Structure of the Tertiary Lava Pile in Iceland

The Tertiary basalt areas on both sides of the axial zone in Iceland provide an opportunity for a comparison with the model. The deeply incised basalt pile in eastern, northern and western Iceland gives in some areas structural information to a depth of 1–1.5 km from the original top of the pile.

The structural properties which are of particular importance for comparison with the model are (a) the relative dyke volume fraction, (b) the regional dip of the lavas, and (c) the deposition rate of lavas. All these properties vary with depth, and it is therefore necessary to have a method of estimating the depth of observation from the original surface of the lava pile, since the uppermost part of the pile has been eroded away. Walker (1960) used the zeolite zoning of the uppermost crust together with the dyke fraction and its vertical variation. Following Walker we will assume that the top of the analcite zone lies at a depth of 600 m from the original top of the lava pile in other parts of Iceland. This assumption implies that the maximum surface thermal gradient in the crust, which controls the level of the zeolite zones, was the same in other parts of Iceland as in eastern Iceland.

Before going into a detailed comparison of the model with field observations from the Tertiary basalt pile in Iceland, the model

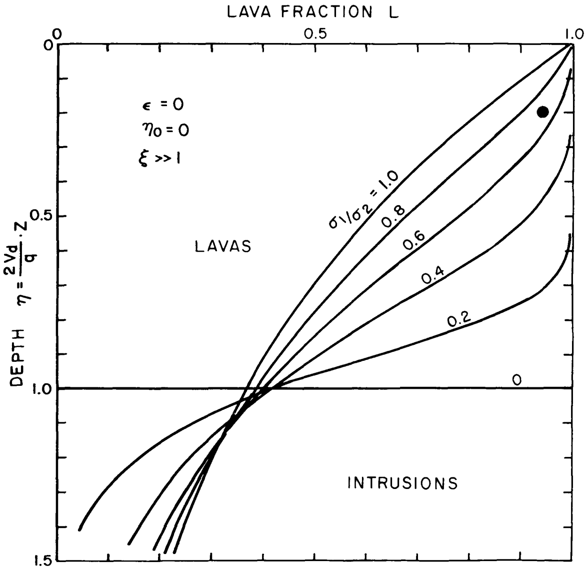

Fig. 3. Lava fraction vs. dimensionless depth in the distant model crust, for various values of the ratio σ_1/σ_2 (crustal strain/lava deposition standard deviations). The normal fault parameters ε and η_0 are discussed in a later section. The *black dot* denotes the average conditions at sea-level in the eastern Iceland lava pile (with $q = 4/3 \times 10^{-4}$ km^2/a and $V_d = 1$ cm/a)

relationships needed for this purpose will be derived. The three structural properties mentioned above, i.e., the dip of the lavas, the deposition rate and the dyke fraction, can all be expressed as functions of depth in the crust, and of the parameters of the model, i.e., V_d, q, σ_1 and σ_2. It turns out that for a constant depth the structural properties depend more ome of the model parameters than on others. The sensitivity of the structural properties to variations in the model parameters will be discussed in some detail. This sensitivity is important when an attempt is made to put constraints on the model parameters from observations of the structural properties in the lava pile.

For the purpose of comparison with the Tertiary lava pile in Iceland we will look only at the structure of th model crust well outside the accretion zone, i.e., for $\xi \gg 1$.

The structural properties can be expressed as functions of the parameter $\xi(0)$, i.e., the horizontal coordinate of the point of origin of a lava trajectory which reaches a depth η in the frozen-in crust. The relationship between η and $\xi(0)$ is given by Eq. (3) with $\xi \gg 1$. This equation gives $\eta = \eta(\xi(0), \sigma_1/\sigma_2)$ and thus contains the ratio σ_1/σ_2 as a parameter, i.e., the crustal strain/lava deposition standard deviations ratio.

The lava fraction in the crust is expressed by Eq. (7). When the velocities v_x and v_z depend only on x this equation reduces to:

$$L = \frac{v_x(x(0))}{v_x(x_1)} = \frac{\operatorname{erf}\left(\frac{\sigma_2}{\sigma_1} \cdot \xi(0)\right)}{\operatorname{erf}\left(\frac{\sigma_2}{\sigma_1} \cdot \xi\right)} \simeq \operatorname{erf}\left(\frac{\sigma_2}{\sigma_1}\ \xi(0)\right). \tag{8a}$$

This formula, together with Eq. (3), thus gives $L = L(\eta, \sigma_1/\sigma_2)$. A family of curves for this relationship is given in Fig. 3. This diagram gives, for an observed value of L at a certain depth z, a relationship between possible values of V_d/q (spreading velocity/total lava deposition rate) and σ_1/σ_2.

The regional dip of the lavas (isochrons) in the upper crust is given by the following formula [Pálmason, 1973; p. 458 Eq. (7a)], assuming that both velocity components depend only on x:

$$\left(\frac{dz}{dx}\right)_t = -\frac{v_z(x(0)) - v_z(x)}{v_x(x)}. \tag{12}$$

In the distant crust well away from the accretion zone ($\xi \gg 1$) this may be written as follows, using the dimensionless variables and the expressions for v_z and v_x.

$$\left(\frac{d\eta}{d\xi}\right)_\tau = -\frac{2}{\sqrt{\pi}} \cdot \exp(-\xi(0)^2). \tag{13}$$

As before, $\xi(0)$ is related to η by Eq. (3). This expression thus gives $(d\eta/d\xi)_\tau$ as a function of η and σ_1/σ_2 (the ratio of standard deviations of crustal strain and lava deposition). In Fig. 4 this relationship is shown for the two limiting cases $\sigma_1/\sigma_2 = 0$ and 1. It is evident that the dip depends little on σ_1/σ_2 in the upper part of the crust. The diagram in Fig. 4 gives, for an observed value of the dip $(dz/dx)_t$ at a certain depth z, a relationship between possible values of V_d/q and σ_2. This relationship depends slightly but not significantly on the ratio σ_1/σ_2.

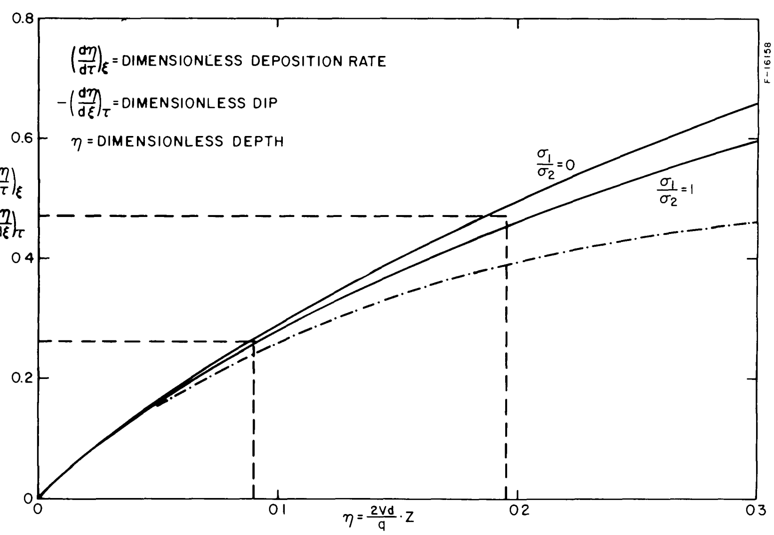

Fig. 4. Dimensionless deposition rate and dip in the distant model crust. The *solid curves* show the range of variation with σ_1/σ_2 (crustal strain/lava deposition standard deviations). The *dot-dashed curve* shows the lower boundary of variation with the normal fault parameters ε and η_0 discussed in a later section

The last structural property, the lava deposition rate $(dz/dt)_x$, may be deduced from equations given by Pálmason (1973). A simple way to find this is to use the mathematical relationship

$$\left(\frac{d\eta}{d\tau}\right)_\xi = -\left(\frac{d\eta}{d\xi}\right)_\tau \left(\frac{d\xi}{d\tau}\right)_\eta \tag{14}$$

Noting that by the definitions of ξ and τ:

$$\left(\frac{d\xi}{d\tau}\right)_\eta = 1, \quad \text{since} \left(\frac{dx}{dt}\right)_z = V_d, \quad (\xi \gg 1)$$

it follows that

$$\left(\frac{d\eta}{d\tau}\right)_\xi = -\left(\frac{d\eta}{d\xi}\right)_\tau = +\frac{2}{\sqrt{\pi}} \cdot \exp(-\xi(0)^2). \tag{15}$$

Thus the same relationship is found between $(d\eta/d\tau)_\xi$ and η as was previously found between $-(d\eta/d\xi)_\tau$ and η, and both are shown by the curves in Fig. 4. In the case of the deposition rate, however, the diagram gives, for an observed value of the deposition rate $(dz/dt)_x$ at a certain depth z, a relationship between possible values of V_d/q and σ_2/V_d.

The curves in Fig. 4 allow constraints to be placed on the model parameters on the basis of observed values of regional dip and deposition rate at certain depths in the lava pile. In this connection it is important to realize that the curves constrain some of the parameters more than others. This is perhaps best seen by imagining that the functions in Fig. 4 were straight lines instead of being slightly curved. Then an observation of the dip at a certain depth in the lava pile would give a unique determination of σ_2, independently of both spreading velocity, V_d, and lava deposition rate per unit length, q. In the same way an observation of the deposition rate at a certain depth would give a unique determination of σ_2/V_d, independently of q. The relatively slight curvature of the functions in Fig. 4 means that they constrain the parameters σ_2 and σ_2/V_d much more than they do the parameter q. The significance of this is that it is possible to make more meaningful statements about possible values of σ_2 and σ_2/V_d on the basis of observations of the dip and deposition rate respectively, than one can make about possible values of q. The parameter q can vary within wide limits with only a small effect on the dip and deposition rate as measured at a certain depth in the lava pile.

It is worth noting finally that the two observable quantities, the regional dip of lavas and the rate of deposition, are closely related to each other. One of them can be deduced from the other if the spreading rate for the particular section of the lava pile is known. Alternatively, the spreading rate can be deduced if these two quantities are both known. This follows from the mathematical relationship mentioned earlier [Eq. (14)]:

$$-\left(\frac{dz}{dt}\right)_x \Big/ \left(\frac{dz}{dx}\right)_t = \left(\frac{dx}{dt}\right)_z \quad (= V_d \text{ for } \xi \gg 1). \tag{16}$$

The model relationships derived above and shown in Figs. 3 and 4 will in the following be used to put some constraints on the model parameters on the basis of structural observations in the Tertiary lava pile of Iceland.

An Estimate of σ_1/σ_2 From the Tertiary Lava Pile in Eastern Iceland

According to the studies of Walker (1959, 1960, 1974) in eastern Iceland of the dyke fraction at various levels in the Tertiary lava pile the average value of D at sea level is about 0.06. The sea level corresponds to a depth of about 1.3 km from the original top of the lava pile according to studies of secondary mineral zones and the upwards decrease of dyke fraction in the pile (Walker, 1960). The diagram in Fig. 3 then gives a relationship between possible values of q/V_d (lava deposition rate per unit length/spreading velocity) and of σ_1/σ_2 (crustal strain/lava deposition standard deviations). If we assume the parameter q/V_d to have had the same value during the formation of the eastern Iceland lava pile, as is indicated at the present time,

$$q = \tfrac{4}{3} \times 10^{-4}\,\text{km}^2/\text{a}, \qquad V_d = 10^{-5}\,\text{km/a},$$

the diagram gives a value of about 0.69 for σ_1/σ_2. The main uncertainty in this estimate with the present model assumptions is in the value of the lava production rate q, since for V_d a relatively steady spreading rate for the Iceland area during the last 47.5 Ma may be deduced from the magnetic anomaly pattern on the Reykjanes Ridge (Herron and Talwani, 1972). A greater lava production rate would give a higher estimate for σ_1/σ_2 and vice versa.

Later on, where a more elaborate version of the model will be discussed, it will be shown that the dyke observations in eastern Iceland cannot constrain the ratio σ_1/σ_2 more than within the range 0.69–1.0.

Three Estimates of the Standard Deviation of Lava Deposition Rate, σ_2, From the Tertiary Lava Pile in Eastern and Western Iceland

The relationships deduced earlier (Fig. 4) between the structural properties and the model parameters will now be used to place constraints on the model parameters on the basis of observations of dip and deposition rate. We will use (a) Walker's observations on the regional dip at sea level in the eastern Iceland Tertiary lava pile, (b) the results of McDougall et al. (1977) on the deposition rate in the Tertiary lavas of central western Iceland, and (c) the results of Watkins and Walker (1977) on the deposition rate in Tertiary lavas in eastern Iceland.

(a) According to Walker (1959, 1960, 1974) the regional dip at sea level of the Tertiary lavas in eastern Iceland is in the range 6–8°. The sea level is at a depth of about 1.3 km from the original top of the lava pile (Walker, 1960). From Fig. 4 we may then deduce a relationship between possible values of σ_2 and q/V_d, according to the model. If we assume that the ratio q/V_d had the same value at the time of formation of the eastern Iceland lavas as it has today, a value of 0.195 is obtained for η. The curves in Fig. 4 then give $(d\eta/d\xi)_\tau = 0.47$, assuming $\sigma_1/\sigma_2 \simeq 0.5$. From the definitions of the dimensionless variables ξ and η we find:

$$\sigma_2 = \frac{z}{\sqrt{2}\left(\frac{dz}{dx}\right)_t} \cdot \frac{\left(\frac{d\eta}{d\xi}\right)_\tau}{\eta}$$

from which we obtain:

$\sigma_2 = 21.1$ km for a dip of 6°

$\sigma_2 = 15.8$ km for a dip of 8°

It may be seen from the above expression for σ_2 that a variation of the other model parameters V_d and q has a relatively small effect on the value obtained for σ_2. This is because the ratio $(d\eta/d\xi)_\tau/\eta$ changes slowly when η is varied by changing the value of q/V_d. Multiplying the value of q/V_d by 2 gives $\sigma_2 = 18.8$–25.1 km instead of 15.8–21.1 km. Dividing it by 2 gives $\sigma_2 = 12.3$–16.5 km. Thus multiplying or dividing q/V_d by 2 gives a variation in σ_2 of about $\pm 20\%$.

(b) In the second example we will estimate σ_2, or rather the ratio σ_2/V_d, from the lava deposition rate $(dz/dt)_x$, as deduced from a study of McDougall et al. (1977) of a lava succession 2–7 Ma old from Borgarfjördur in western Iceland, which has been mapped in detail by Jóhannesson (1975). The samples for K-Ar datings were taken from a 32-km-long profile, at levels ranging mostly ± 200 m around the boundary between the chabazite-thomsonite zone and the analcite zone of secondary mineralization. In order to estimate the depth range of the samples from the original top of the lava pile, it will be assumed that the top of the analcite zone is at a depth of about 600 m, i.e., the same depth as was deduced by Walker (1960) for the lava pile in eastern Iceland. The samples thus are mainly from the depth range 400–800 m. The average rate of lava deposition given by McDougall et al. (1977) was 730 m/Ma, and we will assume for the purpose of calculation that this applies for a depth of 600 m.

With the above values for the lava deposition rate and depth we may use Fig. 4 to deduce a relationship between possible values of σ_2/V_d and q/V_d. If we assume that the ratio q/V_d had the same value at the time of formation of the Borgarfjördur lavas as it has today, a value of 0.09 is obtained for η. The curves in Fig. 4 then give $(d\eta/d\tau)_\xi = 0.261$, assuming $\sigma_1/\sigma_2 \simeq 0.5$. From the definitions of the dimensionless variables ξ and η we find:

$$\frac{\sigma_2}{V_d} = \frac{z}{\sqrt{2}\cdot\left(\frac{dz}{dt}\right)_x} \cdot \frac{\left(\frac{d\eta}{d\tau}\right)_\xi}{\eta}$$

and with inserted values:

$$\frac{\sigma_2}{V_d} = 1.69\ \text{Ma}.$$

With $V_d = 1$ cm/a this gives for the standard deviation:

$$\sigma_2 = 16.9\ \text{km}.$$

As before, the value of σ_2/V_d is rather insensitive to moderate variations in q/V_d. Varying q by a factor of 2, leads to σ_2/V_d values in the range 1.43–1.94 Ma or a variation of about $\pm 15\%$. Varying V_d on the other hand leads to an approximately proportionate variation in σ_2. Thus, observations of the deposition rate of lavas constrain primarily the ratio σ_2/V_d, while q can vary within much wider limits and still be consistent with the observations.

The influence of a possible error in the depth value of 600 m can easily be estimated. For a ± 100 m variation in the depth, i.e., 500–700 m, a range of 14.7–19.1 km is obtained for the standard deviation σ_2, assuming $V_d = 1$ cm/a.

(c) In the third example we will estimate σ_2, or σ_2/V_d, from the results of Watkins and Walker (1977) on the deposition rate of lavas in eastern Iceland. The authors deduced an average lava deposition rate of 620 m/Ma at the top of the analcite zone (assumed to be at a depth of 600 m from the original top of the pile) for a lava sequence 2.0–13.6 Ma old. Corrections were made to take into account that the samples for radiometric dating originated from various depths in the range of about 200–1600 m in the lava pile. The data used included those of McDougall et al. (1976a and b). Some variation in the deposition rate was deduced from the data but this was greatly reduced when the corrections for depth were applied.

Applying the diagram in Fig. 4 to the deposition rate in the same manner as before gives the following:

$$\sigma_2/V_d = 1.98\ \text{Ma}$$

or, with $V_d = 1$ cm/a, $\sigma_2 = 19.8$ km. Allowing for a variation by a factor of 2 in q gives σ_2 values in the range 16.9–22.8 km.

In the three examples discussed above, similar values of σ_2 are obtained, based on data from different localities and crustal sections of partly different age. The values of 14–25 km for σ_2 predicted by the model for the Tertiary lava pile in eastern and western Iceland is in rather good agreement with the situation in the present day zone of rifting and volcanism of Iceland, where postglacial lavas are distributed over a zone some 50–100 km wide. These results may indicate a rather uniform intensity of volcanism in the Icelandic zone for the past 10–15 Ma. It should be emphasized, however, that the deposition rate and the regional dip at a certain depth in the Tertiary lava pile depend only very weakly on q, the total lava production rate per unit length of the axial zone. This is perhaps somewhat unexpected but may easily be ascertained from the curves in Fig. 4. On the other hand, both quantities are rather strongly dependent on the standard deviation σ_2, and in addition the deposition rate is also strongly dependent on the spreading velocity V_d. The conclusion that may be drawn is that it is not possible to infer from measurements of deposition rate and dip about past variations in q. Measurements of dip give primarily information on σ_2, and measurements of deposition rate give primarily information on σ_2/V_d.

Determination of the Spreading Rate From Observations of Regional Dip and Lava Deposition Rate

We will now use Eq. (16) to estimate the spreading velocity from two areas in Iceland where measurements of dip and deposition rate have been made at the same level in the Tertiary lava pile.

As the first example we will take the western Iceland (Borgarfjördur) section of McDougall et al. (1977) discussed earlier. According to the authors the regional dips in the section are in the range 2°–8° towards southeast. Taking the average 5° as a representative value, and the deposition rate 730 m/Ma, we obtain for the spreading velocity

$$V_d = \frac{0.073}{0.087} = 0.84\ \text{cm/a}.$$

In the second example we will take the data on dip and deposition rate given by Watkins and Walker (1977) for a section in eastern Iceland. A mean dip at the top of the analcite zone is 3.9° while the mean deposition rate is 620 m/Ma. This gives for the spreading velocity:

$$V_d = \frac{0.062}{0.068} = 0.91\ \text{cm/a}.$$

Direct measurement of the spreading velocity on the same profile with the $^{40}Ar/^{39}Ar$ method gives according to Ross and Mussett (1976) a value of 'not less than 0.8 cm/a'.

The above estimates of spreading velocities from the Tertiary lava pile on both sides of the axial zone in Iceland are in reasonably good agreement with the spreading rates deduced from magnetic anomalies on the Reykjanes Ridge (Herron and Talwani, 1972).

The Origin of the Visible Tertiary Basalt Pile in Iceland Relative to the Axis of the Accretion Zone

The trajectories of lava elements erupted at the surface in the axial zone give the average paths of such elements from their place of origin at the surface to a final depth in the distant crust. The volcanic products erupted near the axis are buried deep into the crust while those erupted farther from the axis remain closer to the surface (cf. Figs. 1 and 2). The model permits a simple evaluation of the origin with respect to the axis of lavas reaching a certain depth in the crust, provided the model parameters are known.

At the relatively shallow depth in the crust, which we are concerned with here, there is little difference between the trajectories calculated for $\sigma_1/\sigma_2=0$ and $\sigma_1/\sigma_2=1$ as discussed earlier. In Eq. (4) we may put $\xi \gg 1$ for the distant crust, and obtain for the trajectories:

$$\begin{aligned} &\text{I.} \quad \sigma_1/\sigma_2=0; \quad \operatorname{erf}(\xi(0))=1-\eta \\ &\text{II.} \quad \sigma_1/\sigma_2=1, \quad \operatorname{erf}(\xi(0))=\exp(-\eta). \end{aligned} \tag{4a}$$

From these formulas we will estimate the distance from the axis corresponding to the origin of the deepest (sea-level) visible parts of the lava pile in eastern Iceland, assuming the same values of q and V_d as before. The sea level corresponds to a depth of about 1.3 km, which gives $\eta=0.195$. This gives for $x(0)$:

$$\begin{aligned} &\text{I.} \quad \xi(0)=0.916; \quad x(0)=1.30x\sigma_2 \\ &\text{II.} \quad \xi(0)=0.970; \quad x(0)=1.37x\sigma_2. \end{aligned}$$

The standard deviation σ_2 has earlier been estimated to be in the range 14–25 km. Taking 20 km as a representative value gives $x(0)$ in the range 26–27 km. This means, according to the model, that the visible eastern Iceland basalt pile was originally deposited outside an axial zone some 50–55 km wide. The lavas deposited within this zone have been buried to depths in the crust which are below sea level and are thus not accessible to direct observation in the fjords of eastern Iceland.

This result is of considerable interest when comparing the visible structure and surface topography of the Tertiary flood basalt areas with the present-day active volcanic zone. It is well known that there is a conspicuous difference in appearance between the two areas. In the active zone volcanic vents of various forms abound. While some of the topography of the axial zone is no doubt due to the controlling influence of an ice cover during the last glacial period on the behaviour of magma erupted at the surface, it is nevertheless likely that the sheer number and distribution of vents is a more permanent feature of the zone and largely independent of an ice cover. The Tertiary flood basalt areas on the other hand are characterized by a relatively smooth horizontal surface at its highest levels, underlain by horizontal lava flows. Deeper down in the basalt pile signs of eruptive vents become more frequent, dyke swarms and extinct central volcanic complexes begin to appear. At the same time the regional dip of the lava sections increases.

The difference in the appearance of the two main volcanic regions of Iceland is easily understood in terms of the model, and no major change in the behaviour of the volcanic processes in Iceland is required to account for this difference. Assuming that the density of volcanic vents is distributed normally across the accretion zone with the same standard deviation σ_1 as the horizontal strain rate one may easily calculate what fraction of volcanic vents, produced in the axial zone, will be visible in the uppermost 1,300 m of the crust, e.g., in the eroded fjords of eastern Iceland. With $\sigma_1/\sigma_2=0.7$ this is found to be 5.4%, and with $\sigma_1/\sigma_2=1.0$ the fraction is 17.7%. The first value is likely to be more realistic since $\sigma_1/\sigma_2=1$ is a limiting case not likely to correspond to the real situation. In any case, the fraction of the volcanic vents produced in the accretion zone, which would be visible in the uppermost 1,300 m of the crust is small, probably well within 10% of the total produced in the axial zone. The remaining 90% disappear below sea level. This explains the relative scarcity of such vents in the Tertiary flood basalts, not only in eastern Iceland but in other parts of Iceland as well.

Further Modification of the Model. The Effect of Normal Faults

So far only the simplest assumptions have been made regarding the flow field of lava elements in the model. Both the horizontal strain rate dv_x/dx and the vertical velocity v_z have been assumed to be independent of the depth coordinate. This may prove to be an oversimplification, if the intrusive activity associated with the horizontal strain varies with depth. In particular normal faults may take up some of the horizontal strain in the upper part of the crust.

Although normal faults are of relatively minor importance in the structure of the visible Tertiary basalt pile of Iceland (Bodvarsson and Walker, 1964), this should not be taken as an indication that they were uncommon in the axial zone of those times. For the same reason as discussed before for the volcanic vents, most traces of normal faults in the axial zone would have been buried to depths below sea level and thus not be accessible to observation. Normal faults are quite conspicuous in some parts of the present day axial zone in Iceland, in particular in the Thingvellir area in southwest Iceland, and at Tjörnes along the western margin of the active zone in northeast Iceland. The observation of Saemundsson (1967) that the vertical displacements along some of these faults is greater where they cut through older rocks than where they cross younger rocks shows that these faults have been active over a long time, perhaps moving episodically during rifting events such as is taking place in northern Iceland at present (Björnsson et al., 1977). The relative exposure of the normal faults along the axial zone may be related to short term variations in the rate of deposition of lavas along the zone.

In the axial zones of the mid-ocean ridges normal faults appear to be quite common, at least where an axial rift valley is well developed. The normal faults then show up in the topography across the valley (Atwater and Mudie, 1973; Needham and Francheteau, 1974; Tapponnier and Francheteau, 1978; Atwater, 1979).

An attempt will be made to incorporate the normal faults into the model by treating them as a continuous process of horizontal strain and an associated subsidence. It will be assumed that the horizontal strain rate associated with the normal faults is highest at the surface and decreases with increasing depth where dykes and other intrusions take up a larger fraction of the horizontal strain. It is necessary to introduce two new parameters to describe this process. One of them is the fraction ε of the total horizontal strain rate at the surface that is taken up by normal faults. The other, z_0, describes how this part of the strain rate decreases with depth, as will be discussed below.

The volume strain rate associated with the emplacement of intrusions in the crust is given by the divergence of the velocity vector $\bar{v}$ of lava elements. Using the parameters described above this will be assumed to be of the form.

$$\operatorname{div} \bar{v} = \frac{\sqrt{2} \cdot V_d}{\sqrt{\pi} \cdot \sigma_1} \cdot \exp\left(-\frac{x^2}{2\sigma_1^2}\right) \cdot \left(1 - \varepsilon \cdot \exp\left(-\frac{z}{z_0}\right)\right). \tag{2a}$$

The function in the second parenthesis takes into effect the influence of the normal faults. Its form is chosen mainly for reasons of convenience. The parameter ε is the fraction of the total horizontal strain rate at the surface that is taken up by the normal faults, while the parameter z_0 represents the decrease with depth of the influence of the normal faults. For $\varepsilon = 0$ the expression reduces to that given earlier [Eq. (2)].

It will furthermore be assumed that the total horizontal strain rate dv_x/dx has the same form as before and depends only on x:

$$\frac{dv_x}{dx} = \frac{\sqrt{2} \cdot V_d}{\sqrt{\pi} \cdot \sigma_1} \cdot \exp\left(-\frac{x^2}{2\sigma_1^2}\right); \qquad v_x = V_d \cdot \operatorname{erf}\left(\frac{x}{\sqrt{2}\sigma_1}\right). \tag{1a}$$

Then $\dfrac{\partial v_z}{\partial z} = \operatorname{div} \bar{v} - \dfrac{\partial v_x}{\partial x}$, from which it is found by integration that

$$v_z(x,z) = v_z(x,0) - \frac{\sqrt{2} \cdot V_d}{\sqrt{\pi} \cdot \sigma_1} \cdot \varepsilon z_0 \cdot \exp\left(-\frac{x^2}{2\sigma_1^2}\right) \cdot \left(1 - \exp\left(-\frac{z}{z_0}\right)\right). \tag{17}$$

The vertical velocity component at the surface is assumed to have the same form as before:

$$v_z(x,0) = \frac{q}{\sqrt{2\pi} \cdot \sigma_2} \cdot \exp\left(-\frac{x^2}{2\sigma_2^2}\right). \tag{1b}$$

Using the dimensionless variables and introducing the dimensionless velocities

$$U_x = v_x/V_d \quad \text{and} \quad U_z = 2\sqrt{2} \cdot \sigma_2 \cdot v_z/q$$

the flow field components may be written:

$$U_x = \operatorname{erf}\left(\frac{\sigma_2}{\sigma_1} \cdot \xi\right) \tag{18a}$$

$$U_z = \frac{2}{\sqrt{\pi}} \cdot \exp(-\xi^2) - \frac{2}{\sqrt{\pi}} \cdot \frac{\sigma_2}{\sigma_1} \cdot \varepsilon \eta_0 \exp\left(-\xi^2 \cdot \left(\frac{\sigma_2}{\sigma_1}\right)^2\right) \cdot (1 - \exp(-\eta/\eta_0)). \tag{18b}$$

This flow field is consistent with the boundary conditions at the surface, at the axis, and in the distant crust. The introduction of the two new parameters ε and η_0 gives the flow field new properties which were not present in the previously discussed case, where $\varepsilon = 0$. In particular the vertical velocity, which is a function of both ξ and η, can now become zero for certain values of ξ and negative at greater depth. The physical explanation for this is that the lava supply from the surface which makes up the crust has been exhausted at a certain depth, and material from below is brought up to form the lower crust.

The properties of this flow field are perhaps most clearly shown by considering the special case $\varepsilon = 1$ and $\eta_0 \to \infty$. This is an extreme case where all the horizontal strain rate in the upper crust is taken up by normal faults. In this case the vertical velocity component is:

$$U_z = \frac{2}{\sqrt{\pi}} \cdot \exp(-\xi^2) - \frac{2}{\sqrt{\pi}} \cdot \frac{\sigma_2}{\sigma_1} \cdot \eta \cdot \exp\left(-\xi^2 \left(\frac{\sigma_2}{\sigma_1}\right)^2\right). \tag{18c}$$

At the axis $U_z = 0$ for $\eta = \dfrac{\sigma_1}{\sigma_2}$. The trajectory leading from this point on the axis into the crust approaches asymptotically a depth of $\eta = 1$ in the distant crust. This trajectory forms in this special case a sharp boundary between 100% lavas and material brought up from below (intrusions). This is thus another case, in addition to the one mentioned earlier, $\sigma_1/\sigma_2 = 0$, that gives a sharp boundary between lavas and intrusive rocks.

In the less extreme cases where ε and η_0 have lower values, the boundary between predominantly lavas and predominantly intrusions becomes blurred. Lava trajectories, lava isochrons and lava fraction may be calculated numerically as was done for the simpler cases discussed earlier. As an example, the isochrons in the distant crust ($\xi \gg 1$) for the case $\sigma_1/\sigma_2 = 1$ and $\eta_0 \to \infty$ may be calculated from the following equations deduced from Eqs. (5), (18), and (A7) in the Appendix:

$$\eta = \frac{1}{\varepsilon}(1 - \operatorname{erf}(\xi(0))^{\varepsilon}) \qquad \text{Trajectories}$$

$$-\left(\frac{d\eta}{d\xi}\right)_\tau = \frac{2}{\sqrt{\pi}} \cdot \exp(-\xi(0)^2) \cdot (1 - \varepsilon\eta) \quad \text{Isochrons.}$$

The form of the lava isochrons in the distant crust for this case, with $\varepsilon = 1$, is shown in Fig. 5.

The variation of the lava fraction L with depth and its dependence on the parameters ε, η_0, and σ_1/σ_2 is of particular interest for comparison with observations in the Tertiary lava pile of Iceland. The three sets of curves in Figs. 3 and 6 are calculated for the distant crust ($\xi \gg 1$) from formulas derived from Eqs. (7) and (18). The black dots inserted show average conditions in the eastern Iceland lava pile according to Walker (1960) (with $q = \frac{4}{3} \times 10^{-4}\,\text{km}^2/\text{a}$, $V_d = 1$ cm/a).

On the basis of the model curves in Figs. 3 and 6 it is possible to put some constraints on possible model parameters that are compatible with Walker's observations of the dyke fraction in eastern Iceland. Assuming $q = 4/3 \times 10^{-4}\,\text{km}^2/\text{a}$ and $V_d = 1$ cm/a, the following constraints are obtained:

(i) $0.69 < \sigma_1/\sigma_2 < 1$. This is valid regardless of the values of ε and η_0. The lower limit $\sigma_1/\sigma_2 = 0.69$ is obtained for $\varepsilon = \eta_0 = 0$.

(ii) for $\sigma_1/\sigma_2 = 1$ the following limits are found for ε and η_0: $0.74 < \varepsilon < 1$ and $0.35 < \eta_0 < \infty$. The value $\varepsilon = 0.74$ corresponds to $\eta_0 = \infty$, and $\eta_0 = 0.35$ corresponds to $\varepsilon = 1$. For each value of σ_1/σ_2 there is a minimum value of ε corresponding to $\eta_0 = \infty$, and a minimum value of η_0 corresponding to $\varepsilon = 1$.

From Figs. 3 and 6 it may be seen that the normal fault parameters ε and η_0 together with the ratio σ_1/σ_2, have a very pronounced effect on the dyke fraction in the upper part of the

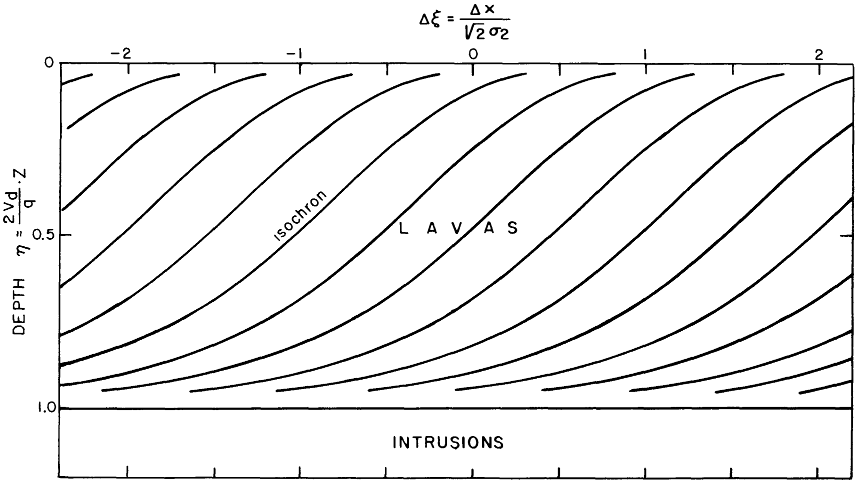

Fig. 5. Lava isochrons in the distant model crust for the extreme case $\sigma_1/\sigma_2 = 1$, $\varepsilon = 1$, $\eta_0 = \infty$. A sharp boundary is obtained between the upper lava layer and the underlying intrusive layer. (ε = relative influence of normal faulting on total crustal strain rate at surface; η_0 = depth constant of this influence)

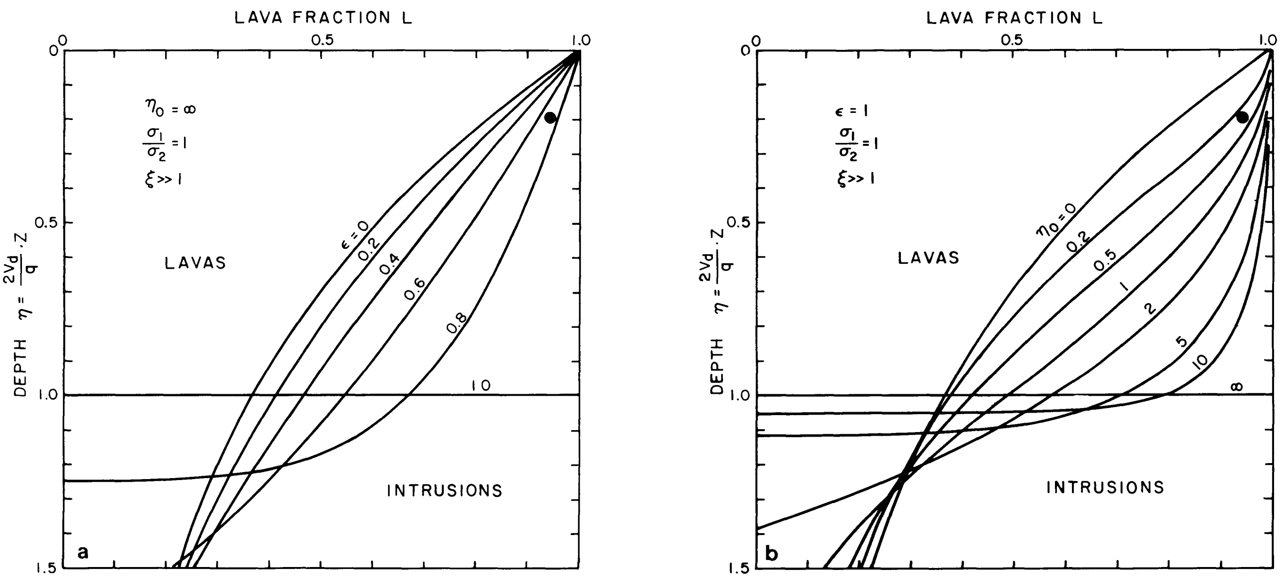

Fig. 6a and b. Lava fraction vs. dimensionless depth in the distant model crust, for various values of the normal fault parameters (**a**) ε and (**b**) η_0. The *black dots* denote the average conditions at sea-level in the eastern Iceland lava pile (with $q = 4/3 \times 10^{-4}$ km^2/a and $V_d = 1$ cm/a). (ε, η_0 see caption of Fig. 5)

model crust. Therefore, only rather broad conclusions regarding the values of individual parameters can be drawn from observations of the dyke fraction in the Tertiary lava pile of Iceland.

Finally, one may ask how the normal fault parameters ε and η_0 will affect the earlier conclusion, that the dip and deposition rate at a certain depth in the upper part of the distant crust (corresponding to the Tertiary lava pile in Iceland) depend mainly on the parameters σ_2 and σ_2/V_d, and only weakly on q. Referring to Fig. 4, the lowest curve (dot-dashed) is obtained for the extreme case $\varepsilon = 1$, $\eta_0 = \infty$ and $\sigma_1/\sigma_2 = 1$. Between this curve and the one labeled $\sigma_1/\sigma_2 = 0$ is the possible range of variation. Considering that the probable range of variation is much smaller, it is evident that the above conclusion regarding the dip and deposition rate remains largely valid. It also follows that the earlier derivations of the parameters σ_2 and σ_2/V_d from the dips and deposition rate measured in various parts of Iceland would not be affected significantly by taking into account the probable range of values for the normal fault parameters ε and η_0.

The normal fault parameters have a major effect on the thermal state of the model crust. This is because extension by dykes implies a strong heat source, while extension by normal faults does not. The detailed discussion of the thermal state of the model crust is, however, outside the scope of this paper.

Discussion

In Iceland regional structural data from two distinct geological environments, the active zone of rifting and volcanism, and the Tertiary flood basalt areas on both sides, can be brought to bear on the process of crustal accretion. A major objective of the present paper has been to analyse such data from both areas in terms of a common model of crustal accretion, to investigate whether they can be fitted to a single kinematic model with a common set of parameters, which would indicate that the accretion process had been going on more or less continuously in the same way it is today, since Iceland began to be formed.

The observational data include regional dips, lava deposition rate, and dyke fraction in the Tertiary areas, width, total lava production rate, and faulting in the active volcanic zone. The analysis in terms of the model shows that from the regional dips (isochrons) of lavas and the local deposition rate of lavas one can obtain estimates of the width of the Tertiary volcanic zone (in terms of the standard deviation σ_2) and the spreading rate. The standard deviation σ_2 is estimated to be in the range 14–25 km, compared to an estimated 10–30 km for its present day value. There is thus no indication that the width of the volcanic zone has changed in a significant way during the last 10–15 Ma. Another result of the analysis is that from measurements of the local deposition rate in the Tertiary lava pile, no conclusions can be drawn about the total lava production rate per unit length of the volcanic zone. This can vary within wide limits without being detected in the uppermost crust. Evidence of volcanic and tectonic processes in the central part of the active zone disappears gradually into the deeper part of the distant crust, where it can not be observed directly. An estimate based on the model indicates that the eastern Iceland visible Tertiary lava pile, some 1,300 m thick, was formed outside a 50–55-km-wide central part of the Tertiary volcanic zone.

An analysis of the factors influencing the dyke fraction in the upper crust, and its variation with depth, indicates that this is influenced both by the ratio σ_1/σ_2 and by the relative importance of dyke injection and normal faulting in the extensional process of the axial zone. Therefore, only rather wide constraints can be put individually on the three parameters describing these processes, from observations of the dyke fraction in the eastern Iceland lava pile.

The same parameters which govern the dyke fraction in the upper crust, also control the transition from an upper lava-dominated layer to a lower intrusion-dominated layer. This transition can be a sharp one, e.g., for $\sigma_1/\sigma_2=0$, as discussed by Kidd (1977). But it is also possible to produce a sharp boundary between lavas and intrusions by a wider extensional zone ($\sigma_1>0$), where normal faults take up the horizontal strain instead of dykes. Both these possibilities should be considered in discussing the relatively sharp boundary observed in some ophiolite sequences. In the Iceland case, likely model parameters would give a gradual transition from the lava layer to the intrusion layer (cf. Fig. 6).

Although the kinematic model appears to correlate fairly well with some regional properties in the active volcanic zone and in the Tertiary flood basalts, care should be taken not to carry the model analogy too far. The simple geometry of the model contrasts rather strongly with the complicated pattern of the present volcanic zones, parts of which have shifted or jumped from one location to another during the geological history of the Iceland area. Structural complications resulting from such jumps are of course not incorporated into the model in its present form. It appears that a volcanic accretion zone has to be active for several million years in the same location to develop the steady-state properties described by the model.

Appendix

A general expression for the isochrons in the model crust may be derived in the following way. A trajectory of lava elements originating at the surface position $\xi(0)$ can be written

$$\eta=\eta(\xi,\xi(0)) \tag{A1}$$

and the time variable (age of lavas) along the trajectory

$$\tau=\tau(\xi,\xi(0)). \tag{A2}$$

Taking the differentials

$$d\eta=\left(\frac{\partial\eta}{\partial\xi}\right)_{\xi(0)} d\xi+\left(\frac{\partial\eta}{\partial\xi(0)}\right)_{\xi} d\xi(0) \tag{A3}$$

$$d\tau=\left(\frac{\partial\tau}{\partial\xi}\right)_{\xi(0)} d\xi+\left(\frac{\partial\tau}{\partial\xi(0)}\right)_{\xi} d\xi(0). \tag{A4}$$

From (A4): $$\left(\frac{d\xi(0)}{d\xi}\right)_{\tau}=-\frac{\left(\frac{\partial\tau}{\partial\xi}\right)_{\xi(0)}}{\left(\frac{\partial\tau}{\partial\xi(0)}\right)_{\xi}}. \tag{A5}$$

From (A3) and (A5):

$$\left(\frac{d\eta}{d\xi}\right)_{\tau}=\left(\frac{\partial\eta}{\partial\xi}\right)_{\xi(0)}-\left(\frac{\partial\eta}{\partial\xi(0)}\right)_{\xi}\cdot\frac{\left(\frac{\partial\tau}{\partial\xi}\right)_{\xi(0)}}{\left(\frac{\partial\tau}{\partial\xi(0)}\right)_{\xi}}. \tag{A6}$$

This is the general differential equation for the isochrons in the model crust. In the distant crust, i.e., $\xi\gg1$, we have

$$\left(\frac{\partial\tau}{\partial\xi}\right)_{\xi(0)}=1 \quad \text{and} \quad \left(\frac{\partial\eta}{\partial\xi}\right)_{\xi(0)}=0,$$

and (A6) reduces to:

$$\left(\frac{d\eta}{d\xi}\right)_{\tau}=-\frac{\left(\frac{\partial\eta}{\partial\xi(0)}\right)_{\xi}}{\left(\frac{\partial\tau}{\partial\xi(0)}\right)_{\xi}}. \tag{A7}$$

References

Atwater, T.: Constraints from the FAMOUS area concerning the structure of the oceanic section. In: Deep Drilling Results in the Atlantic Ocean: Ocean Crust, (M. Talwani, C.G. Harrison, D.E. Hayes, eds.) Am. Geophys. Union, Maurice Ewing Series **2**, 33–42, 1979

Atwater, T., Mudie, J.D.: Detailed near-bottom geophysical study of the Gorda Rise. J. Geophys. Res. **78**, 8665–8686, 1973

Björnsson, A., Saemundsson, K., Einarsson, P., Tryggvason, E., Grönvold, K.: Current rifting episode in north Iceland. Nature **266**, 318–322, 1977

Bodvarsson, G., Walker, G.P.L.: Crustal drift in Iceland. Geophys. J.R. Astron. Soc. **8**, 285–300, 1964

Cann, J.R.: A model for the oceanic crustal structure developed. Geophys. J.R. Astron. Soc. **39**, 169–187, 1974

Daigniéres, M., Courtillot, V., Bayer, R., Tapponnier, P.: A model for the evolution of the axial zone of mid-ocean ridges as suggested by Icelandic tectonics. Earth Planet. Sci. Lett. **26**, 222–232, 1975

Harrison, C.G.A.: Formation of magnetic anomaly patterns by dyke injection. J. Geophys. Res. **73**, 2137–2142, 1968

Herron, E.M., Talwani, M.: Magnetic anomalies on the Reykjanes Ridge. Nature **238**, 390–392, 1972

Jakobsson, S.P.: Chemistry and distribution pattern of Recent basaltic rocks in Iceland. Lithos **5**, 365–386, 1972

Jóhannesson, H.: Structure and petrochemistry of the Reykjadalur central volcano and the surrounding areas, midwest Iceland. Ph.D. Thesis. Durham University, 273 pp., 1975

Kidd, R.G.W.: A model for the process of formation of the upper oceanic crust. Geophys. J.R. Astron. Soc. **50**, 149–183, 1977

Matthews, D.H., Bath, J.: Formation of magnetic anomaly pattern of mid-Atlantic ridge. Geophys. J.R. Astron. Soc. **13**, 349–357, 1967

McDougall, I., Saemundsson, K., Jóhannesson, H., Watkins, N.D., Kristjánsson, L.: Extension of the geomagnetic polarity time scale to 6.5 m.y.: K-Ar dating, geological and paleomagnetic study of a 3,500 m lava succession in western Iceland. Geol. Soc. Am. Bull. **88**, 1–15, 1977

McDougall, I., Watkins, N.D., Kristjánsson, L.: Geochronology and paleomagnetism of a Miocene-Pliocene lava sequence at Bessastadaá, eastern Iceland. Am. J. Sci. **276**, 1078–1095, 1976a

McDougall, I., Watkins, N.D., Walker, G.P.L., Kristjánsson, L.: Potassium-argon and paleomagnetic analysis of Icelandic lava flows: Limits on the age of anomaly 5. J. Geophys. Res. **81**, 1505–1512, 1976b

Needham, H.D., Francheteau, J.: Some characteristics of the rift valley in the Atlantic Ocean near 36° 48′ north. Earth Planet. Sci. Lett. **22**, 29–43, 1974

Pálmason, G.: Kinematics and heat flow in a volcanic rift zone, with application to Iceland. Geophys. J.R. Astron. Soc. **33**, 451–481, 1973

Ross, J.G., Mussett, A.E.: $^{40}Ar/^{39}Ar$ dates for spreading rates in eastern Iceland. Nature **259**, 36–38, 1976

Saemundsson, K.: Vulkanismus und Tektonik des Hengill-Gebietes in SüdwestLIsland. Acta Nat. Isl. **II** (7), 195 pp., 1967

Tapponnier, P., Francheteau, J.: Necking of the lithosphere and the mechanics of slowly accreting plate boundaries. J. Geophys. Res. **83**, 3955–3970, 1978

Walker, G.P.L.: Geology of the Reydarfjördur area, eastern Iceland. Q. J. Geol. Soc. London **114**, 367–391, 1959

Walker, G.P.L.: Zeolite zones and dyke distribution in relation to the structure of the basalts in eastern Iceland. J. Geol. **68**, 515–528, 1960

Walker, G.P.L.: The structure of eastern Iceland. In: Geodynamics of Iceland and the North Atlantic Area, L. Kristjánsson, ed.: pp. 177–188. Dordrecht: Reidel 1974

Watkins, N.D., Walker, G.P.L.: Magnetostratigraphy of eastern Iceland. Am. J. Sci. **227**, 513–584, 1977

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