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KINEMATICS AND HEAT FLOW IN A VOLCANIC RIFT ZONE, WITH APPLICATION TO ICELAND

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CONTENTS

SUMMARY

A kinematic model of a volcanic rift zone, based on plate tectonics concepts of crustal accretion is presented. Quantitative relationships between observable parameters of the model are tested by a comparison with estimates from Iceland of the rate of production of extrusive rocks by eruptions, the width of the volcanic zone, the drift velocity of the lithospheric plates, the regional dips of the flood basalts and the rate of increase of dyke volume fraction with depth in the eastern Iceland lava pile.

Approximate calculations of surface heat flow in the volcanic zone of the model and the adjacent lithospheric plates have been made and compared with heat flow observations from various parts of Iceland.

The model appears to describe fairly well certain regional structural properties of the Icelandic lava pile, such as regional dips of the flood basalts and the average dyke distribution with depth. It is also compatible with the general pattern of heat flow values in Iceland. It is necessary, however, to assume that crustal accretion has shifted between at least 2 zones during the time involved in building up the Icelandic lava pile.

Kinematics and Heat Flow in a Volcanic Rift Zone, with Application to Iceland.

Gudmundur Pálmason

1. Introduction.

Within the framework of plate tectonics, an understanding of the mechanism of generation of new crust at diverging plate boundaries is of great importance. It has implications for interpreting the magnetic anomaly pattern observed over large parts of the oceans and it may also be of importance for correctly interpreting the seismic boundaries observed in the oceanic crust. Observational material, on which to base models of accreting plate boundaries, is scanty, since most such zones are in oceanic areas. Iceland may be a sufficiently large land area to permit a valid test of a model to be made by a comparison with various field observations, which are available.

The purpose of this paper is to present a kinematic model of crustal generation in a volcanic rift zone. The model is primarily intended for the Icelandic rift zone, but it may perhaps with some modifications also apply to other rift zones, where rifting and volcanism occur on a comparable scale. In particular the model may be applicable to some mid-ocean ridge segments. The model will be tested primarily by a comparison with Icelandic data on the regional dips in the Tertiary lava pile, the rate of increase of dyke volume fraction with depth, and with heat flow values in various parts of Iceland.

The Icelandic zone of rifting and volcanism forms a rather complicated pattern. A simple model of the kind here presented can only be expected to account for certain regional properties. A detailed description of the zone will not be given here, but the reader is referred to Thorarinsson (1965), Björnsson (1967), Sigurdsson (1970), Ward (1971) and Saemundsson (1973). Fig. 1 shows the main features of the Icelandic zone of rifting and volcanism.

Fig. 1

The studies of Walker and his collaborators on the structure of the Tertiary lava pile in eastern Iceland have been of fundamental importance for an understanding of the structure of the Icelandic lava pile (Walker, 1959, 1960). The regional structure was found to be one of a general dip of the lava flows towards the present volcanic zone. The dip increases downwards in the pile, commonly reaching 6-8° at sea level. Upwards the dip approaches zero at an elevation corresponding to the original top of the lava pile (cf. Fig. 6). This top level was by independent methods inferred to be some 1500 meters above the present sea level. A further property studied by Walker was the relative volume of dykes at various levels in the lava pile. The dykes occur in swarms, the relative dyke volume increasing from zero at the top of the pile to some 6% on the average at sea level. It was stressed by Walker (1960) that the regional dip of the lavas was caused by subsidence as a result of the piling up of large volumes of volcanic products in an active zone.

Bodvarsson (1954) concluded on the basis of theoretical studies that the abnormally high heat flow in many parts of Iceland is caused by intrusions in the lower part of the crust. On the basis of this work and Walker's in eastern Iceland, these authors (Bodvarsson and Walker, 1964) proposed a process of crustal drift in Iceland by dyke injection. Their model is in many respects similar to the model that is presented in this paper, but is less quantitative and not explicitly based on plate tectonics concepts.

Einarsson (1962) has studied the structure of the Tertiary basalt pile in other parts of Iceland. In many respects his results are similar to those of Walker in eastern Iceland. The regional dips are generally smallest in the uppermost part of the pile, and increase downwards, sometimes jumpwise across discordances. The uppermost part of the pile usually has a grey, fresh appearance due to a lack of zeolite infillings.

The directions of the dips, however, do not always show as clear a relationship to the presently active volcanic zones as the eastern Iceland dips do. Einarsson interprets his data differently from what Walker does. According to Einarsson (1962, 1967) the relatively steep dips in the lower part of the pile are due to a major tectonic phase after the main part of the basalt pile was formed. In this "first tectonic phase" flexures and low synclines and anticlines were formed. After this a widespread peneplanation followed, whereupon a "second tectonic phase" set in, a few million years ago, consisting essentially of vertical uplift without tilting, which presumably has affected the whole of the country.

Einarsson (1967) has furthermore summarized existing knowledge of the arrangement of dykes in the Icelandic basalt pile. He concluded that, 1) the overwhelming majority of the dykes seems to have a trend between NNE and NE, 2) in each area the strike of the dykes and that of the dipping basalts is mostly the same, and 3) the dykes nearly always cut the basalt banks at a right angle. It appears that these observations can in a general way be reconciled with the present model as well as with Bodvarsson and Walker's model, if one assumes that some parts of the Icelandic lava pile were generated in volcanic zones which today are inactive.

The regional heat flow pattern around a volcanic rift zone is of considerable importance for correctly assessing the mechanism of crustal accretion. In some models (e.g. McKenzie, 1967 ; Sleep, 1969 ; Sclater and Francheteau, 1970) the rift zone crustal temperatures enter only as postulated boundary conditions. Such models are useful for the lithospheric plates at relatively great distances from the axial zone. Models of the axial zone crustal temperatures, however, have to be based on a consideration of the mechanism of generation of new crust.

2. Description of the model.

The model consists of a two-dimensional volcanic rift zone column of half-width x_o bounded vertically by rigid lithospheric plates. Only the properties of the crust above the depth of the solidus temperature will be considered. The lithospheric plates are assumed to be moving apart perpendicular to the rift zone with a drift velocity v_d. The forces moving the plates are not specified but are assumed to originate in the underlying mantle. As a result of the plate separation, magma is intruded into the rift zone column, partly reaching the surface as lava flows, and partly remaining as dykes or other irregular intrusions. The intrusive activity is of course irregularly distributed in space and time, but for the present purpose it will be described by a density function f(x), which depends on the horizontal coordinate only, and gives the horizontal strain associated with the intrusions. One may envisage the intrusions in the form of vertical dykes, although this may not necessarily correspond to the real geological situation in all cases. The rift zone column will be assumed to be in isostatic equilibrium, so that the building up of volcanic products at the surface leads to an equivalent amount of subsidence. The process of subsidence is of course very irregular and complicated, but for the present purpose it will be described by another density function g(x), which also depends on the horizontal coordinate only. For simplicity the surface of the rift zone will be assumed to be a horizontal plane at the same level as the surface of the lithospheric plates. The assumption of isostatic equilibrium is in Iceland supported by the available gravity data (Einarsson, 1954, and later unpublished gravity work).

It will be assumed that the process of crustal generation is a stationary one and that the parameters of the model are independent of time. Time dependence will be considered only in connection with the cooling of the lithospheric plate as it moves away from the rift zone.

3. Kinematics of the model.

Fig. 2 shows schematically the model used. A mass element originally erupted as lava at the surface position x(o) moves along a trajectory determined by the velocities v_x and v_z, and reaches the plate at a depth z(o), when the horizontal coordinate is x_o. The velocities are described by the two density functions f(x) and g(x), which depend on x only.

Fig. 2

The intensity of dyke injection, or the horizontal strain, is expressed as follows

$$\frac{dv_x}{dx} = A\cdot f(x) \qquad \text{(A=constant)}$$

The subsidence velocity is similarly expressed as

$$v_z = B\cdot g(x) \qquad \text{(B=constant)}$$

Normalizing the density functions so that $\int_0^{x_o} f(\xi)d\xi = \int_0^{x_o} g(\xi)d\xi = 1$ the two velocities can be expressed as follows:

$$v_x = v_d \int_0^x f(\xi)d\xi \qquad (1)$$

$$v_z = \frac{q}{2}\cdot g(x) \qquad (2)$$

where v_d is the drift velocity of the lithospheric plate relative to the axis of the volcanic zone, and q is the volume rate of production of volcanics at the surface per unit length of the zone. Both v_d and q are assumed to be known from field observations and measurements. If erosional effects are of significance in removing material from the volcanic zone this has to be taken into account when estimating q.

3.1 Trajectories of lava mass elements.

Equations (1) and (2) give the trajectories of crustal mass elements originating as lava at the surface. The relationship between the depth z and the horizontal coordinate x_1 is given by

$$z = \frac{q}{2v_d} \int_{x(o)}^{x_1} \frac{g(x)dx}{\int_o^x f(\xi)d\xi} \qquad (3)$$

A special case of some importance is when g(x) = f(x). This means geologically that the ratio of the lava build-up rate (or rate of subsidence) to the intensity of dyke injection in the underlying crust is the same everywhere in the volcanic zone. In this case eqn. (3) becomes

$$z = \frac{q}{2v_d} \cdot \ln \frac{v_x(x_1)}{v_x(x(o))} \qquad (3a)$$

This special case will be used in numerical calculations (Case I-II).

3.2 Dyke volume fraction in volcanic zone and in lithospheric plate.

By considering the material transported between two adjacent trajectories (Fig. 3) from the surface in the volcanic rift zone to a depth z corresponding to the horizontal coordinate x_1, an expression for the volume fraction of dykes at various depths may be obtained. The volume transport rates dQ_1 and dQ_2 in Fig. 3 can be expressed as follows

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Fig. 3

$$dQ_1 = v_z(x(o)) \cdot dx(o)$$

$$dQ_2 = v_x(x_1) \cdot |dz|$$

The dyke fraction D in the volcanic zone can be expressed as follows

$$D = \frac{dQ_2 - dQ_1}{dQ_2} = 1 - \frac{dQ_1}{dQ_2}$$

From the above expressions for dQ_1 and dQ_2 and by using eqn. (3) one obtains for the dyke fraction

$$D = 1 - \frac{v_x(x(o))}{v_x(x_1)} \tag{4}$$

For the special case f(x) = g(x) one obtains by using eqn. (3a)

$$D = 1 - \exp\left(-\frac{2v_d \cdot z}{q}\right) \tag{4a}$$

This result is independent of the horizontal coordinate and thus gives the same relative dyke volume at the same depth everywhere in the volcanic zone. Since conditions in the lithospheric plate are the same as at the boundary of the volcanic zone this result also applies to the lithospheric plate, provided the assumption f(x) = g(x) is valid.

The rate of increase of dyke volume with depth at the surface is obtained from eqns. (4) and (3)

$$\frac{dD}{dz} = \frac{dD}{dx(o)} \cdot \frac{dx(o)}{dz} = \frac{v_d \cdot f(x(o)) \cdot v_x(x(o))}{v_x(x_1) \cdot v_z(x(o))}$$

As $z \to o$, then $x(o) \to x_1$ and the expression becomes

$$\left(\frac{dD}{dz}\right)_{z=o} = \frac{2v_d}{q} \cdot \frac{f(x_1)}{g(x_1)} \qquad (5)$$

The rate of increase of dyke volume with depth in the lithospheric plate is obtained by setting $x_1 = x_o$ in eqn. (5). If both f(x) and g(x) approach zero at the boundary of the volcanic zone, the limiting value for their ratio should be used.

The assumption f(x) = g(x) may not be valid, especially near the edge of the volcanic zone, as the surface lavas can be expected to flow over a larger area than the main zone of dyke injection. Deviations from this assumption will critically affect the dyke volume fraction in the lithospheric plate, especially near the surface. In order to investigate this effect, the case of a narrower dyke injection zone of half-width $x_0' = \beta \cdot x_0$ ($\beta<1$) will be considered. For simplicity the function f(x) will be taken as constant for $0<x<x_0'$ and zero for $x>x_0'$. The function g(x) will also be assumed to be constant for $0<x<x_0$. The details of the calculations will not be given here, but the dyke volume fraction in the lithospheric plate is found to be

$$\left.\begin{aligned} D &= 0 && \text{for } z < z' \\ D &= 1 - \exp\left\{-\frac{2v_d}{\beta\, q}(z - z')\right\} && \text{for } z > z' \end{aligned}\right\} \qquad (6)$$

where $z' = \frac{q}{2v_d} \cdot (1 - \beta)$

Examples of calculations of dyke volume fractions according to these formulas are given in Fig. 7 for various values of β. It is of interest to note that at a depth of $z = \frac{q}{2v_d}$ the dyke fraction according to eqn. (6) is 0.63, independently of the value of β. At this depth there will be a very rapid increase in the dyke fraction provided $\beta << 1$.

3.3 Lava isochrons in volcanic rift zone and in lithospheric plate.

The lava isochrons in the volcanic rift zone are obtained from eqns (1), (2) and (3). The differential equation for the isochrons is found to be

$$\frac{dz}{dx} = - \frac{v_z(x(o)) - v_z(x)}{v_x(x)} \qquad (7)$$

with the boundary condition $z = v_z(o) \cdot t$ at $x=0$.
Here $x(o)$ is related to x and z by eqn. (3), writing x instead of x, in eqn. (3).

At the boundary of the volcanic zone $x=x_0$ eqn. (7) becomes

$$\left(\frac{dz}{dx}\right)_{x=x_0} = - \frac{v_z(x(o))}{v_d} = - \frac{q}{2v_d} \cdot g(x(o)) \qquad (7a)$$

where $x(o)$ is related to z by eqn. (3) with $x_1=x_0$.

In the lithospheric plate the isochrons move horizontally with velocity v_d, and eqn. (7a) may therefore be regarded as the differential equation for the isochrons in the plate with the boundary condition $x = x_0+v_d \cdot t$ at $z=0$.

3.4 Dips of lava flows (lava isochrons) in lithospheric plate.

The dip of the lava isochrons at various depths in the plate is obtained directly from eqn. (7a). If ϕ is the dip angle one obtains

$$\operatorname{tg} \phi = -\left(\frac{dz}{dx}\right)_{x=x_0} = \frac{q}{2v_d} \cdot g\,(x(o)) \qquad (7b)$$

where $x(o)$ is related to z by eqn. (3) with $x_1 = x_0$.

The dip angle approaches zero at the surface in the lithospheric plate, if $g(x) \to 0$ as $x \to x_0$. The expression (7b) shows that the regional dips in the upper part of the plate depend primarily on the behaviour of the function $g(x)$, i.e. the distribution of surface lavas, and the ratio q/v_d. The dips would not be significantly affected if the zone of dyke injection were narrower than the zone of subsidence (zone of lava flows).

3.5 Numerical calculations for two cases.

To exemplify the formulas that have been given, numerical calculations will be made for two cases. In both cases the functions $f(x)$ and $g(x)$ will be assumed to be equal. In Case I it will be assumed that both functions are constant, i.e. $f(x) = g(x) = 1/x_0$. This is clearly an oversimplification but may nevertheless be of some value. In Case II the functions are chosen to be of a parabolic form. This gives a better agreement with field observations of regional dips, and avoids the somewhat unnatural discontinuity in the subsidence velocity at the boundary of the subsidence zone.

Table 1 gives a summary of the formulas for numerical calculations for the two cases. The time variable is expressed in a dimensionless form

$$\tau = \frac{v_d}{x_0} \cdot t$$

Fig. 4

TABLE 1

Model calculations for two cases

(v.r.z. = volcanic rift zone ;

l.p. = lithospheric plate)

Property	Case I	Case II
$f(x)$ $(=g(x))$	$1/x_0$	$\frac{3}{2x_0}\left(1-\left(\frac{x}{x_0}\right)^2\right)$
Trajectories in v.r.z	$x=x(o)\cdot\exp\left(\frac{2v_d}{q}\cdot z\right)$	$\frac{x}{x_0}\left(3-\left(\frac{x}{x_0}\right)^2\right) = \frac{x(o)}{x_0}\left(3-\left(\frac{x(o)}{x_0}\right)^2\right)\cdot\exp\left(\frac{2v_d}{q}\cdot z\right)$
Lava isochrons in v.r.z.	$z=\frac{q}{2v_d}\cdot\tau$	$\exp\left(\frac{4}{3}\frac{v_d}{q}\cdot z\right) = \left(1-\frac{1}{3}\left(\frac{x}{x_0}\right)^2\right)\cdot\exp(\tau)+\frac{1}{3}\left(\frac{x}{x_0}\right)^2\cdot\exp(-2\tau)$
Lava isochrons in l.p.	$z=\frac{q}{2v_d}\left(1-\frac{x}{x_0}+\tau\right)$	$\exp\left(\frac{4}{3}\frac{v_d}{q}\cdot z\right) = \frac{2}{3}\exp\left(1-\frac{x}{x_0}+\tau\right)+\frac{1}{3}\cdot\exp\left(-2\left(1-\frac{x}{x_0}+\tau\right)\right)$
Slope of lavas in l.p.	$\frac{dz}{dx}=\frac{q}{2v_d\cdot x_0}$	$\frac{dz}{dx} = -\frac{q}{2v_d\cdot x_0}(\exp(\tau)-\exp(-2\tau))\cdot\exp\left(-\frac{4}{3}\frac{v_d}{q}\cdot z\right)$
		where $2\exp(\tau)+\exp(-2\tau) = 3\cdot\exp\left(\frac{4}{3}\frac{v_d}{q}\cdot z\right)$
Dyke volume fraction in l.p. and v.r.z.	$D = 1-\exp\left(-\frac{2v_d}{q}\cdot z\right)$	$D = 1-\exp\left(-\frac{2v_d}{q}\cdot z\right)$

Fig. 4 shows the calculated lava isochrons and trajectories for the two cases, assuming a value of 1 cm/yr for v_d and $4/3 \times 10^{-4}$ km^2/yr for q. The latter value is based on various estimates that have been made of the average rate of production of volcanics in Iceland (Einarsson, 1954 ; Bödvarsson and Walker, 1964 ; Thorarinsson, 1967 ; Jakobsson, 1972)

The dyke volume fraction is identical for both cases, and is shown graphically in Fig. 7 (curve 1).

4. Comparison of the kinematic model with observations in Iceland.

One test of the kinematic model is provided by a comparison of the quantitative relationships given in Table I with certain observations that are available from Iceland. For a comparison with the lithospheric plate part of the model, the eastern Iceland area is particularly suitable, as it is well outside the active volcanic zone and has been studied in relatively great detail.

In the volcanic zone estimates have been made of the present rate of production of volcanics (cf. sect. 3.5). Some data are available on the rate of subsidence in the volcanic zone (Tryggvason, 1968, 1970 ; Björnsson et al., 1972). They indicate a rate of a few millimeters per year, which is consistent with the assumption made earlier in the kinematic model on the basis of gravity data, that the subsidence rate is equivalent to the production rate of volcanics from eruptions.

There are essentially three parameters, q, v_d and x_0, which characterize the model. Derived properties which can be observed in the eastern Iceland lava pile include the regional dip of the flood basalts and the rate of increase of dyke volume fraction with depth. These observations refer to conditions at the time of formation of the plate segments involved. Observations of the primary parameters, q, v_d and x_0 refer to their values at the present time. The assumption is made that these parameters have not changed significantly during the time period involved in the formation of Iceland. This assumption is not without support from the magnetic pattern on the Reykjanes Ridge (Heirtzler et al., 1966 ; Talwani et al., 1971), as far as the drift velocity is concerned.

According to Walker (1960) the regional dip of the eastern Iceland flood basalts at sea level is 7-8° towards the volcanic zone. The sea level corresponds to a depth of about 1500 meters from the original top of the pile according to Walker. By using the formula in Table I, Case II, for the slope of the lavas in the lithospheric plate, and assuming a drift velocity of 1 cm/yr, one obtains the relationships between the parameters q and x_0 which are shown in Fig. 5 for three values of the dip angle at sea level. The present average rate of production $q = 4/3 \times 10^{-4}$ km^2/yr corresponds to a half-width of $x_0 = 51$ km for a dip angle of 7°.

Fig. 5

When comparing this with the present-day conditions it is necessary to keep in mind that the volcanic rift zone is here defined as the zone over which surface lavas flow, since this is the area affected by subsidence. The main intrusive activity may be confined to a considerably narrower zone.

In this sense the total width of the volcanic zone in northern Iceland is about 70 km. In southern Iceland the total width is larger, about 100 km or more. The agreement between the present-day observed width and that deduced above from the basalt pile in eastern Iceland is thus relatively good.

Fig. 6

Fig. 6 shows a section through the lava pile in eastern Iceland according to Walker (Bodvarsson and Walker, 1964). The model sections below are drawn in the same scale and show isochrons at 1 Myrs intervals, assuming the parameters of the model to be $x_0 = 50$ km, $v_d = 1$ cm/yr and $q = 4/3 \times 10^{-4}$ km^2/yr.

The rate of increase of dyke volume fraction with depth has been studied in eastern Iceland by Walker and his collaborators (Walker, 1960 ; Gibson et al., 1966). The dykes occur predominantly in swarms, but when averaged over horizontal sections of several tens of kilometers, the average rate of increase of dyke volume with depth is about 4% per kilometer in the uppermost 1500 meters of the pile. Gibson and Piper (1971) have argued on the basis of the swarm distribution that the 4% rate of increase gives too low an estimate for the dyke volume fraction at greater depth.

The model calculations in Table 1 give a rate of increase of dyke volume fraction

$$\left(\frac{dD}{dz}\right)_{z=o} = \frac{2v_d}{q} = 0.15 \text{ km}^{-1}, \text{ or } 15\% \text{ per km.}$$

This is nearly four times the average observed rate. The discrepancy is, however, not difficult to explain. According to the discussion in sect. 3.2 the rate of increase of dyke volume fraction with depth is critically affected by the ratio of the functions f(x) and g(x) in the outer part of the volcanic zone. The assumption f(x) = g(x), on which the formulas in Table 1 are based, can lead to too high estimates for the rate of increase of dyke volume with depth. This is shown very clearly by eqn. (6) which gives the dyke volume behaviour if the dyke injection zone is narrower than the zone of subsidence.

Fig. 7

A comparison of the calculated curves in Fig. 7 with the observed rate of increase of dyke volume fraction with depth shows that good agreement can be obtained by allowing the main dyke injection zone of the model to be narrower than the zone of lava flows and subsidence. It is necessary, however, to assume that some small dyke injection also takes place in the outer part of the subsidence zone, as otherwise no dykes would be observed in the uppermost part of the pile. The assumption of a narrower dyke injection zone is in general agreement with conditions in the present-day volcanic zone of Iceland.

It appears from the above comparisons with the eastern Iceland basalt pile, that the kinematic model gives a satisfactory quantitative description of the regional dips of the flood basalts as well as the average dyke volume relationships in the upper part of the lava pile, if the dyke injection zone is assumed to be narrower than the zone of lava flows.

5. Heat flow in the volcanic rift zone and the adjacent lithospheric plates.

The kinematics of the model allow some deductions to be made about the crustal temperature field in the volcanic rift zone and the adjacent lithospheric plate. It will be assumed that the dyke intrusions are the only significant heat sources in the volcanic zone. Other heat sources, such as hydrothermal alteration reactions, may be of importance also, but will not be considered here, as reliable estimates of the rate of heat release by such reactions are apparently not available (Arnórsson, 1972). The effect of radioactivity is negligible and will not be considered.

5.1 Heat flow in the volcanic rift zone.

The crustal temperature field in the volcanic zone of the model is described by the equation of heat conduction in a moving medium with heat sources (Carslaw and Jaeger, 1959, p. 13)

$$\alpha \nabla^2 T + \frac{A}{\rho c} - \bar{v} \cdot \nabla T = \frac{\partial T}{\partial t} \qquad (8)$$

where α = thermal diffusivity $(= \frac{k}{\rho c})$, T = temperature, A - heat source function, $\bar{v}$ velocity of crustal mass elements, t = time, k = thermal conductivity, ρ = density and c = specific heat.

It will be assumed here that the volcanic zone is a stationary phenomenon and that steady-state temperature has been reached, i.e. that $\frac{\partial T}{\partial t} = 0$. It would be possible to solve eqn. (8) numerically for a given velocity field and boundary conditions. It is more instructive, however, to study how the solution depends on the parameters of the model.

A few simplifying assumptions will be made, which can be expected to be valid in a relatively wide volcanic zone, especially in its central part. The temperature is assumed to depend on the vertical coordinate only, i.e. lateral conduction is neglected. It will furthermore be assumed that the functions f(x) and g(x) are constant, i.e. $f(x) = g(x) = 1/x_0$.

The heat source function A can be written approximately as follows

$$A = n\rho c \frac{dv_x}{dx}(T_k - T)$$

Here $T_k = T_l + \frac{L}{c}$, where T_l, the liquidus temperature of basalt, will be taken as 1200°C, and L is the latent heat of fusion. The factor n (>1) is introduced to take into account the heat conducted from dykes into the adjacent rock during volcanic eruptions. This is to be added to the heat content of the magma in the dyke, which is conducted to the surrounding rock at the end of the magma flow in the dyke. The factor n which depends primarily on the average dyke width and the duration of magma flow in the dyke channel can be estimated from heat conduction theory. For a normal dyke width of 3 meters and a duration of magma flow of 1 week the factor n is about 1.5. For a duration of 1 month it is about 1.9 and for 1 year about 3.2. On the basis of observations in Iceland, both of historic eruptions (Thorarinsson, 1968) and rock magnetism around dykes (Kristjánsson, 1970), it appears likely that the duration of flow in one channel rarely exeeds 1 month, although the total duration of an eruption may be considerably longer. A value of n = 2 will be used in the following calculations.

With the assumptions made above, and inserting appropriate expressions for A and v, eqn. (8) becomes

$$\frac{d^2T}{dz^2} - a\frac{dT}{dz} + b\,(T_k - T) = 0 \qquad (8a)$$

where $a = \frac{q}{2\alpha x_0}$ and $b = \frac{nv_d}{\alpha x_0}$

With the boundary conditions T = 0 and $\frac{dT}{dz} = g_0$ at z = o, the solution can be written as follows

$$T = T_k - \frac{T_k}{2}\left\{(h+1)\exp\{-(\sqrt{\tfrac{a^2}{4}+b} - \tfrac{a}{2})z\} - (h-1)\exp\{(\sqrt{\tfrac{a^2}{4}+b} + \tfrac{a}{2})z\}\right\} \qquad (9)$$

$$\text{where} \quad h = \frac{\frac{g_0}{T_k} + \frac{a}{2}}{\sqrt{\frac{a^2}{4} + b}} \qquad (10)$$

It is readily seen from the second exponential term that, for relatively large values of z, the solution is very sensitive to variations in the parameter h. To show this the curves in Fig. 8 have been drawn, using the following numerical values: $q = 4/3\times10^{-4}$ km^2/yr , $v_d = 10^{-5}$ km/yr , $x_0 = 50$ km , $\alpha =$ 2×10^{-5} km^2/yr , L = 100 cal/g , c = 0.25 cal/g °C , $T_k =$ $1200 + \frac{100}{0.25} = 1600$°C .

Fig. 8

Curves which do not reach the melting range of basalts obviously are not physically acceptable. This sets for the parameter h a lower limit, which is only a few per cent below the value unity. Curves with values of h appreciably greater than unity would be physically acceptable in the crust shallower than the solidus temperature, but would probably not be consistent with realistic boundary conditions across the solidus. The temperature can be expected to increase slowly below the depth of the solidus temperature where partial melting occurs, and the boundary conditions probably require the temperature gradient to be relatively small above the solidus also, suggesting an h-value close to unity.

The above discussion leads to the conclusion that if the temperature function, eqn. (9), is to be physically acceptable, the parameter h is confined to a relatively narrow interval around the value unity. Adopting the value h = 1 leads directly to a calculation of the surface gradient from the expression (10). In this case the temperature according to (9) becomes

$$T = T_k(1 - \exp(- \frac{g_0}{T_k} \cdot z)) \qquad (9a)$$

and the surface gradient from (10) becomes, when the expressions for a and b are substituted

$$g_0 = \frac{q \cdot T_k}{4\alpha x_0} \left\{ h \sqrt{1 + \frac{16 n v_d \alpha x_0}{q^2}} - 1 \right\} \qquad (10a)$$

Fig. 9

The gradient is shown in Fig. 9 as a function of x_0. Numerical values used are the same as in Fig. 8. For a half-width of 50 km, as indicated by the eastern Iceland data, and with h = 1, the surface gradient is found to be 180°C/km. With a thermal conductivity of 1.9 W/m°C (0.0045 cal/cm sec °C this corresponds to a heat flow of 340 mW/m² (about 8.1 HFU).

When comparing the surface gradient (10a) with observed gradients in Iceland, it is necessary to note that the heat discharged in the geothermal areas derives from the process of crustal heating by intrusive activity. It is thus included in the heat flow calculated above. Bödvarsson (1961) estimated the total steady heat discharge from geothermal areas in Iceland to be 10^9 cal/sec, of which 90% is contributed by the high-temperature areas in the volcanic zone. Taking the half-width of the volcanic zone as 50 km, this heat flow is equivalent to an average gradient of 67°C/km, if the heat were transported by conduction. Subtracting this from the calculated gradient gives a value of 113°C/km, corresponding to a heat flow of about 215 mW/m^2 (5.1 HFU), as an estimate for the conductive part of the heat transport to the surface in the Icelandic volcanic zone. A comparison of these estimates with observed heat flow values will be discussed in sect. 6.

The temperature function (9a) is valid only for temperatures below the solidus temperature. For approximate calculations, e.g. of the cooling of the lithospheric plates, it will be used at higher temperatures also, with the understanding that it represents a hypothetical temperature corresponding to a constant specific heat of the partially molten lower lithosphere.

5.2 The decay of the temperature field in the lithospheric plate.

It is of some importance to estimate how the surface temperature gradient in the model falls off with distance from the volcanic rift zone. This phenomenon can be observed and a further test is thereby provided on the model under discussion.

As crustal segments from the active zone join the lithospheric plates, the intrusive activity ceases as well as the subsidence. An estimate of the decrease of the surface gradient with distance from the volcanic zone may then be obtained by calculating how the temperature function given by eqn. (9a) changes with time after the heat source function and the subsidence both vanish. The spreading rate thereafter gives the variation with distance. The same calculations also apply if an active volcanic zone becomes inactive, and give the decrease of the surface temperature gradient with time. In all these calculations the horizontal heat flow is neglected compared to the vertical one.

From formulas given in Carslaw & Jaeger (1959, p.59) it may be shown that if an initial temperature distribution is given by eqn. (9a), then the temperature at a time t thereafter, in a medium without heat sources, is given by the following expression

$$T(z,t) = T_k \{\mathrm{erf}(\frac{z}{2\sqrt{\alpha t}}) + \exp(p^2\alpha t)\cdot\sinh(pz) +$$

$$+ \frac{1}{2}\exp(p^2\alpha t)\cdot\left[\exp(-pz)\cdot\mathrm{erf}(p\sqrt{\alpha t} - \frac{z}{2\sqrt{\alpha t}}) - \exp(pz)\cdot\mathrm{erf}(p\sqrt{\alpha t}+\frac{z}{2\sqrt{\alpha t}})\right]\} \tag{11}$$

where $p = g_0/T_k$ and g_0 is the initial surface gradient. The temperature function (11) is shown in Fig. 10 for various values of the time, assuming $g_0 = 180°C/km$. From eqn. (11) the surface gradient g_t at a time t is found to be

$$g_t = g_0 \cdot e^{y^2} \cdot \mathrm{erfc}\ y \tag{12}$$

where $y = \frac{g_0}{T_k}\cdot\sqrt{\alpha t}$

The function on the right side of eqn. (12) is tabulated in Carslaw & Jaeger (1959, p. 485). For large values of t the surface gradient approaches the function

$$g_t = \frac{T_k}{\sqrt{\pi\alpha t}} \tag{12a}$$

which is independent of the original surface gradient. Eqn. (12a) is the expression for the surface gradient of a semi-infinite solid originally at a temperature T_k.

Fig. 10

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6. Comparison with heat flow observations in Iceland.

The relatively shallow crustal temperature field in Iceland is in many places disturbed by the movement of water. Such disturbances may reach a depth of 2 km or more, and are usually associated with surface manifestations of thermal activity. Cases are known, however, where a high surface gradient without surface thermal activity has by drilling been shown to be due to water movement at depth. Such conditions may be expected in tectonically active areas where the permeability of the rocks may be relatively great. In the active volcanic rift zones the high permeability of the near-surface volcanic rocks also often leads to disturbances of the shallow temperature field by cold ground-water movement.

These observations show that heat flow data which are to be used to predict temperatures at a depth of several kilometers have to be very carefully selected. Single boreholes can give erroneous results, but if consistent results are obtained from several boreholes in an area extending over several tens of kilometers, one may have some confidence in the results. In eastern Iceland which is practically without thermal activity, significant results may be expected from relatively few boreholes.

The latest compilation of regional heat flow values from Iceland was given by Pálmason (1967). Several new holes, which appear to be of significance, have been drilled since. Some of them have been drilled specifically to obtain heat flow data. They are usually 100 meters deep. Fig. 11 shows the available temperature gradient data, which are considered significant for predicting crustal temperatures.

Fig. 11

From the available data on the thermal conductivity of basalts it appears that a conductivity of 1.9 W/m°C (0.0045 cal/cm sec °C) can be used for estimating the corresponding heat flow values. Table 2 gives some additional data on the holes. Boreholes shallower than 90 meters have been ommitted as they are considered less reliable than the deeper ones.

Fig. 12 shows heat flow values from southern and western Iceland projected on the line A-B in Fig. 11, which is roughly perpendicular to the main trend of the volcanic zones in southern Iceland. A zone of maximum heat flow near the Reykjanes-Langjökull volcanic zone is clearly indicated. Equally conspicuous is the absence of a corresponding conductive heat flow anomaly associated with the eastern zone in southern Iceland. This is noteworthy since this zone has in historical times (about 1100 years) produced a far greater volume of volcanic rocks by eruptions than the Reykjanes - Langjökull zone (Thorarinsson, 1967).

Since the eastern volcanic zone does not seem to have a major conductive heat flow anomaly associated with it, it is tempting to look at the available heat flow data on the assumption that the Icelandic flood basalt pile was primarily generated in the Reykjanes - Langjökull volcanic rift zone. Following a suggestion of Saemundsson (1967) it will then be assumed that a zone to the north of Langjökull, schematically shown in Fig. 11, was an active zone of spreading and volcanism until a few million years ago when it became inactive. Recent K/Ar age determinations (Everts et al., 1972) support the relatively young age of rocks from this area. Of thirteen determinations nine give an age between 1 and 3 million years. For the present purpose it will be assumed that volcanic activity ceased in this zone 3 million years ago.

Fig. 13

TABLE 2

Heat flow data from boreholes in Iceland.

Borehole	Latitude N	Longitude W	Elevation m	Depth m	Gradient °C/km	Surface heat flow[1] mW/m^2	(HFU)
Arnarholt	64°14'.8	21°51'.9	20	240	165	315	(7.5)
Ferstikla	64°24'.3	21°36'.2	(5)	100	145	275	(6.6)
Akranes I	64°19'.4	22°04'.6	(5)	1400	129	245	(5.9)
Akranes II	64°18'.3	21°56'.2	(5)	100	153	290	(6.9)
Akranes III	64°22'.0	21°57'.9	(5)	100	150	285	(6.8)
Borgarnes	64°33'.7	21°55'.6	(5)	100	109	205	(4.9)
Stadastadur	64°48'.4	23°01'.0	(5)	100	66	125	(3.0)
Tindar	65°18'.8	22°13'.6	15	105	111	210	(5.0)
Hvammstangi	65°23'.9	20°56'.8	(10)	103	70	133	(3.2)
Hólar	65°43'.9	19°07'.6	140	103	58	110	(2.6)
Akureyri	65°41'.1	18°06'.1	40	100	64	120	(2.9)
Eidar	65°22'.9	14°20'.3	40	100	37	70	(1.7)
Hvalnes	64°25'.5	14°33'.1	30	110	45	85	(2.0)[2]
Prestbakki	63°49'.4	18°02'.6	40	100	47	90	(2.1)
Vestmannaeyjar	63°26'.7	20°17'.2	20	1565	63	120	(2.9)
Búrfell (BH-14)	64°06'.5	19°48'.2	247	160	51	95	(2.3)
Thykkvibaer	63°44'.7	20°37'.5	10	90	93	175	(4.2)
Eyrarbakki	63°51'.9	21°09'.2	(5)	752	88	165	(4.0)
Árbaer	63°56'.7	21°02'.8	20	937	142	270	(6.5)[2]

1) Thermal conductivity assumed 1.9 W/m °C. 1 HFU = 1 microcal/cm^2sec.

2) Borehole partly disturbed by water flow.

In Fig. 13 the thermal gradients from Fig. 11 have been plotted as a function of the age of the respective crustal segments assuming a spreading rate of 1 cm/yr and a direction of spreading as shown in Fig. 11. Theoretical curves for the decay of the surface gradient according to eqn. (12) are also shown, for initial gradients between 100 and 250°C/km. The trend of decreasing heat flow with distance from the Reykjanes-Langjökull-Skagi zone seems clear, but the majority of the values are slightly higher than the theoretical curves which are based on a drift velocity of 1 cm/yr. The discrepancy is, however, not greater than can be expected considering the assumptions made. A wider volcanic zone would give a better agreement between observations and calculated curves.

Perhaps the most important conclusion to be drawn from Fig. 13 is that the eastern Iceland heat flow data seem compatible with the hypothesis that the eastern Iceland basalt pile was generated in the Reykjanes-Langjökull-Skagi zone. They do not appear to be consistent with a crustal generation in the eastern volcanic zone.

It thus appears that the available heat flow data from Iceland with few exceptions are in rather good agreement with the model calculations here presented, if one assumes that the main bulk of the Icelandic lava pile has been generated in the Reykjanes-Langjökull-Skagi zone and that the eastern volcanic zone has been active for only a few million years.

Ade-Hall et al. (1971) have discussed probable temperature gradients in the eastern Iceland lava pile at the time of its formation. They conclude, on the basis of the zeolite zones mapped by Walker (1960) and probable stability fields of various zeolites, that a gradient of 115°C/km existed in the lava pile at the time of formation of the zeolite zones. This is in agreement with the theoretical curves in Fig. 13 if the zeolite zones were formed within some 1-2 million years after volcanic activity came to an end, which corresponds to a distance of 10-20 km from the volcanic zone, if the drift rate is 1 cm/yr.

7. Total heat transport in the Icelandic volcanic rift zone.

As the model calculations of heat flow appear to be compatible with available observations in Iceland, it is of interest to use both to estimate the relative importance of the various modes of heat transfer in the rift zone. The heat is transported to the surface by volcanism, conduction and geothermal activity. Denoting these three components by H_v , H_c and H_g , respectively, per unit length of the zone, they can be expressed as follows, using eqn. (10a).

$$H_v = q \cdot \rho \cdot c \cdot T_k$$

$$H_c + H_g = k \cdot g_0 \cdot 2x_0 = \frac{H_v}{2} \left\{ \sqrt{1 + \frac{16 \cdot n \cdot v_d \cdot \alpha \cdot x_0}{q^2}} - 1 \right\}$$

Inserting numerical values and using Bodvarsson's estimate of 10^9 cal/sec for the steady heat discharge of the geothermal areas, of which 90% is assumed to occur in the volcanic zone, the following results are obtained.

H_v = 21 MW/km
H_c = 21 "
H_g = 14 "

This gives a total heat transport of 56 MW/km to the surface in the volcanic zone. For comparison a rough estimate can be made of the heat transported into the lithospheric plates, using the temperature function, eqn. (9a). The normal heat flow from the upper mantle will be assumed to be 1.0 HFU, corresponding to a gradient of about 22°C/km. With a drift velocity of 1 cm/yr and a thickness of 50 km of the lithospheric plates, the excess heat transported laterally into the lithospheric plates from the volcanic zone is found to be

H_l = 78 MW/km

This heat is gradually conducted to the surface of the lithospheric plates as they move away from the volcanic zone and cool down.

These estimates are average ones for the Icelandic part of the axial zone of the Mid-Atlantic Ridge. Jakobsson (1972) has shown that the rate of postglacial lava production in Iceland varies along the volcanic zone, being greatest near central Iceland and decreasing to the north and to the south. Similar variations in the other heat transport components may also be expected along the axial zone of the ridge.

8. General discussion of the results.

Despite the complexity of the volcanic zone in Iceland it appears that the model which is based on plate tectonics concepts can give a fair description of the regional structure of the Icelandic lava pile and can account for the regional pattern of heat flow values.

The data from eastern Iceland on the rate of increase of dyke volume fraction with depth, as well as observations of volcanic activity in the Icelandic zone, indicate that the main zone of dyke injection is considerably narrower than the zone of lava flows which is affected by subsidence. This result has only partly been taken into account in the model calculations which have been presented, e.g. not at all in the heat flow calculations. In an improved model this should be taken into account.

The main implications of a narrower dyke injection zone for the structure of the upper crust are as follows. In the dyke injection zone the rate of increase of dyke volume fraction with depth is relatively great, probably greater than corresponding to curve 1 in Fig. 7. In the outer part of the subsidence zone a pile of gently dipping lava flows with a relatively small dyke volume fraction covers crustal sections which originally were in the dyke injection zone. This will also be the structure of the lithospheric plates. The resulting dyke volume fraction behaviour corresponds to the dotted curve in Fig. 7.

This structure appears to be compatible with the magnetic pattern observed over some mid-ocean ridges, e.g. the Reykjanes Ridge. The strong linear central anomaly would be caused partly by a relatively high intensity of dykes in the injection zone and partly by the normally magnetized lava flows accumulated in the zone of subsidence, which would have their greatest thickness at the axis. The outer anomalies on the other hand would be caused either by alternately normally and reversely magnetized dipping lava groups, or by the dykes which here are largely buried beneath the lava flows accumulated in the outer part of the subsidence zone. The buried dykes are probably more likely to be the sources of the linear anomalies than the gently dipping lava flows, as in this case a plot of anomaly age versus distance from axis would give a straight line through the origin. This is the case for the Reykjanes Ridge anomalies (Talwani et al., 1971). If the anomalies were caused by gently dipping lava flows this line would probably not go through the origin.

The burial could perhaps explain why the outer anomalies are weaker than the central anomaly, without assuming a higher than average magnetization for the axial rocks.

The relatively rapid increase in dyke volume fraction at a depth of the order of $q/2v_d$ may contribute at least partially to forming the seismic layer 3 (velocity about 6.5 km/sec). This would be in agreement with Gibson´s hypothesis (Gibson, 1966 ; Gibson and Piper, 1971) that layer 3 in Iceland is composed of intrusive rocks. There are, however, some difficulties in accepting this hypothesis on the basis of the model, as the model requires a relatively high dyke concentration at shallow depth in the dyke injection zone, which should manifest itself as a relatively shallow depth to layer 3 in this zone. Such a shallow depth appears not to be confirmed by the available seismic refraction data from the Icelandic volcanic zone (Pálmason, 1971). Furthermore direct field measurements of seismic velocities in large gabbroic intrusions in southeastern Iceland do not give velocities higher than about 5 km/sec, corresponding to layer 2 (Pálmason, 1963). Similar results have been obtained by direct measurements of seismic velocities in outcrops of the Troodos igneous complex of Cyprus which is believed to be a fragment of Mesozoic oceanic crust (Lort and Matthews, 1972).

Although the present model can account for the relatively sudden increase in dyke fraction postulated by Gibson and Piper (1971) there are important differences in the underlying reasoning. In the model it is assumed that all dykes reach the surface in the active zone of volcanism. This is a natural assumption from the plate tectonics point of view if the lithospheric plates are assumed to move as rigid bodies. The dykes are subsequently buried gradually beneath surface lava flows as they move laterally to join the lithospheric plate. Gibson and Piper (1971) on the other hand assume that "there is a preferred level to which the intruding magma tends to rise in the crust - a level perhaps governed by some form of hydrostatic equilibrium". This assumption is unnecessary if one assumes that the surface lavas flow over a considerably wider zone than the zone of dyke injection, an assumption borne out by observations in the present day volcanic zone of Iceland.

The detailed studies by Canadian groups of the Mid-Atlantic Ridge at 45°N provide an opportunity of comparing the model with a submarine part of the Mid-Atlantic Ridge. Aumento et al. (1971) have summarized some of the results obtained in this area. A particularly noteworthy result with respect to the present model is that the very young rocks that have been sampled appear to extend over an area considerably wider than the median valley. K-Ar and fission track datings, the thickness of manganese coatings on extrusive rocks, and the distribution of erratic rocks all indicate that the zone of very young rocks extends to about 50 km from the axis as defined by the median valley. This result can be interpreted to mean either a relatively great spreading rate during the Pleistocene, or a relatively wide zone of recent lava flows. The latter interpretation is in good agreement with the present model, which gives a zero age of surface rocks to a distance x_0 , and increasing age at greater distances as determined by the drift rate. If this interpretation is correct, it indicates that the distribution of volcanic activity in some submarine parts of the ridge may be quite similar to what is observed in Iceland.

The regional heat flow data in Iceland indicate that the bulk of the lava pile constituting the upper part of the crust in Iceland was generated in the Reykjanes-Langjökull zone, which is more or less the direct continuation of the present-day axial zone of the Reykjanes Ridge. This implies that the eastern rift zone is a relatively young feature. This conclusion was arrived at independently by Saemundsson (1972) on the basis of stratigraphic and paleomagnetic correlations in NE-Iceland. He estimates the age of the eastern rift zone at 4 myrs. Saemundsson suggests that the shifting of the rift zone was the result of movement of a crustal plate as a whole over a stationary mantle hot-spot plume, at present assumed located beneath the eastern rift zone.

An alternative suggestion is offered here to explain the shifting. The varying glacial load on the crust during the Pleistocene may have been at least a trigger to initiate a new zone of weakness along which crustal accretion could take place. It is noteworthy that the eastern zone appears to be confined more or less to the present land area of Iceland, which has been affected by glaciation to a varying degree. The beginning of the Pleistocene is estimated to be about 3 m.y. ago (McDougall and Wensink, 1966 ; Th. Einarsson et al., 1967), which is compatible with the estimated age of the eastern zone. This suggestion for the origin of the eastern zone of course needs to be critically evaluated, but it is offered here as a tentative possibility.

The axial valley observed on many mid-ocean ridge segments, has been discussed by several authors (e.g. Atwater and Mudie, 1968 ; van Andel and Bowin, 1968 ; Deffeyes, 1970 ; Osmaston, 1971). This does not appear to be a general feature of the mid-ocean ridge crests as for instance evidenced by the Reykjanes Ridge (Talwani et al., 1971) and some ridge segments in the Norwegian-Greenland Sea (Johnson, 1972). No attempt has been made to explain it by the present model. Its existence is probably related to isostatic equilibrium in the main dyke injection zone. It is of some interest to note that models of the crestal zone of mid-ocean ridges, e.g. by the above authors, usually show dips away from the axis as defined by the median valley. This is contrary to what is observed in Iceland where the regional dips of lava flows adjacent to the volcanic zones are usually towards the zones.

The convergence hypothesis of Keith (1972) for explaining the mid-ocean ridges is contradicted by the regional heat flow data from Iceland, as well as the abundant geothermal activity in the axial zone, which require a very high heat flow, not accounted for in the convergence model.

The present model may lead to interesting consequences regarding the source of some of the lavas erupted. Under steady-state conditions the isotherms in the crust in the volcanic rift zone remain fixed with respect to the surface level. The subsidence thus leads to a certain remelting in the lower part of the crust. This can lead to a range of magmas depending on conditions in the lower crust, especially with regard to water content.

9. Conclusions.

The kinematic model, which is based on plate tectonics concepts, appears to be compatible with observations of regional dips of flood basalts as well as dyke volume fraction in the eastern Iceland basalt pile. It is also compatible with dip observations in other parts of Iceland, if the assumption is made that rifting and volcanism has shifted between two or three zones during the time involved in generating the Icelandic lava pile.

The heat flow in the volcanic rift zone as calculated from the model is compatible with observed heat discharge by conduction and geothermal activity. The observed regional heat flow pattern indicates that the main bulk of the Icelandic lava pile was accreted in the Reykjanes-Langjökull volcanic rift zone and its extinct continuation to the north. This implies that the NE-Iceland volcanic rift zone as well as the eastern zone in southern Iceland are relatively young. This is in agreement with Saemundsson´s (1973) conclusions which were based on stratigraphic and paleomagnetic correlations in northeastern Iceland.

Estimates of average heat transport per unit length of the volcanic zone in Iceland, based on observations and model calculations, give the following. Volcanic activity accounts for 21 MW/km, heat conduction for 21 MW/km and geothermal activity for 14 MW/km. The excess heat transported laterally from the volcanic zone to the lithospheric plates above a depth of 50 km is estimated at 78 MW/km, giving a total average heat discharge per unit length of the Icelandic volcanic zone of 134 MW/km.

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Fig. 9. Surface temperature gradient in the volcanic zone of the model, as a function of the half-width x_0.

Fig. 10. Diagram showing the decay of the crustal temperature field in the lithospheric plate. The initial temperature is given by eqn. 9a with g_0 = 180°C/km. The straight dashed line is assumed to represent the solidus.

Fig. 11. Regional temperature gradients (°C/km) from Iceland. The schematic boundaries of the Reykjanes - Langjökull - Skagi zone are used to estimate distances of the heat flow stations, as used in Fig. 13, from this zone. The arrows show the assumed directions of plate motion. The line A-B indicates the location of the heat flow profile in Fig. 12.

Fig. 12. Heat flow values from southwestern Iceland projected on the line A-B in Fig. 11.

Fig. 13. Heat flow values from Iceland vs distance from the Reykjanes-Langjökull-Skagi zone (cf. Fig. 11). Corrections have been applied to take into account that the northern part of this zone is assumed to have been inactive for the last 3 m.y. The solid curves are theoretical cooling curves for the lithospheric plates for initial surface gradients of 100-250 °C/km. The drift velocity is assumed to be 1 cm/yr.

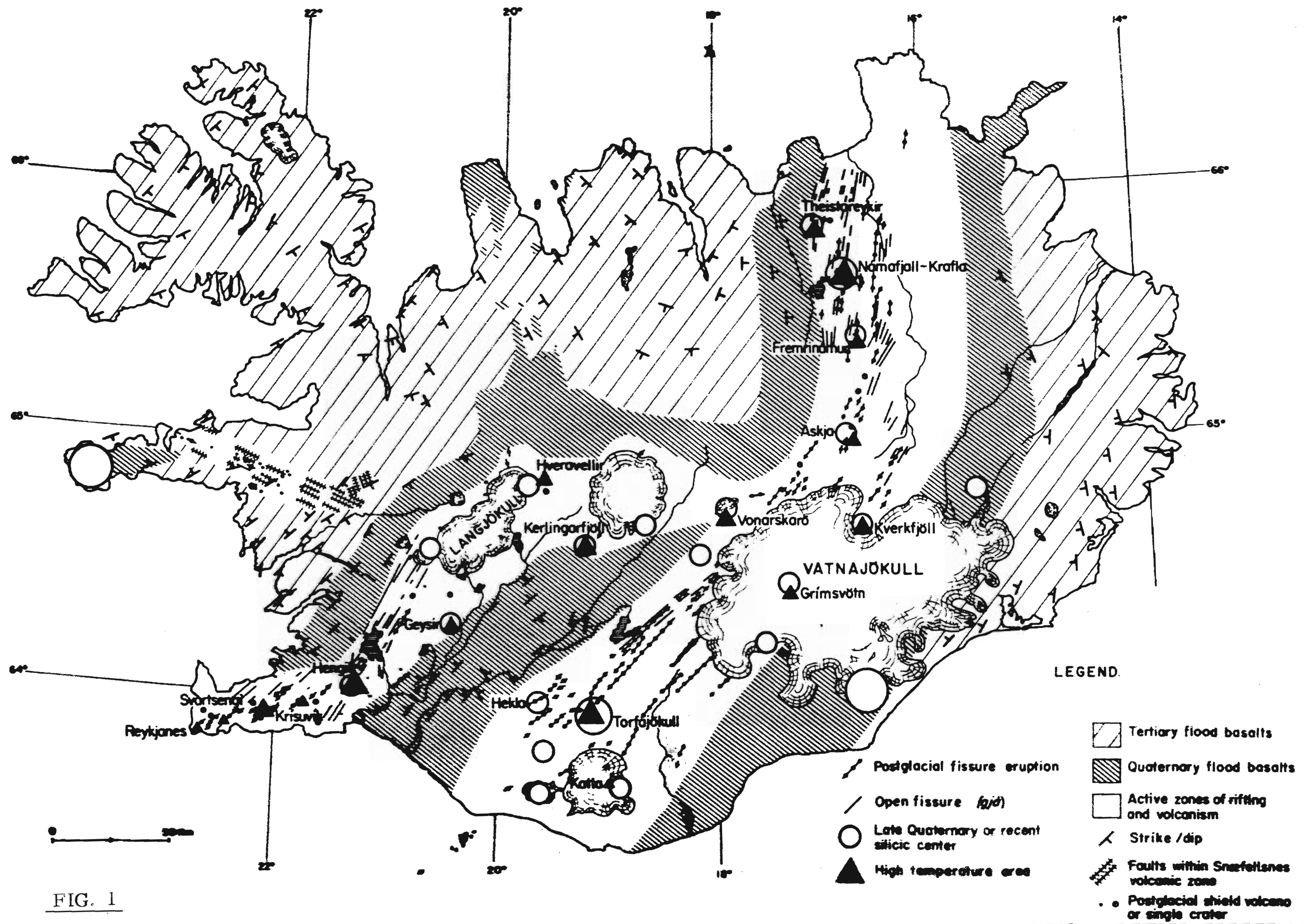

22°
20°
18°
16°
14°
66°
65°
64°
Theistareykir
Námafjall-Krafla
Fremrinámur
Askja
Hveravellir
Kerlingarfjöll
LANGJÖKULL
Vonarskard
Kverkfjöll
VATNAJÖKULL
Grímsvötn
Geysir
Hengill
Svartsengi
Krísuvík
Reykjanes
Hekla
Torfajökull
Katla
LEGEND
Tertiary flood basalts
Quaternary flood basalts
Active zones of rifting and volcanism
Strike /dip
Faults within Snæfellsnes volcanic zone
Postglacial shield volcano or single crater
Postglacial fissure eruption
Open fissure (gjá)
Late Quaternary or recent silicic center
High temperature area

FIG. 1

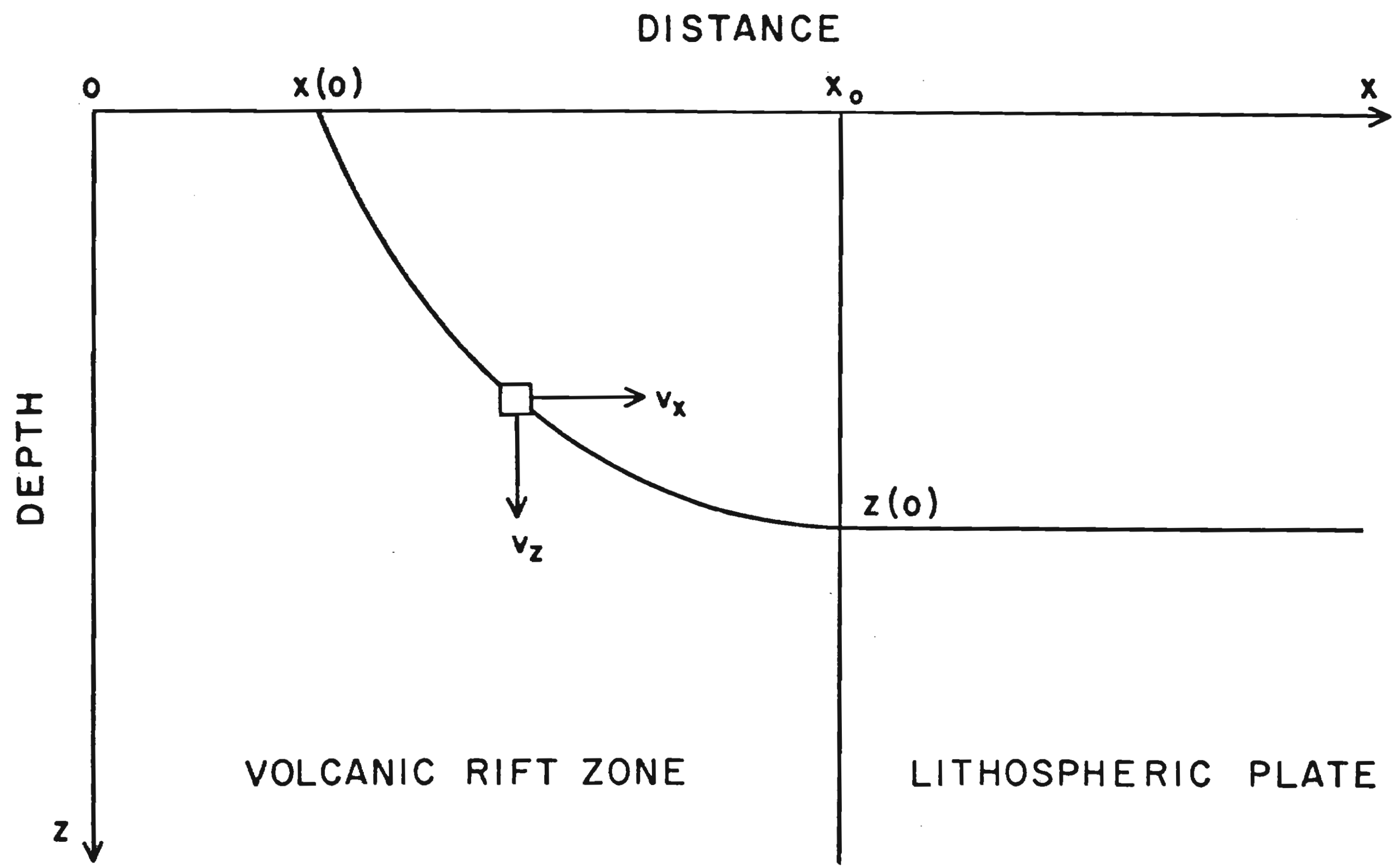

DISTANCE
0
x(o)
x_o
x
DEPTH
v_x
v_z
z(o)
VOLCANIC RIFT ZONE
LITHOSPHERIC PLATE
z

FIG. 2

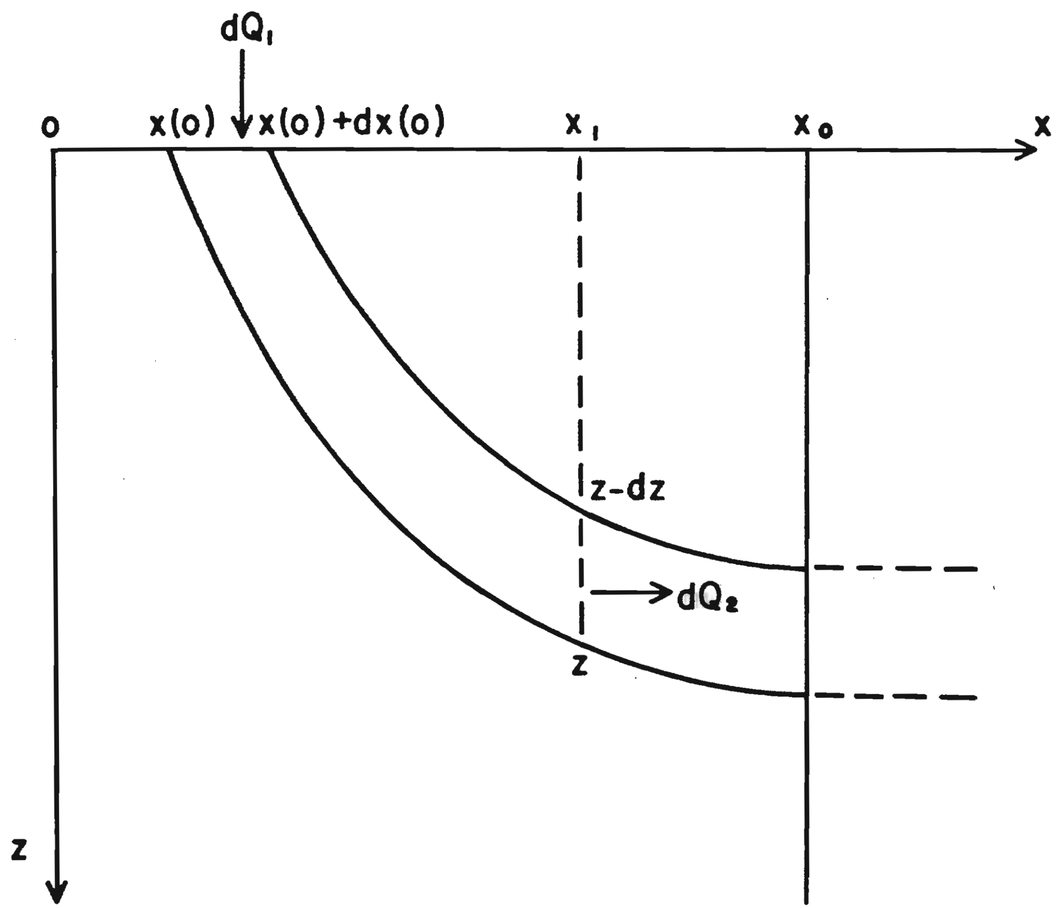
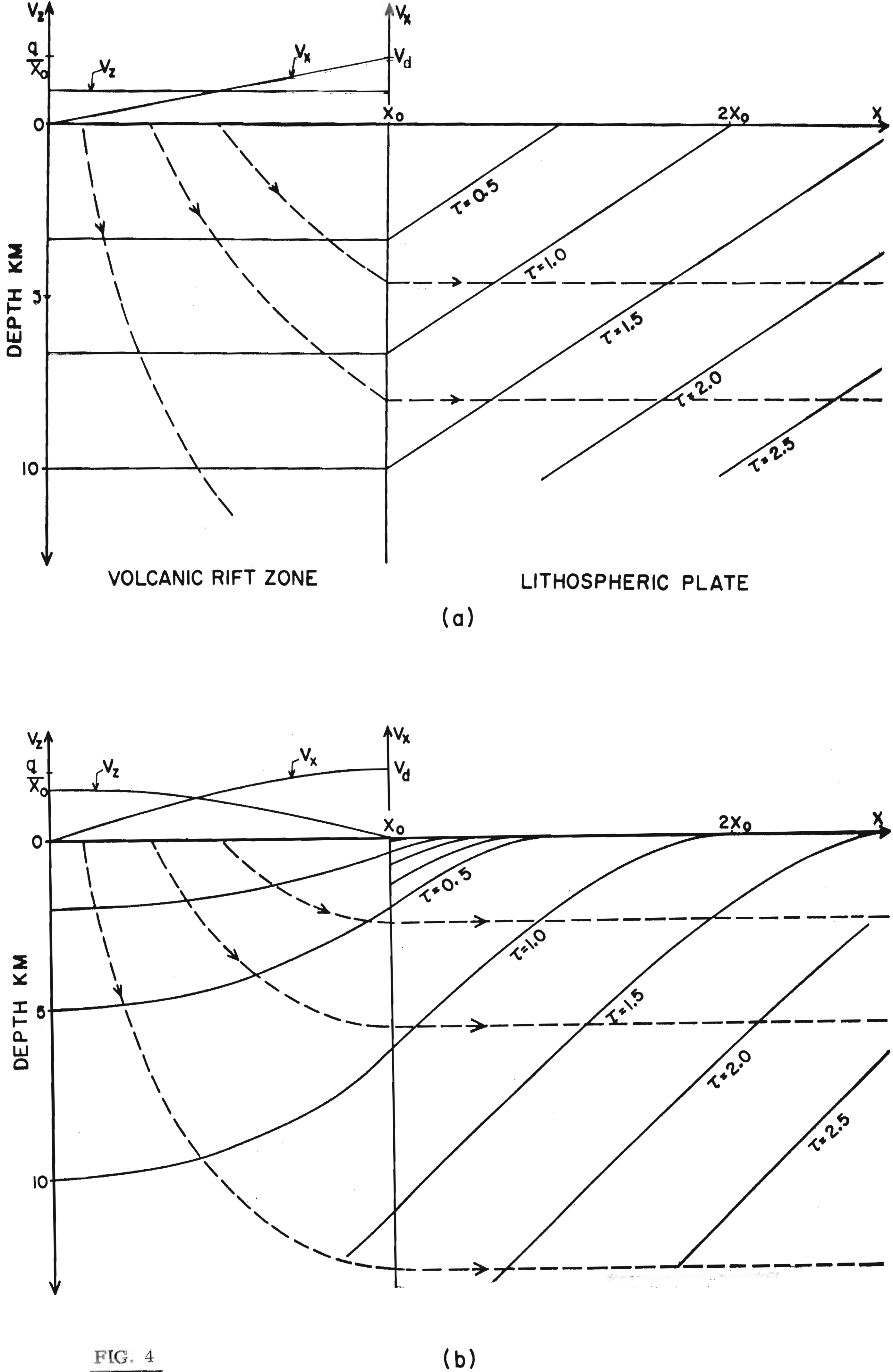


FIG. 3



V_z
V_x
$\frac{q}{X_0}$
V_d
X_0
$2X_0$
X
0
5
10
DEPTH KM
$\tau = 0.5$
$\tau = 1.0$
$\tau = 1.5$
$\tau = 2.0$
$\tau = 2.5$
VOLCANIC RIFT ZONE
LITHOSPHERIC PLATE
(a)
(b)

FIG. 4

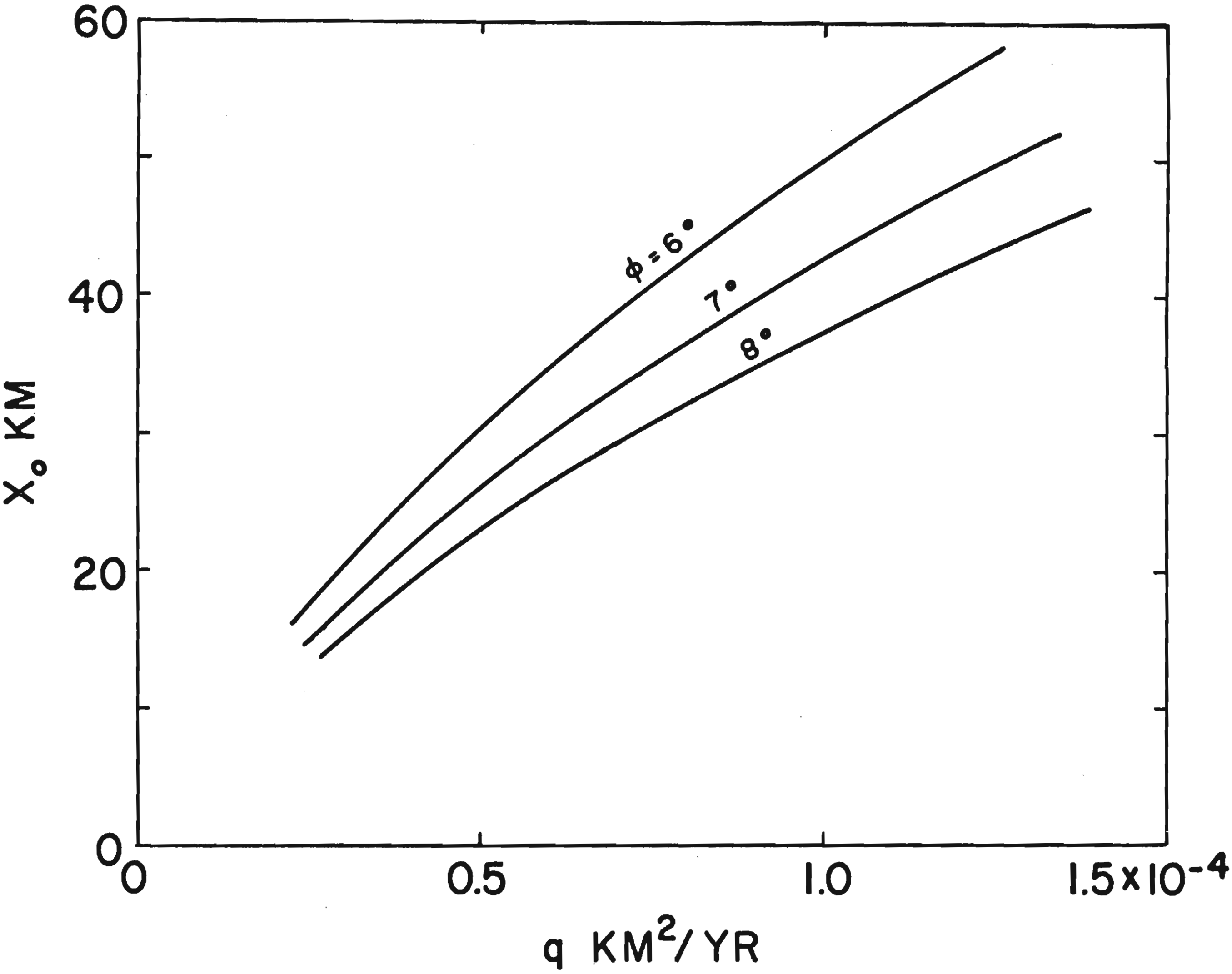


FIG. 5

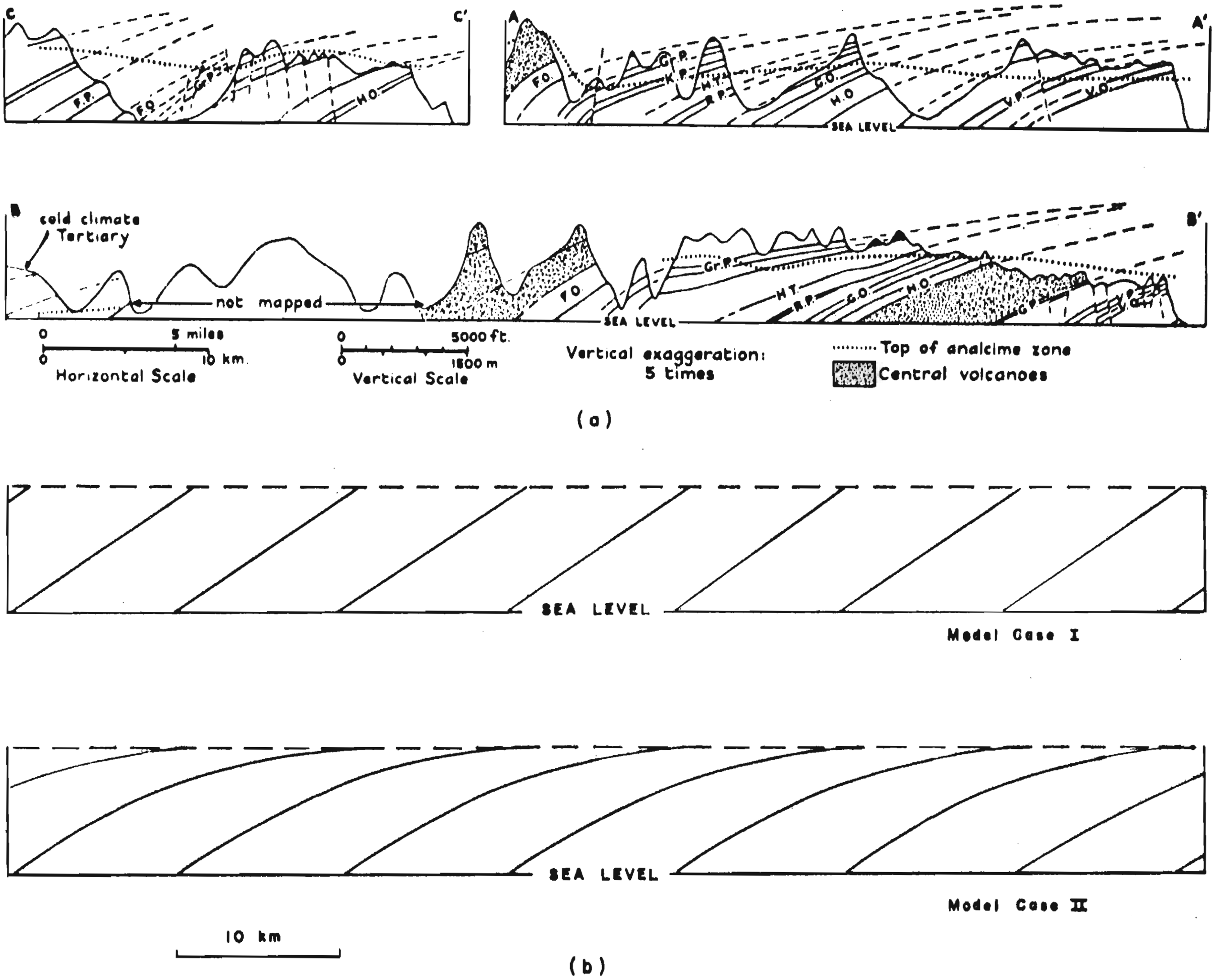

C
C'
A
A'
B
B'
cold climate Tertiary
not mapped
SEA LEVEL
0 5 miles
0 10 km.
Horizontal Scale
0 5000 ft.
0 1500 m
Vertical Scale
Vertical exaggeration: 5 times
Top of analcime zone
Central volcanoes
(a)
1500 m
SEA LEVEL
Model Case I
1500 m
SEA LEVEL
Model Case II
10 km
(b)

FIG. 6

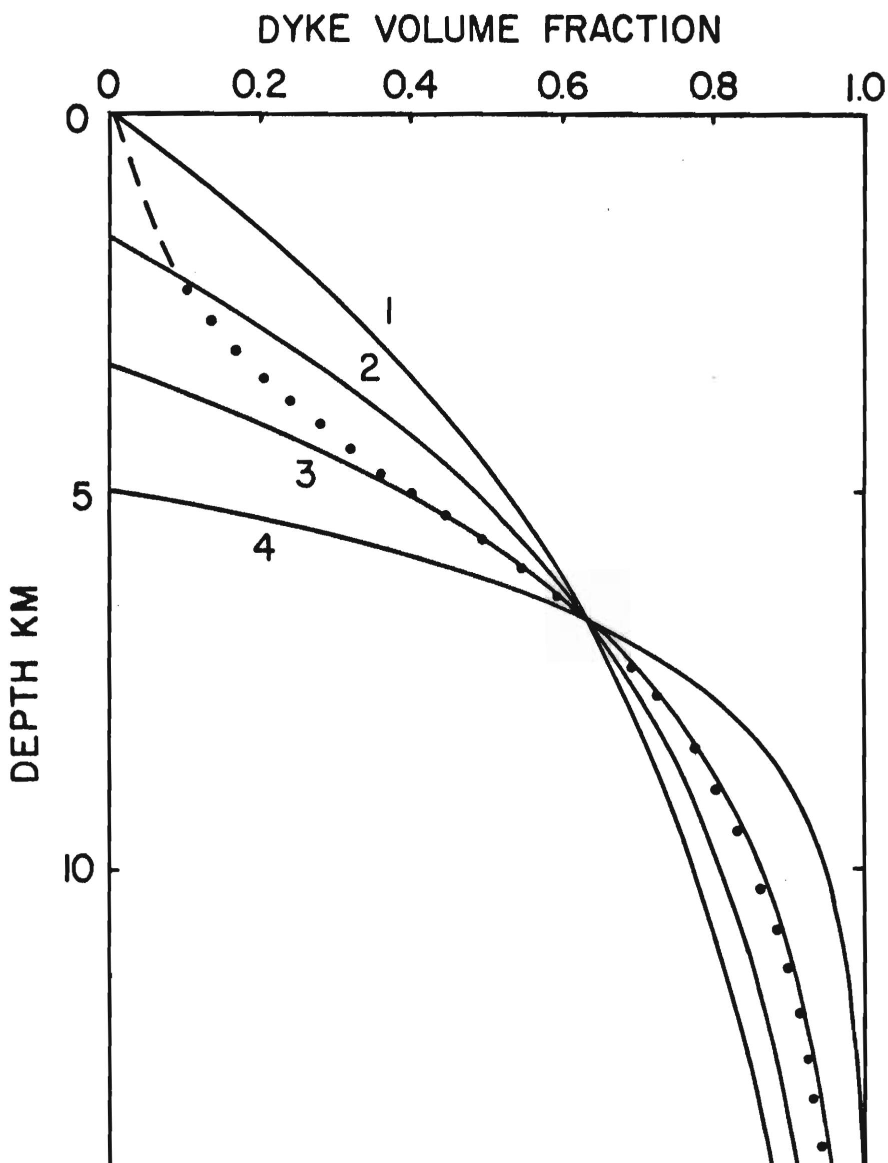
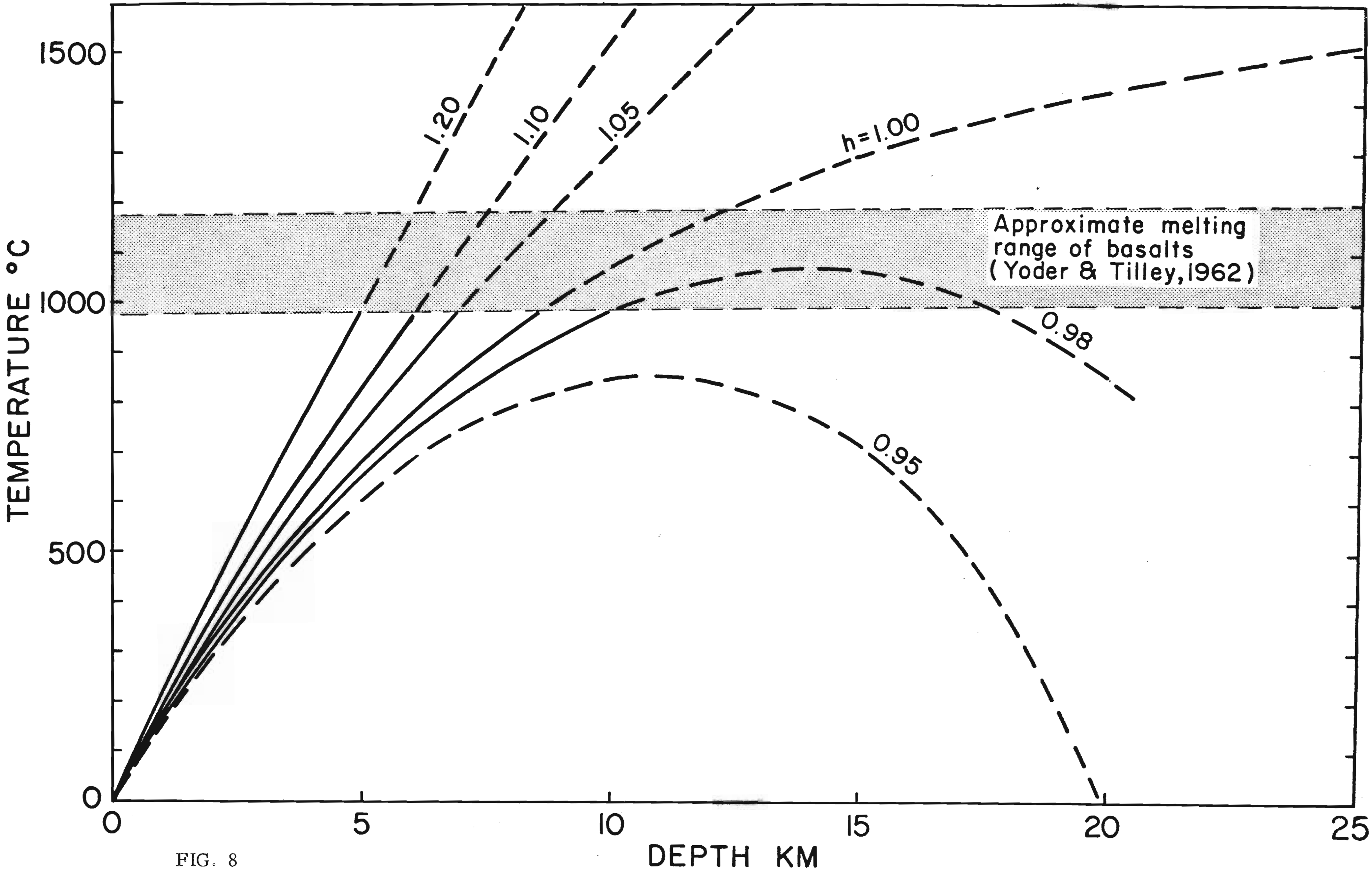


FIG. 7



1500
1000
500
0
TEMPERATURE °C
0
5
10
15
20
25
DEPTH KM
1.20
1.10
1.05
h=1.00
0.98
0.95
Approximate melting
range of basalts
(Yoder & Tilley, 1962)

FIG. 8

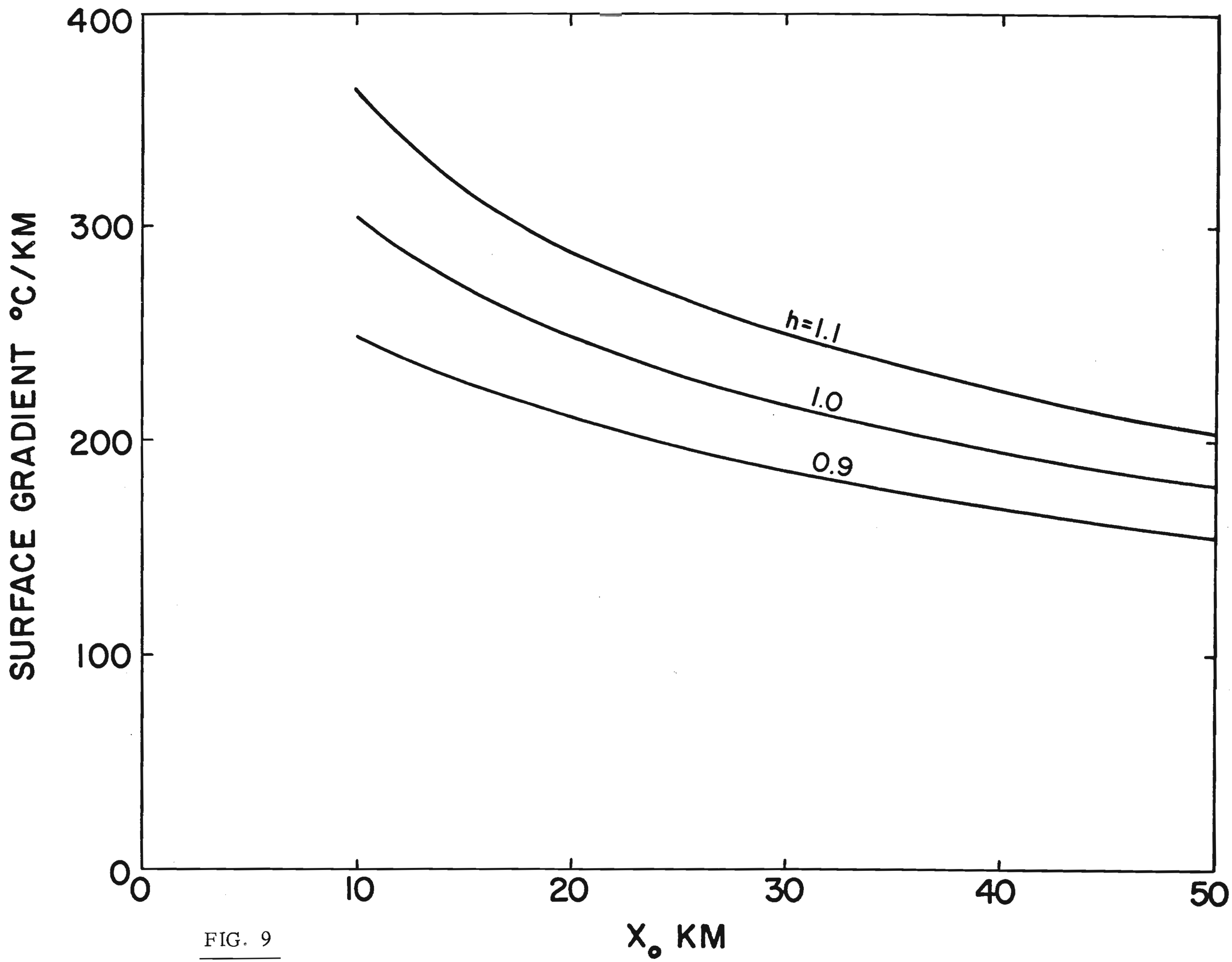

SURFACE GRADIENT °C/KM
400
300
200
100
0
0
10
20
30
40
50
X₀ KM
h=1.1
1.0
0.9

FIG. 9

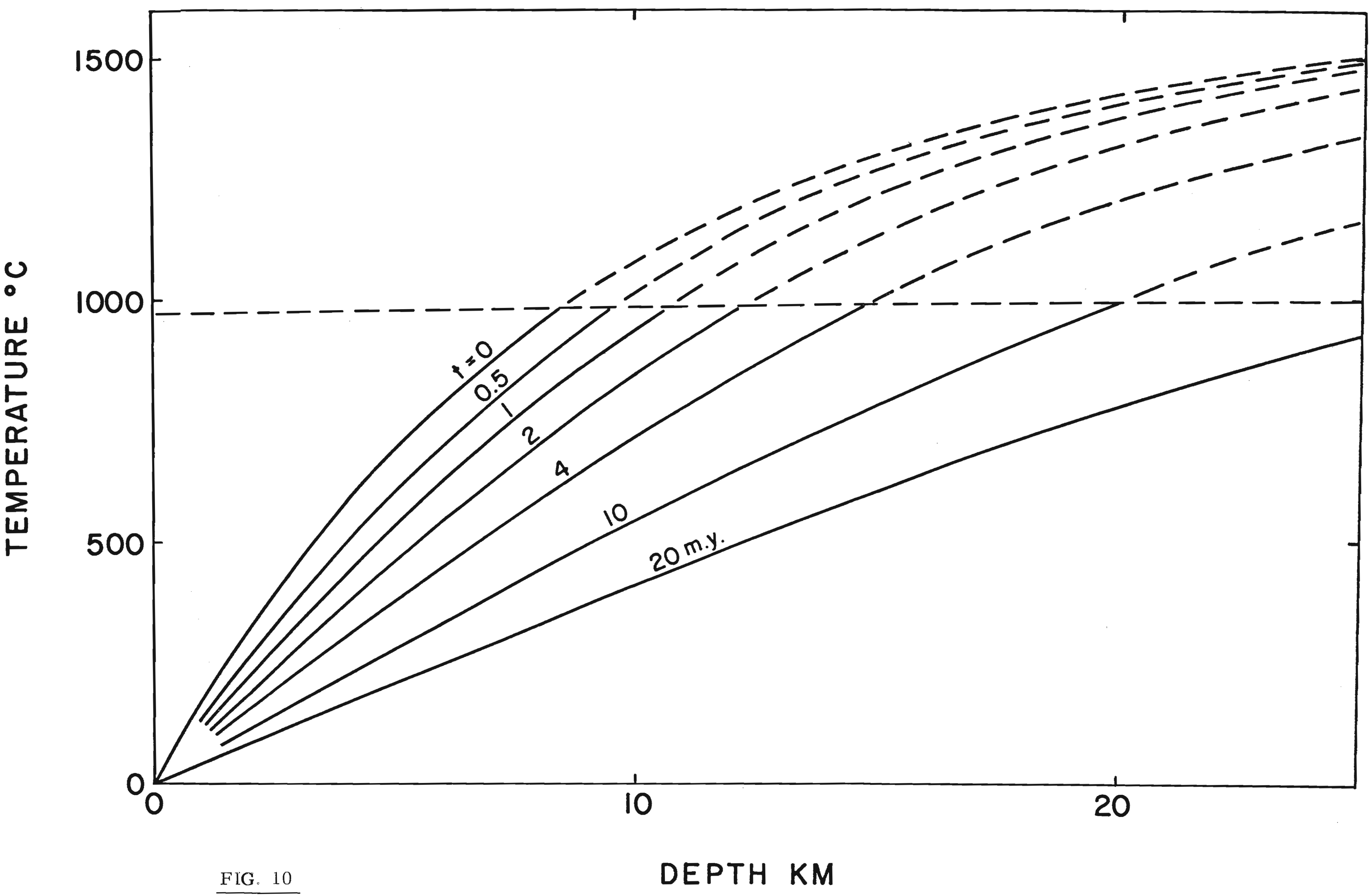
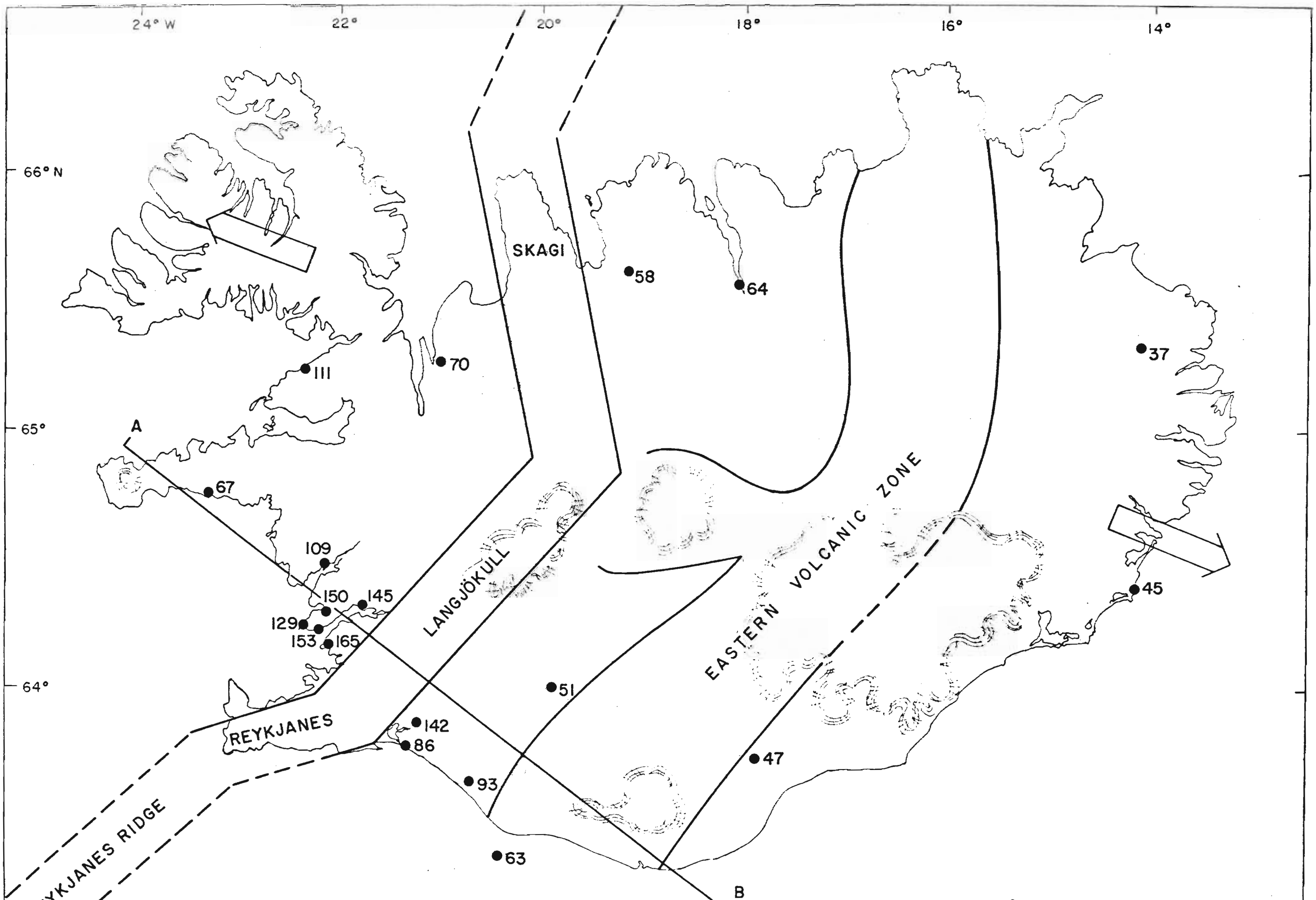
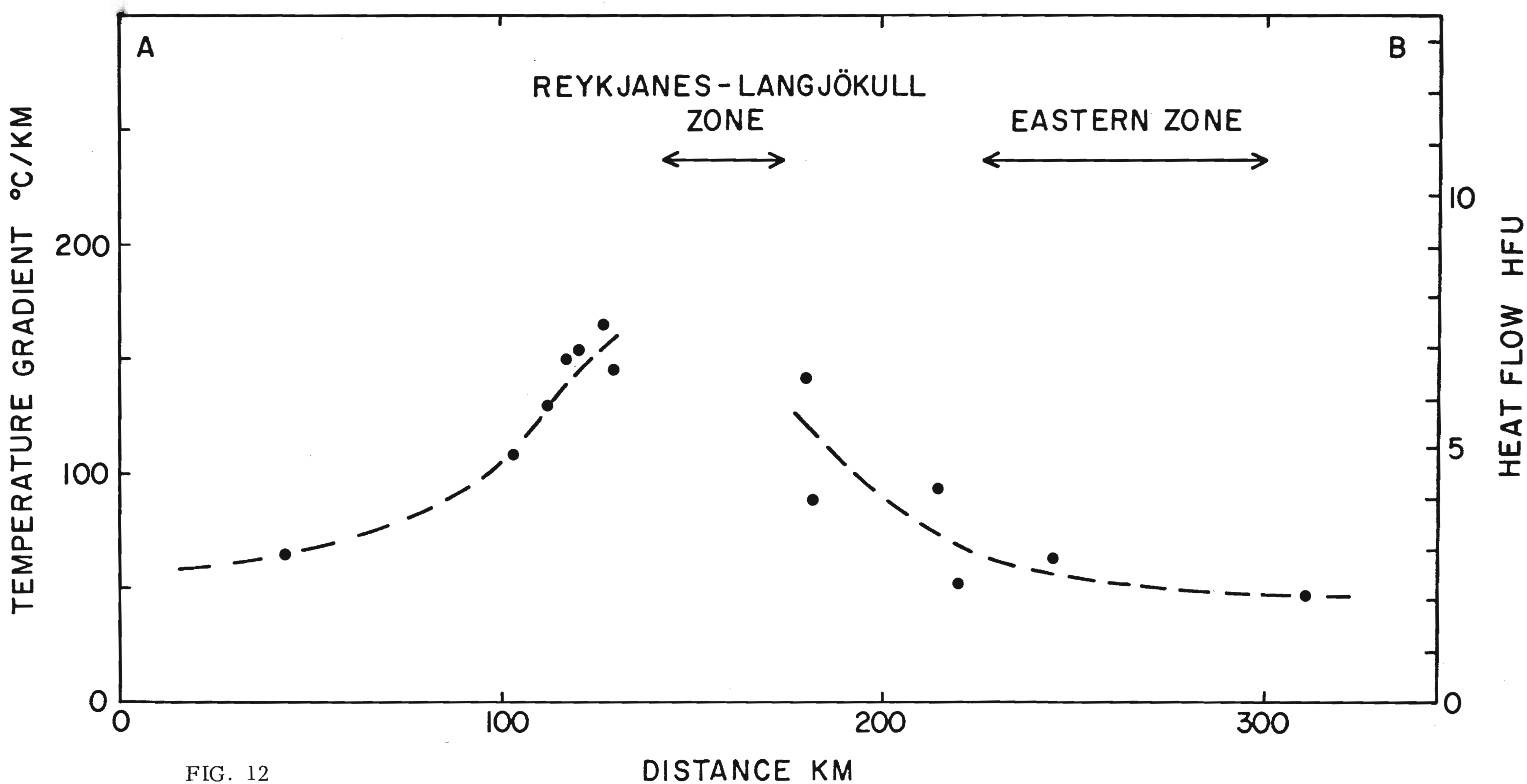


FIG. 10



24° W
22°
20°
18°
16°
14°
66° N
65°
64°
SKAGI
LANGJÖKULL
EASTERN VOLCANIC ZONE
REYKJANES
REYKJANES RIDGE
A
B
58
64
37
70
111
67
109
150
145
129
153
165
51
142
86
93
63
47
45
0
50
100km

FIG. 11



A
B
REYKJANES-LANGJÖKULL
ZONE
EASTERN ZONE
TEMPERATURE GRADIENT °C/KM
HEAT FLOW HFU
DISTANCE KM
0
100
200
300
5
10

FIG. 12

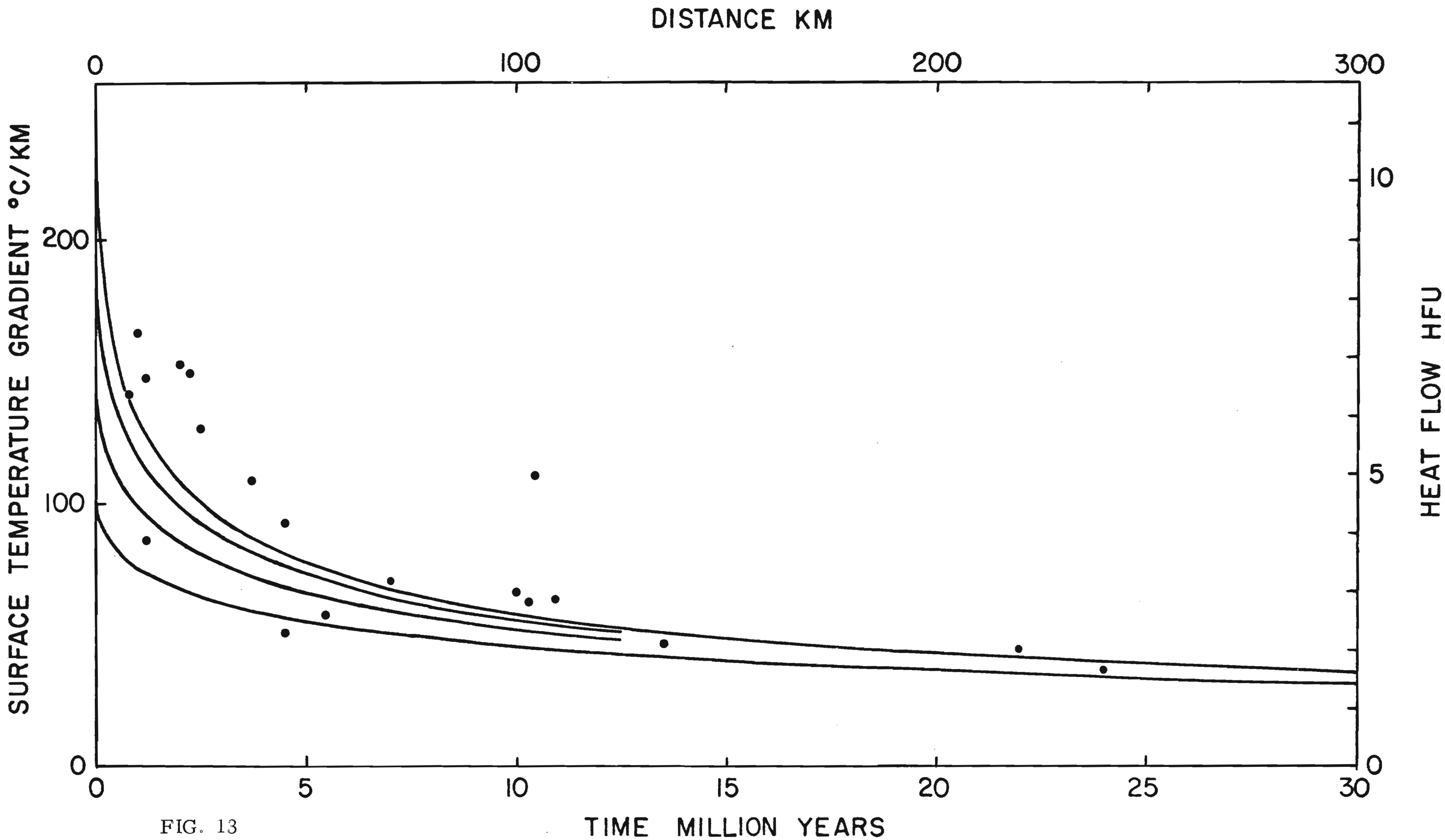

DISTANCE KM
0
100
200
300
SURFACE TEMPERATURE GRADIENT °C/KM
0
100
200
HEAT FLOW HFU
0
5
10
0
5
10
15
20
25
30
TIME MILLION YEARS

FIG. 13