KBS TEKNISK RAPPORT

Groundwater movements around a repository

Repository domain groundwater flow analyses

Part 1 Permeability perturbations
Part 2 Inflow to repository
Part 3 Thermally induced flow

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Edward L Skiba
Robin Carlwood

Hagconsult AB september 1977

GROUNDWATER MOVEMENTS AROUND A REPOSITORY

REPOSITORY DOMAIN GROUNDWATER FLOW ANALYSES
PART 1 PERMEABILITY PERTURBATIONS
PART 2 INFLOW TO REPOSITORY
PART 3 THERMALLY INDUCED FLOW

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TECHNICHAL REPORT 5

REPOSITORY DOMAIN GROUNDWATER FLOW ANALYSES

PART I

PERMEABILITY PERTURBATIONS

GROUNDWATER MOVEMENTS AROUND A REPOSITORY

Technical report 5. Repository Domain Groundwater Flow Analyses
Part I:
Permeability Perturbations

FOREWORD

This report was prepared as one of a series of Technical reports
within a study of the groundwater movements around a repository for
radioactive waste in the precambrian bedrock of Sweden. It was written
in three parts, (I) Permeability Perturbations, (II) Inflow to Repository
and (III) Thermally Induced Flow. This is Part I. The contract for this
study was between KBS - Kärnbränslesäkerhet (Project Fuel Safety) and
Hagconsult AB of Stockholm, Sweden. RE/SPEC Inc. of Rapid City, SD/USA
and Acres Consulting Services Ltd of Niagara Falls, Ontario/Canada
acted as subconsultants to Hagconsult AB.

The principal author of this report is Mr. Joe L. Ratigan of RE/SPEC
Inc. with contributions by Dr. A.S. Burgess of Acres. Review was provided
by Dr. Ulf E. Lindblom of Hagconsult AB, Dr. Paul F. Gnirk of RE/SPEC Inc.
and Dr. Robin G. Charlwood of Acres.

The opinions and conclusions in this document are those of the author and
should not be interpreted as necessarily representing the official policies
on recommendations of KBS.

Stockholm September 1977

Ulf E. Linblom
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REPOSITORY DOMAIN GROUNDWATER FLOW ANALYSES - PERMEABILITY PERTURBATIONS

1. INTRODUCTION

Previous analyses within this study (1)[x] have provided qualitative
evaluations of the perturbation of in situ hydraulic permeability and
groundwater flow resulting from (a) the transient temperature fields
due to the emplacement of radioactive heat generating waste
in a granitic formation (2), and from (b) the stress perturbations
due to the excavation of the repository and the thermomechanical
loading associated with the aforementioned temperature fields (3). In
this report, these perturbations will be treated and analyzed in a
quantitative fashion. In particular, relationships have been derived
and extracted from the literature which relate hydraulic permeability to
rock stress, and viscosity to groundwater temperature. These relation-
ships have been employed to quantitatively evaluate the hydraulic per-
meability around the repository as a function of the temperature fields
presented in (2) and the stress states presented in (3).

The results presented in this report are based on limited data. At this
time, no site specific data are available for these effects. To determine
these effects, extensive field and laboratory testing and advancement
of the state-of-the-art in both field testing methodology and under-
standing of joint behavior will be required. However, the quantitative
results presented in this report can be utilized to quite reliably
identify the importance and influence of the temperature and stress
perturbations on the rock mass hydraulic permeabilities. In fact, the
perturbations of the hydraulic permeability can probably be established
with a greater degree of certainty than the values of in situ permeability.

Future sections of this report will present the assumed permeability -
stress and permeability - temperature relationships employed and the
subsequent perturbations resulting in the local and global repository
regions.

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[x]Numbers in parenthesis refer to references at end of text.

2. PERMEABILITY AS A FUNCTION OF STRESS AND TEMPERATURE

2.1 Stress dependency

The relationships which correlate hydraulic permeability with rock stress as presented in this report should not be interpreted as nessessarily accurately representing the in situ conditions. This topic is the subject of current on going research and will perhaps not be well established within the next decade. Previous investigations are inconclusive in relating permeability changes to pre-loading stress states, and in relating permeability to load path in a quantitative manner. A more detailed description of the present state of the art is provided in (4).

Despite the uncertainty in the permeability-stress relationships, one can conclude with a relatively high degree of certainty that within the elastic range of loading and unloading of a rock joint, the hydraulic permeability is a function of the rock stress normal to the joint plane (or equivilantly to the joint aperture). It is this relationship (permeability as a function of normal stress) that has been utilized in this study to analyze the effects of stress changes on hydraulic permeabilities.

The empirical relationship between permeability and normal stress which has been utilized in obtaining the results presented in this report has been previously cited in (4). In polynomial form, this relationship can be stated as

$$\log_{10} |K(\sigma_n/\sigma_o)| = -5.57 + 0.362 \log_{10} (\sigma_n/\sigma_o) - 0.978 |\log_{10} (\sigma_n/\sigma_o)|^2$$
$$+0.167 |\log_{10} (\sigma_n/\sigma_o)|^3$$

where

σ_n = stress normal to the joint plane (MPa)

K = hydraulic permeability (m/s)

σ_o = ρ_ℓ g $\cdot$ ρ $\cdot$ 1 (MPa)

g = gravational acceleration

ρ = density (t/m^3)

ℓ = unity (m)

Although other relationships have been obtained in the laboratory and in the field (5), this relationship has been employed since it produces higher hydraulic permeabilities and possesses less normal compressive stress dependency than those presented in (5) for the range of stresses which will occur in and around a repository. In this regard, the formula is conservative, since it results in higher groundwater flows.

2.2 Temperature dependency

A temperature dependency of hydraulic permeability occurs through the perturbation of groundwater viscosity. The differential equation for transient groundwater flow in an isothermal porous media is:

$$\nabla \left| \rho \frac{k_{ij}}{\mu} \nabla p \right| + Q(t, x_i) = \rho c \frac{\partial p}{\partial t}$$

where

ρ = density

k_{ij} = intrinsic permeability tensor

μ = dynamic viscosity

c = specific storage

p = potential

$Q(t, x_i)$ = inhomogenieties (fluid source or sink)

reduces to

$$\nabla^2 p = 0$$

in the steady state situation with no inhomogenieties· The flux or groundwater flow is merely

$$q_i = K_{ij} \frac{\partial p}{\partial x_j}$$

where

K_{ij} = hydraulic permeability

However, if the steady state groundwater flow is treated in a "quasi-coupled" fashion, the permeability tensor can be estimated to be

$$\frac{K_{ij}}{\mu(T)} \cdot \mu(T_o)$$

where

$\mu(T)$ = dynamic viscosity at temperature T

T_o = ambient temperature

in order to obtain a first estimate of thermal effects on hydraulic per-
meability and subsequent groundwater flow. This is the form of the
hydraulic permeability - temperature relationship utilized in this study
to obtain the results presented in this report.

The function $\mu(T)$ has been taken from (6) and can be stated as

$$\frac{1}{\mu(T)} = 5380 + 3800A - 260A^3 \qquad (kg/m,s)$$

where

$A = (T -150)/100$

$T = $ temperature (^{o}C)

As stated in (6), this relationship is accurate within 3% over a range
of temperatures of 0 ^{o}C to 300 ^{o}C.

3. LOCAL MODELING PERTURBATIONS

3.1 Excavation Effects

The excavation of an underground opening will result in a perturbation of the hydraulic permeability around the opening. The excavation method, the subsequent accentuation and attenuation of stresses around the opening, the stress relief and dilatation resulting from failure in shear, all affect hydraulic permeability. Due to the complexity of this interaction, if is extremely difficult (if not impossible) to provide a quantitative analysis of the resulting permeability based upon intuition. Perhaps the same statement can be made even for a qualitative analysis.

Figure 1 and 2 illustrate the ratios R_h and R_v, respectively, around the repository storage tunnel after excavation as

$$R_h = K_h(\sigma_e)/K_h(\sigma_0)$$
$$R_v = K_v(\sigma_e)/K_v(\sigma_0)$$

where

$K_h(\sigma_e)$ = horizontal permeability after excavation

$K_v(\sigma_e)$ = vertical permeability after excavation

$K_h(\sigma_0)$ = in situ horizontal permeability

$K_v(\sigma_0)$ = in situ vertical permeability

σ_0 = in situ stress state

σ_e = stress state after excavation

The in situ permeabilities are those identified in (7) as Case 4

The stresses used in producing these figures are those presented in (3) for the excavation in a rock mass with 0° and 90° joint sets. Although this simulated excavation resulted in some regions around the opening experiencing strength failure, no explicit accounting of these regions has been utilized in assessing the resulting permeabilities. The stress dependency equation presented in part 2.1 was used in all regions.

The effect of the excavation on the virgin permeabilities is quite pronounced around the storage tunnel periphery; however, the permeability is at the initial magnitude for distances greater than about 2 meters from the tunnel periphery. This dimension would probably be greater if the regions experiencing failure were considered to have experienced joint dilatation.

It is essential to note that the permeabilities presented in the pre-
ceeding figures are highly dependent upon the assumed in situ stresses
and the storage tunnel geometry. For example, if the rib were nearly
vertical, the horizontal stresses in the floor would be less and the
resulting vertical permeability would be greater. Similarly, if the
ratio of in situ horizontal stress to vertical stress were taken to be
1 rather than 2, the horizontal stress at the roof would be less and
the resulting vertical permeability greater.

Figures 3 and 4 present the permeability ratios resulting from tunnel ex-
cavation based on the assumption that the principal permeability directions
and magnitudes are coincident with, and controlled by, the principal stresses.
In fact, this may not be the case but it is instructive in displaying the
permeability changes normal and tangential to the repository storage tunnel.
As can be seen in the figures, the permeability ratio normal to the tunnel
periphery $(K_h(\sigma_e)/K_h(\sigma_0)$ or that component which dominates flow into the
tunnel, is very low in comparison to the tangential ratio. Physically, this
can be explained by noting that the excavation results in an increased tan-
gential compressive stress (above the in situ magnitude) around the tunnel
which results in a lower permeability than the initial magnitude. The normal
stresses are zero at the tunnel periphery and rapidly increase to in situ
magnitudes, resulting in the high tangential permeability near the tunnel and
near in situ permeabilities at about 2 m from the tunnel periphery.

The stresses resulting from the assumption of joint sets at 0° and 90°
and the stresses from the 45° and -45° joint sets yield very similar per-
meabilities. These permeability ratio fields $R_h = K_h(\sigma_e)/k_h(\sigma_0)$ and $R_v = K_v(\sigma_e)/K_v(\sigma_o)$ are displayed in contour plot form in Figures 5 and 6,
respectively.

3.2 Thermomechanical stress effects

In this section of the report permeabilities resulting from the thermally
induced stresses presented in (3) will be presented for regions around the
storage tunnel.   These permeabilities have been calculated assuming per-
meability to be a function of the total stress normal to the joint plane.
Permeability changes due to viscosity changes resulting from inhomogeneous
temperature fields are presented in a later section of this report.

3.3 Temperature effects

In this section of the report permeability changes are evaluated with
the functional relationships described in section 2.1 and 2.2 including
both normal stress and viscosity effects due to temperature variations.
As was the case with the thermomechanical stresses, the temperature
effect on the permeability is not nearly as significant as the perturba-
tion of the initial magnitudes caused by excavation. The variation due
to temperature perturbation is much smaller than the uncertainty in the
predictions of the in situ permeabilities.

Since a ratio of permeabilities with viscosity changes to permeabilities
without viscosity changes reduces merely to the function $\mu(T)$ presented
in section 2.2, Figures 12 and 13 were constructed with the ratio

$$R = \frac{\mu(T_o)}{\mu(T)}$$

where T_o = pre-emplacement geothermal temperature

These two figures represent the viscosity ratios at 40 and 100 years after
emplacement without ventilation. The greatest magnitudes are in the vicinity
of the waste canister and reach a value of about 2.5 after 40 years.

In an effort to quantify the effect that repository ventilation has upon
post-emplacement hydraulic permeabilities, the ratios R_h and R_v were cal-
culated as

$$R_h = K_h(\sigma_e + \sigma_{Tv}, T_v)/K_h(\sigma_e + \sigma_{Tnv}, T_{nv})$$

$$R_v = K_v(\sigma_e + \sigma_{Tv}, T_v)/K_v(\sigma_e + \sigma_{Tnv}, T_{nv})$$

where T_v = temperatures resulting with 30 years of ventilation

T_{nv} = temperatures resulting without ventilation

The ratios, R_h and R_v are displayed in Figures 14 and 15, respectively,
which result from the total stresses and temperatures occurring 40 years
after emplacement. Near the tunnel periphery the tangential permeability
is greater with ventilation than without, whereas the normal permeability
is about 5-20% less with ventilation than without.

4. GLOBAL MODELING PERTURBATIONS

The evaluation of the global permeability perturbations caused by the
emplacement of radioactive waste has proceeded in the same manner as
that for the local perturbations except that the excavation effects have
not been included. The scale of the global model does not allow this level
of detail.

The permeability perturbations resulting from the thermomechanical stress
and temperature have been evaluated at various depths and consider
various waste emplacement sequences.

4.1 Thermomechanical stress effects

The normal stresses utilized in this section of the report for the
evaluation of permeability include both the in situ stresses and the
thermally induced stresses. Since minimal rock strength failure was
exhibited in the global rock mechanics analysis (3), the results presented
herein are for all practical purposes independent of the choice of joint
orientation. Therefore, the stresses from the 0° and 90° simulations were
be employed.

The ratios of the horizontal and vertical permeability in the repository
domain occurring 100 years after waste emplacement to the respective in
situ values are presented in Figure 16, as a function of depth. The thermo-
mechanical stresses utilized in obtaining these results are those re-
lating to the instantaneous emplacement at 500 m depth (3). As can be
seen in the figures, the stress effect on permeabilities is very small.
Similar small changes in permeability are noted in both the linear waste
emplacement model and the instantaneous waste emplacement at 1000 meters.

4.2 Temperature effects

The permeability changes due to the effect of temperature on viscosity
have been analyzed with the temperatures resulting from both the instan-
taneous and linear waste emplacement sequences. As noted earlier, this
perturbation can be represented as the ratio of viscosities. This ratio
as a function of depth is displayed in Figure 17 for the instan-
taneous waste emplacement at 40 and 100 years after emplacement. The viscosity

perturbation in the linear waste emplacement model is illustrated in Figure 18 for 40 and 100 years after initiation of emplacement. The maximum viscosity ratio for these cases is approximately 2.

Perturbations due to viscosity changes at the 1000 meter level repository model are slightly less than those at the 500 meter level model due to the higher initial temperature at 1000 meters.

5. SUMMARY AND CONCLUSIONS

The perturbations of the in situ hydraulic permeability caused by (a)
stresses resulting from repository storage tunnel excavation and (b)
the thermomechanical stresses resulting from the radiogenic
heat have been evaluated. Changes in the permeability due to the tempera-
ture dependence of the viscosity have also been studied.

The effects of repository ventilation and waste emplacement sequence on
permeability changes have been evaluated for the local and global models,
respectively.

It was found that the most significant perturbation in the hydraulic per-
meabilities was caused by the excavation of the storage tunnels. The
change in permeability caused by the temperature and thermomechanical
stresses is expected to be small in comparison with expected uncertainty
in in situ permeability values. The permeabilities are about 5-20% lower
with storage tunnel ventilation than without ventilation.

The perturbed permeability fields will be used in subsequent analyses
of groundwater flow.

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7. Burgess, A.S., "GROUNDWATER MOVEMENTS AROUND A REPOSITORY,
   Phase 2. Technical Report 5. Repository Domain Groundwater Flow
   Analyses, Part II: Inflow to Repository, to be submitted to
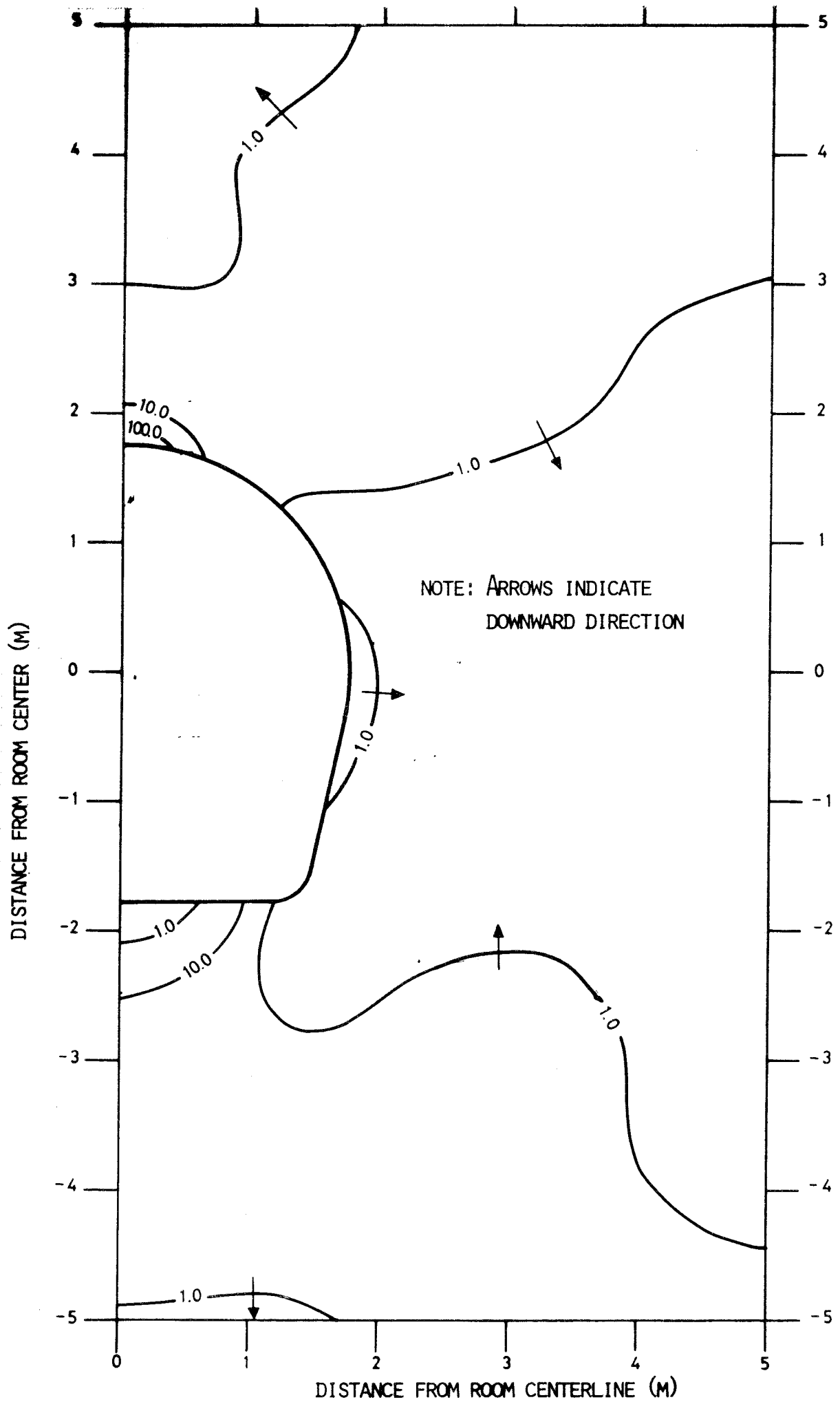   KBS - Kärnbränslesäkerhet, Stockholm, Sweden 1977.

FIGURE 1. RATIOS OF POST EXCAVATION HORIZONTAL PERMEABILITY TO IN SITU HORIZONTAL PERMEABILITY FOR 0^0 AND 90^0 JOINT SETS
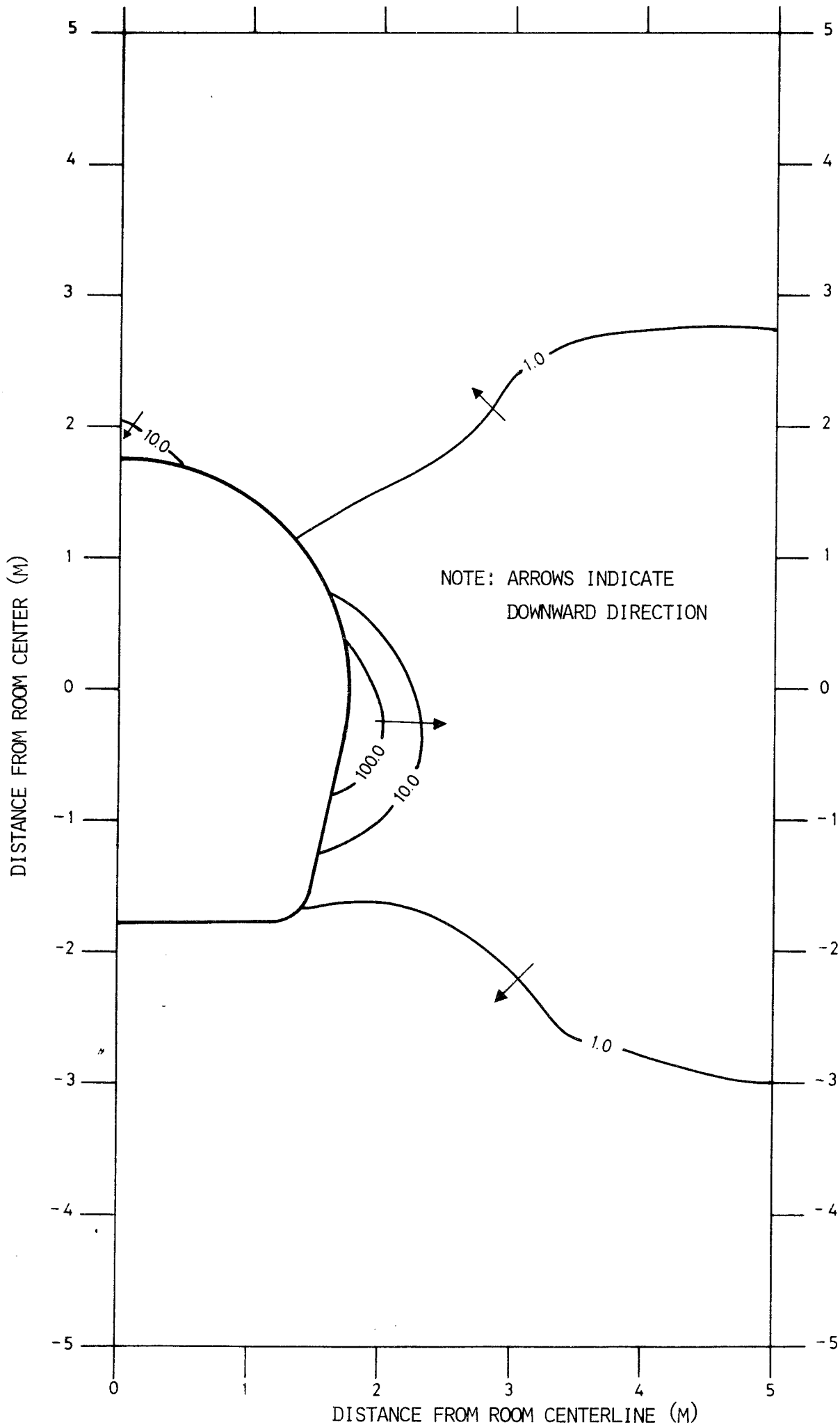
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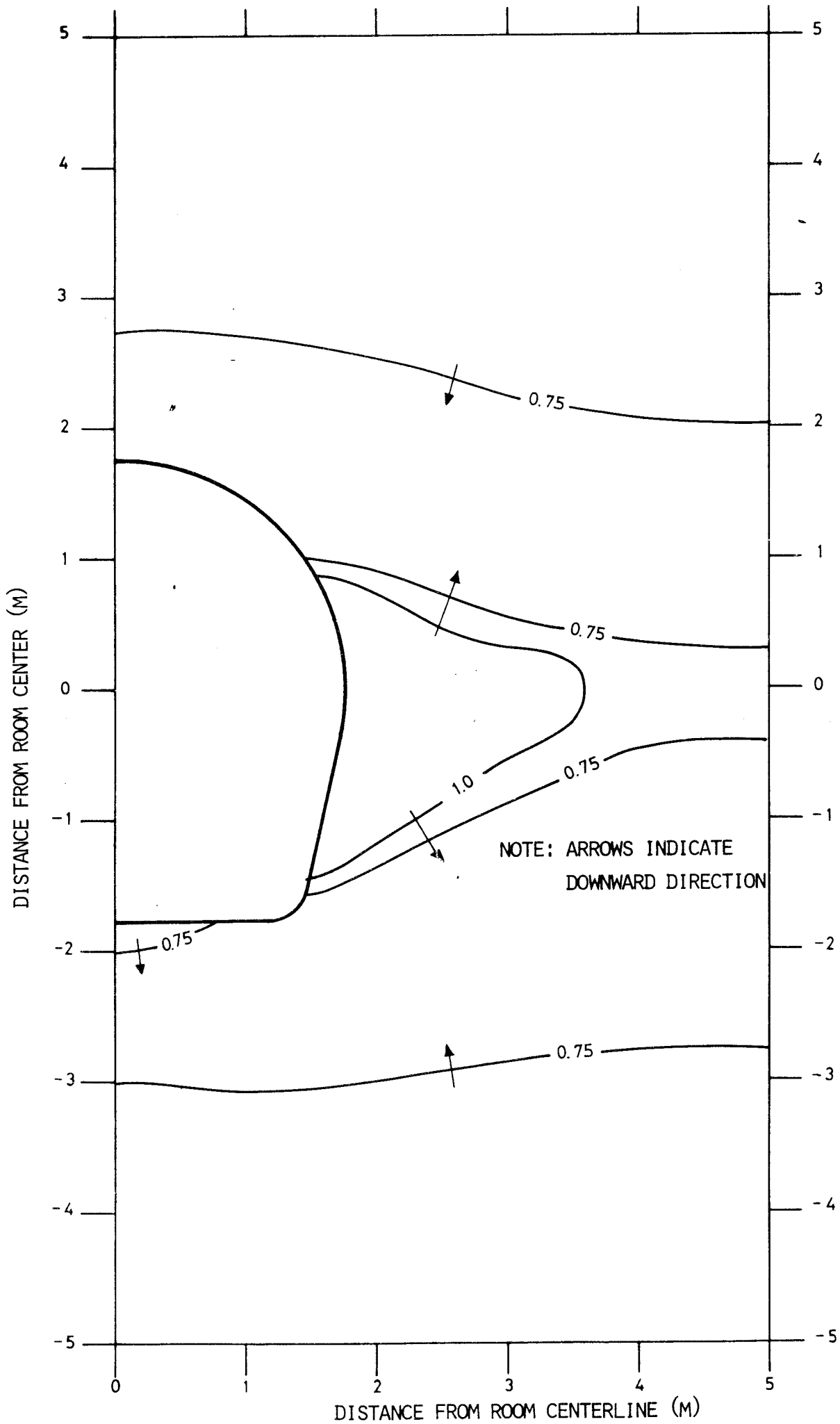VERTICAL PERMEABILITY OF 0° AND 90° JOINT SETS

FIGURE 3. RATIOS OF POST EXCAVATION NORMAL PERMEABILITY TO IN SITU NORMAL PERMEABILITY FOR 0° AND 90° JOINT SETS
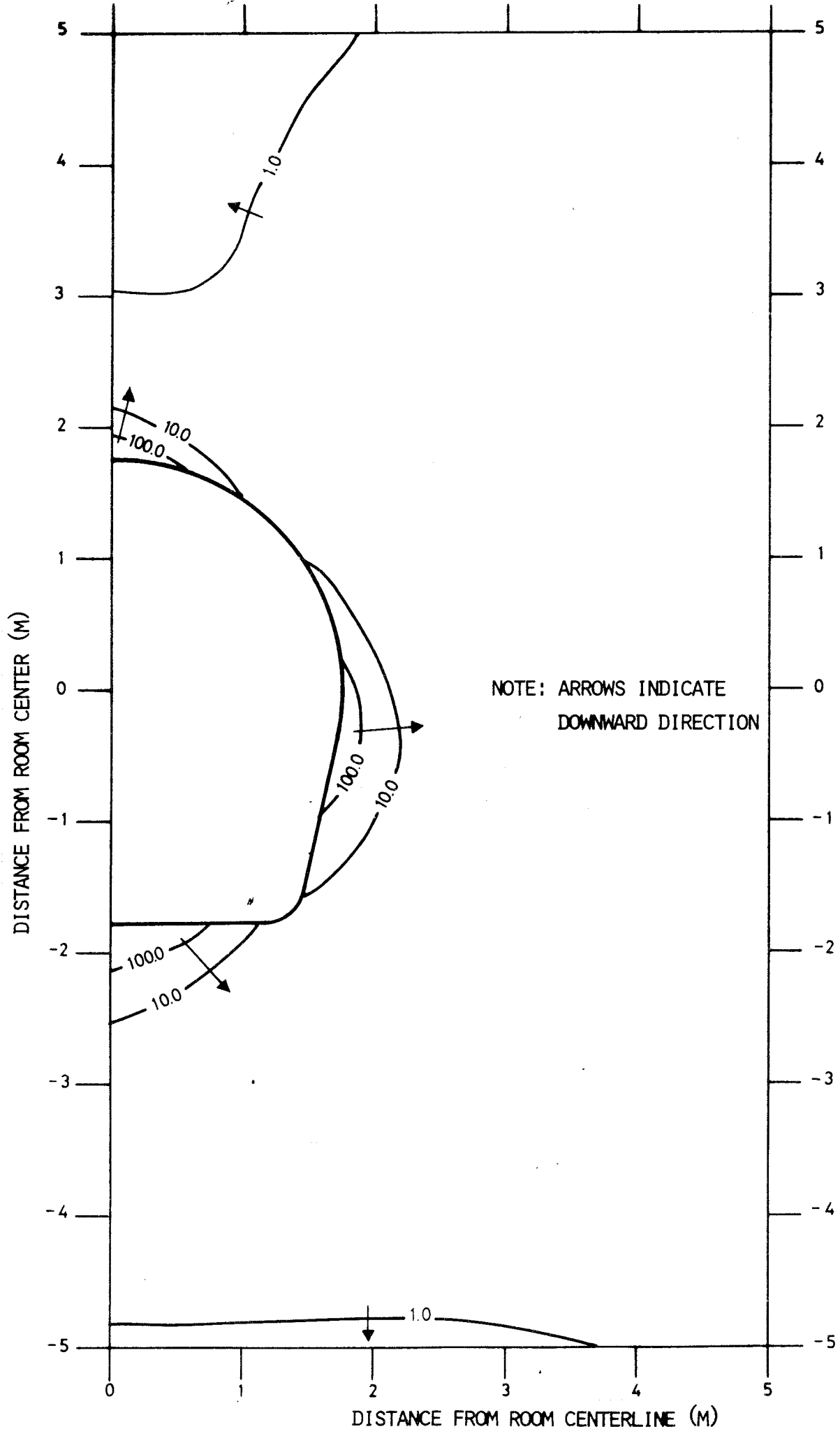
NOTE: ARROWS INDICATE
DOWNWARD DIRECTION

DISTANCE FROM ROOM CENTERLINE (M)

DISTANCE FROM ROOM CENTER (M)

FIGURE 4. RATIOS OF POST EXCAVATION TANGENTIAL PERMEABILITY TO IN SITU
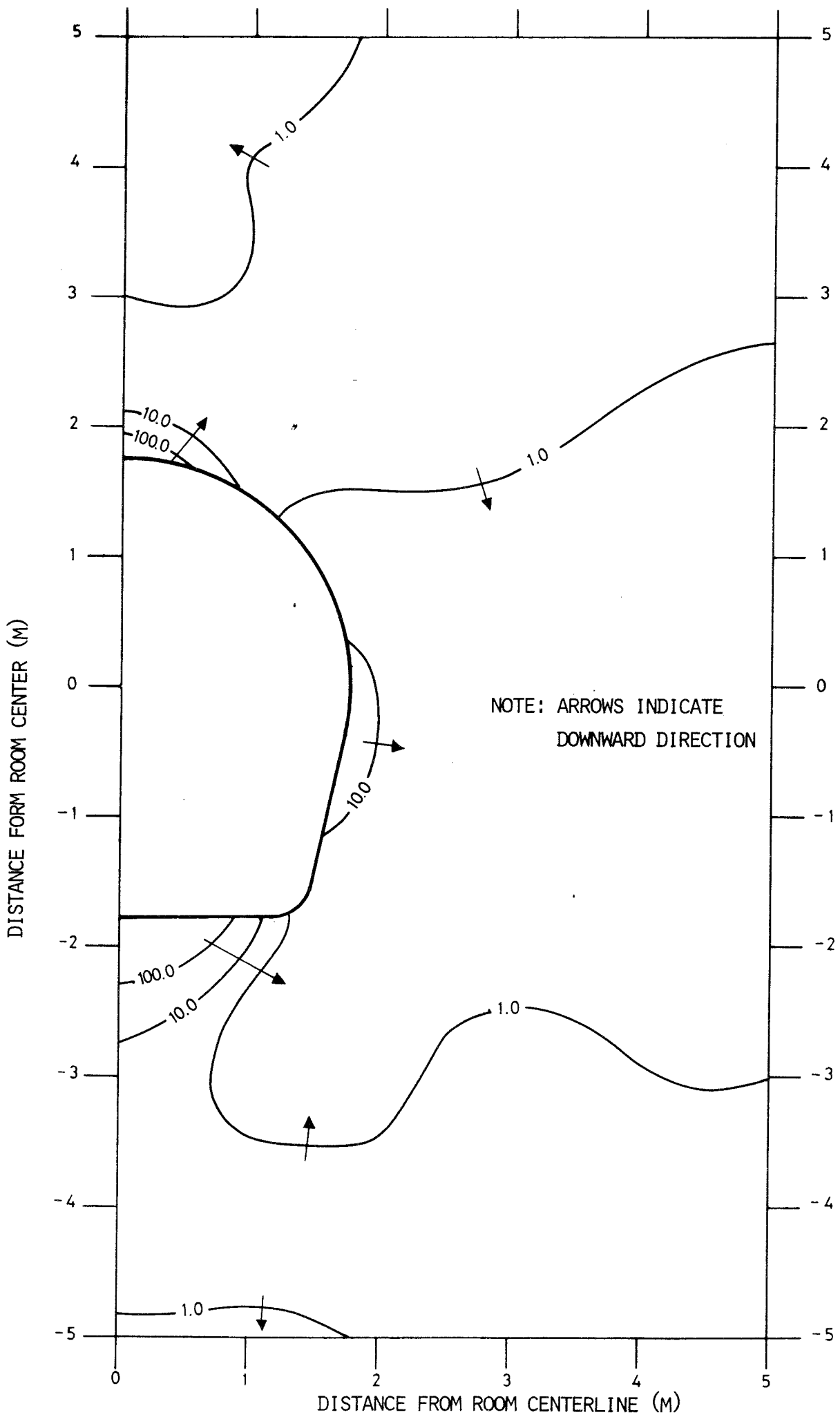TANGENTIAL PERMEABILITY FOR $0°$ AND $90°$ JOINT SETS.

FIGURE 5. RATIOS OF POST EXCAVATION HORIZONTAL PERMEABILITY TO IN SITU
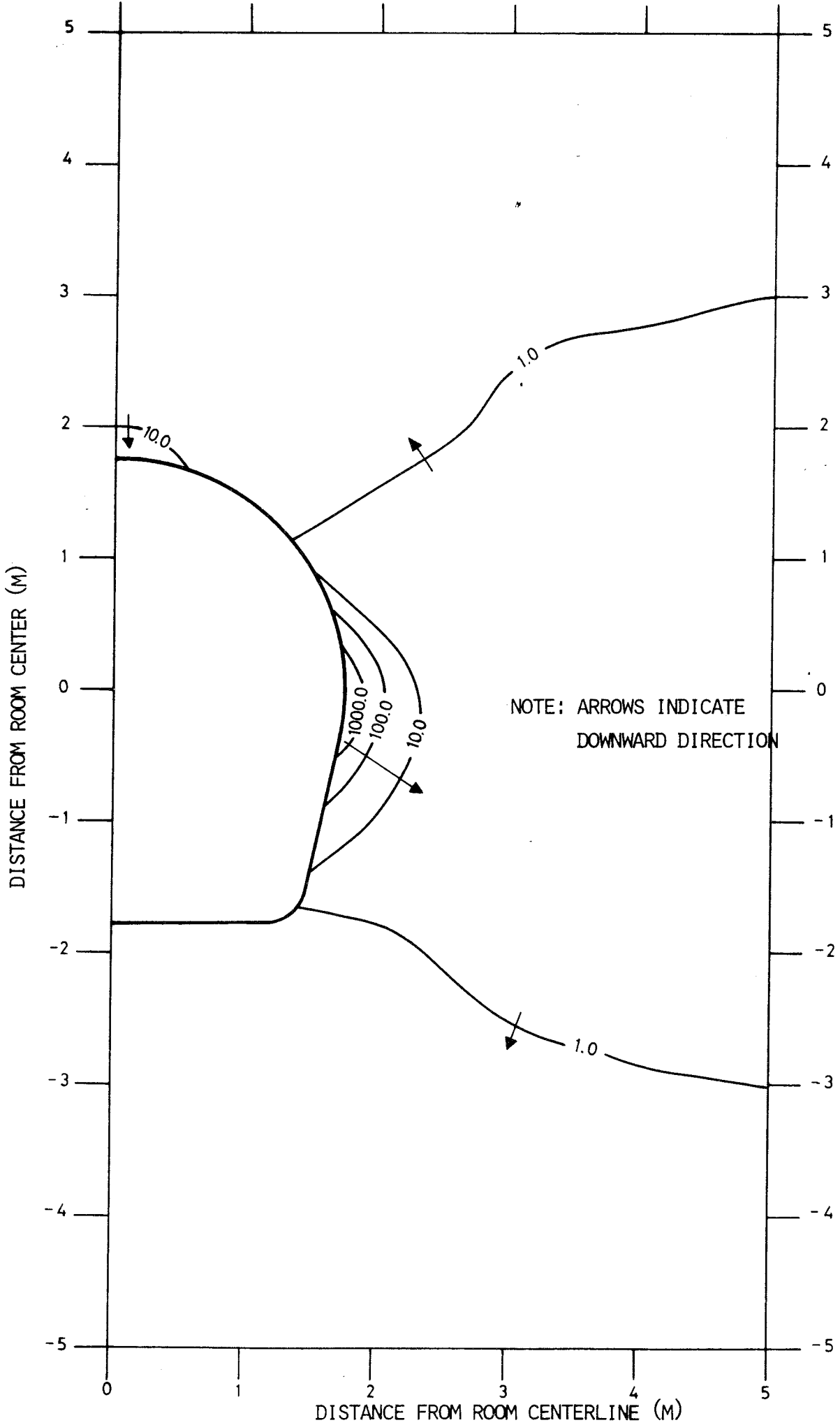HORIZONTAL PERMEABILITY FOR 45° AND -45° JOINT SETS

NOTE: ARROWS INDICATE
DOWNWARD DIRECTION

FIGURE 6, RATIOS OF POST EXCAVATION VERTICAL PERMEABILITY TO IN SITU
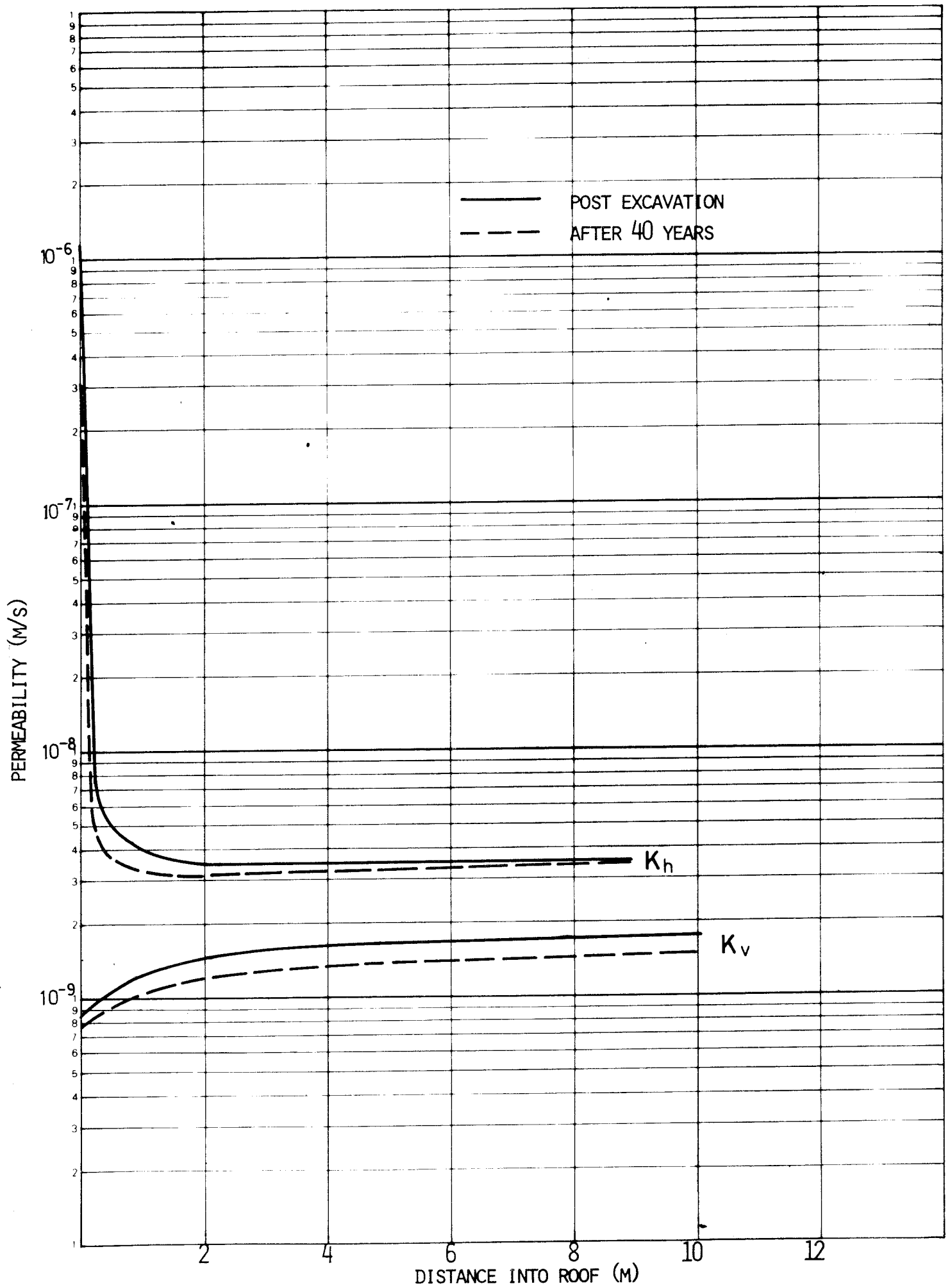VERTICAL PERMEABILITY FOR 45^0 AND -45^0 JOINT SETS

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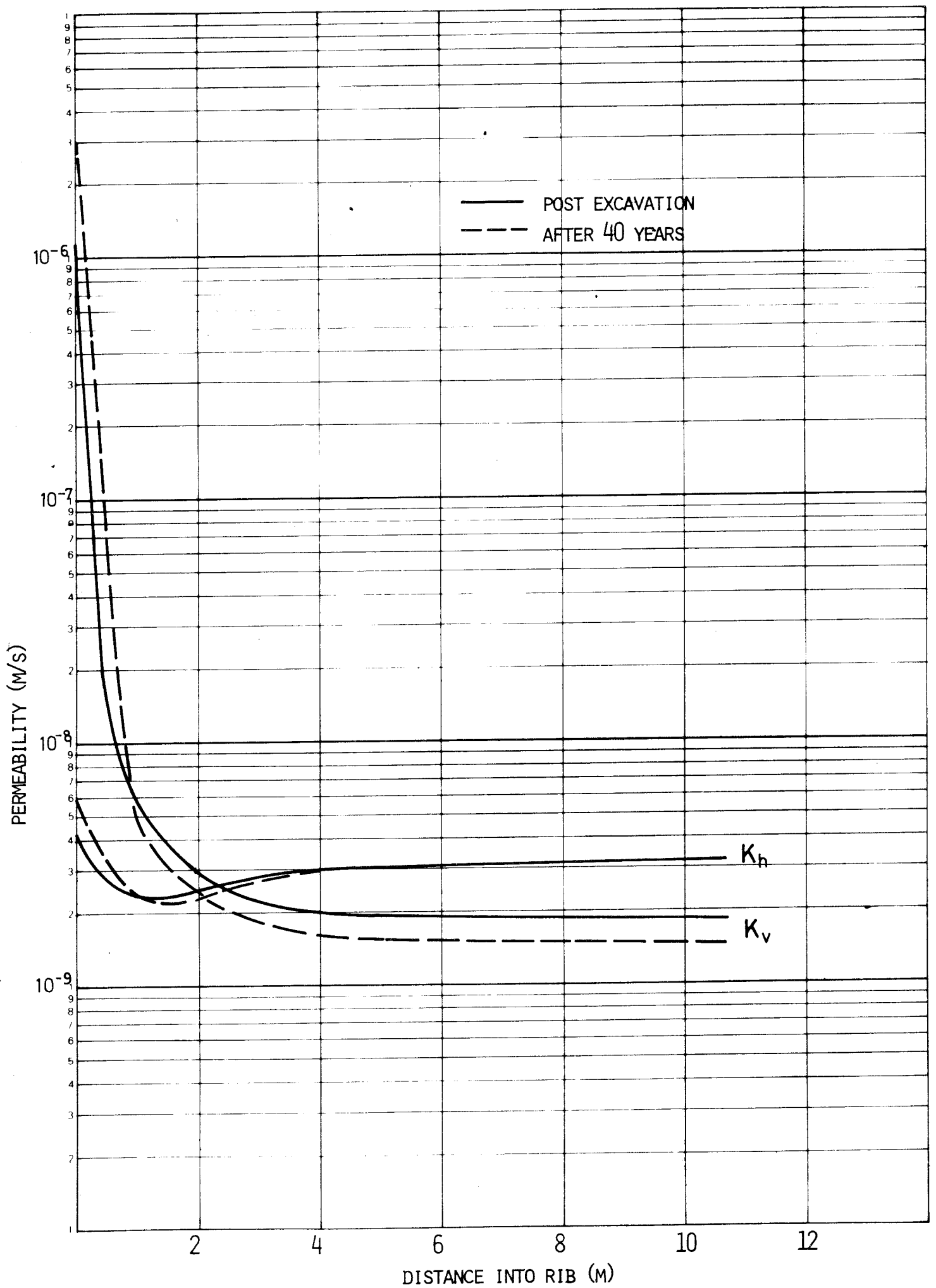NORMAL STRESSES AFTER 40 YEARS IN THE TUNNEL ROOF.

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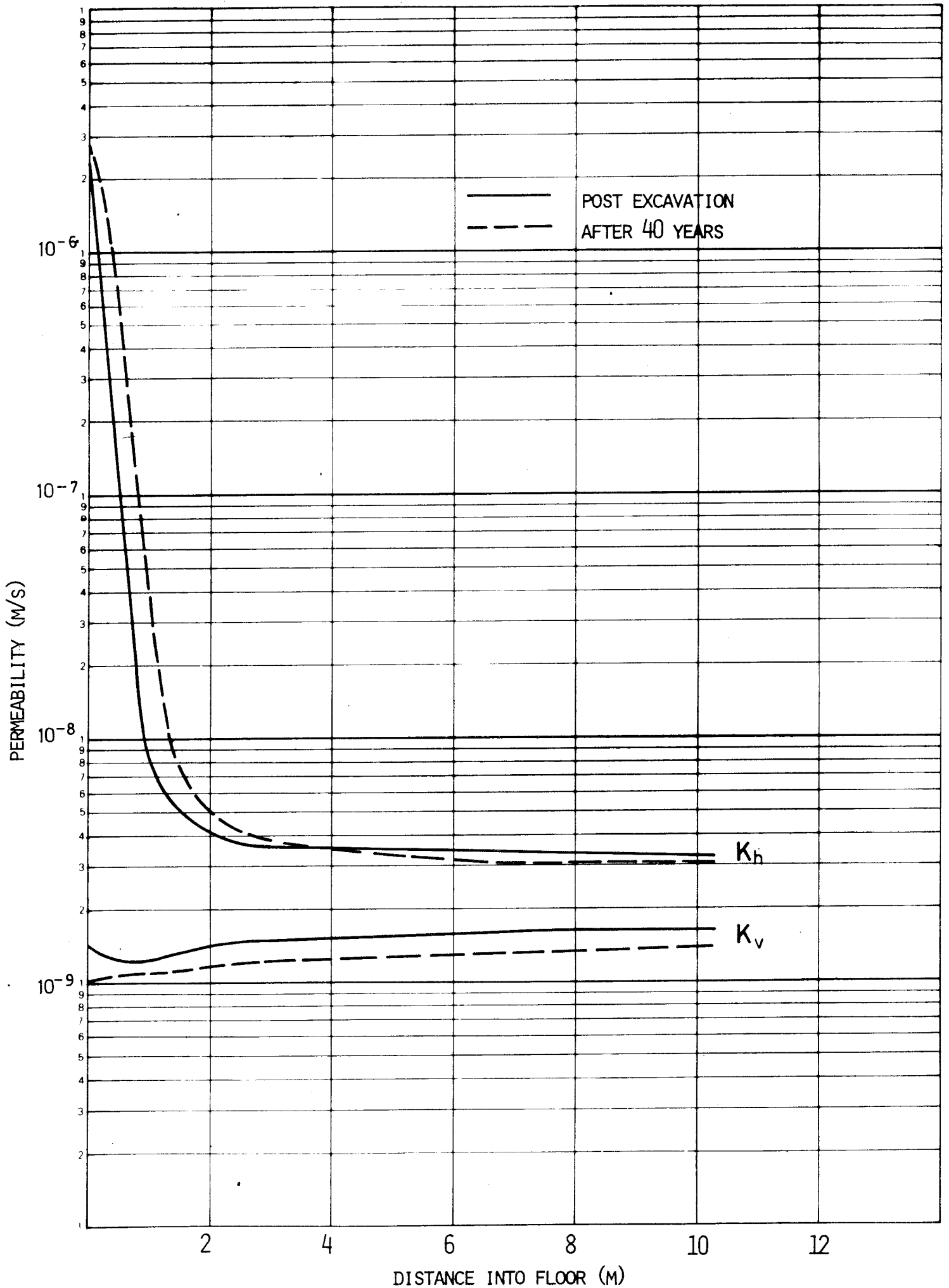
NOTE: ARROWS INDICATE DOWNWARD DIRECTION

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NOTE: ARROWS INDICATE
DOWNWARD DIRECTION

DISTANCE FROM ROOM CENTER (M)

DISTANCE FROM ROOM CENTERLINE (M)
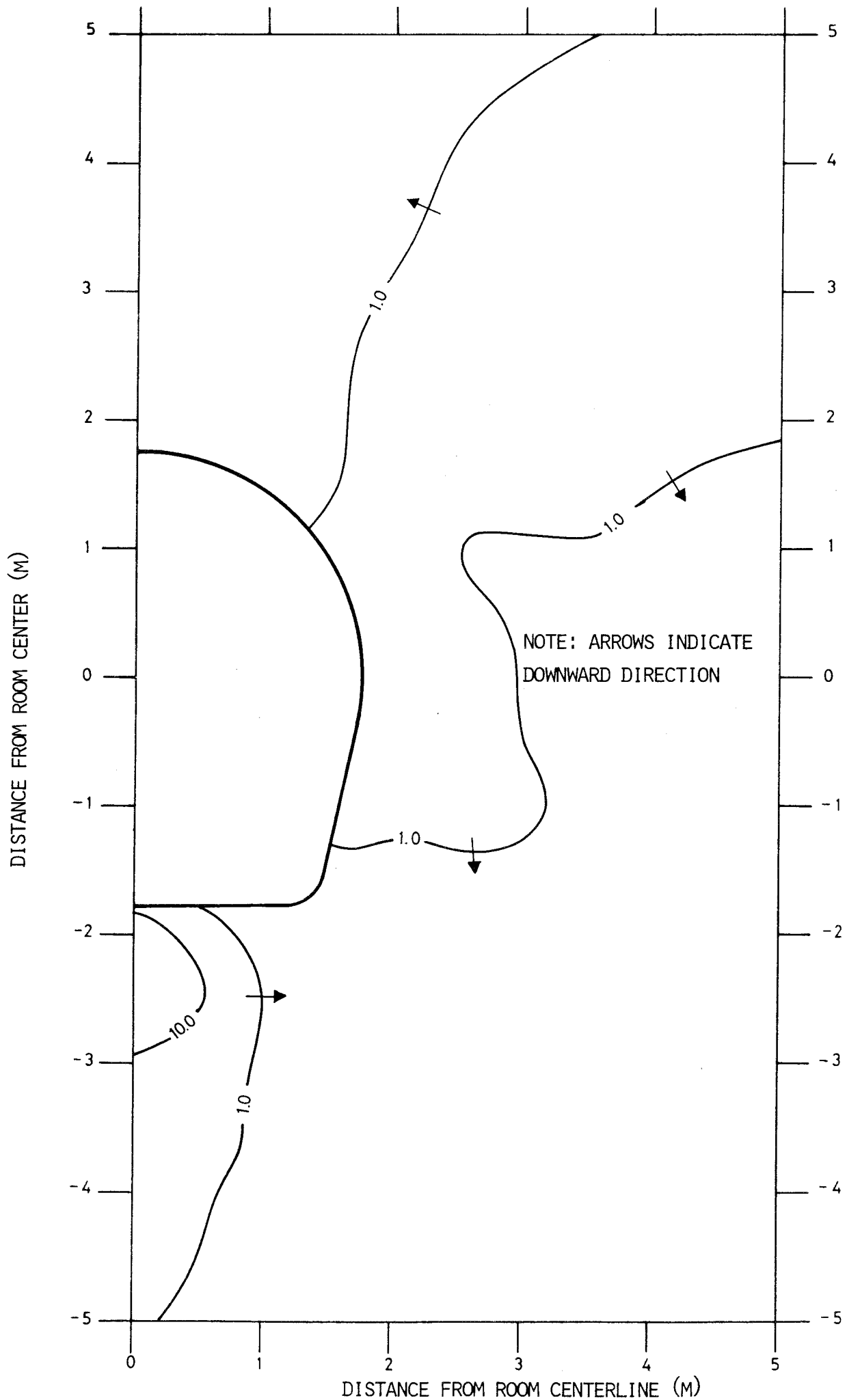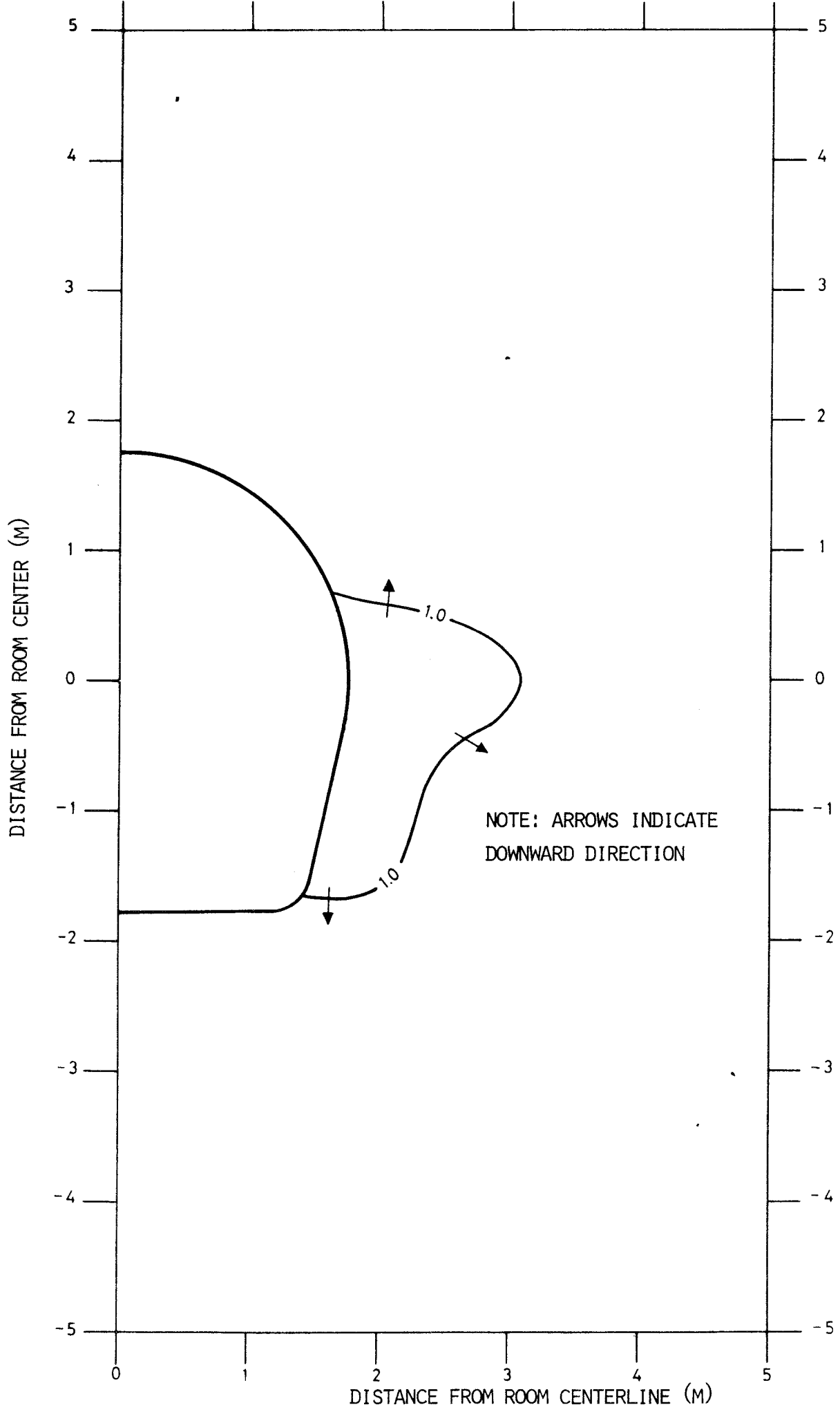
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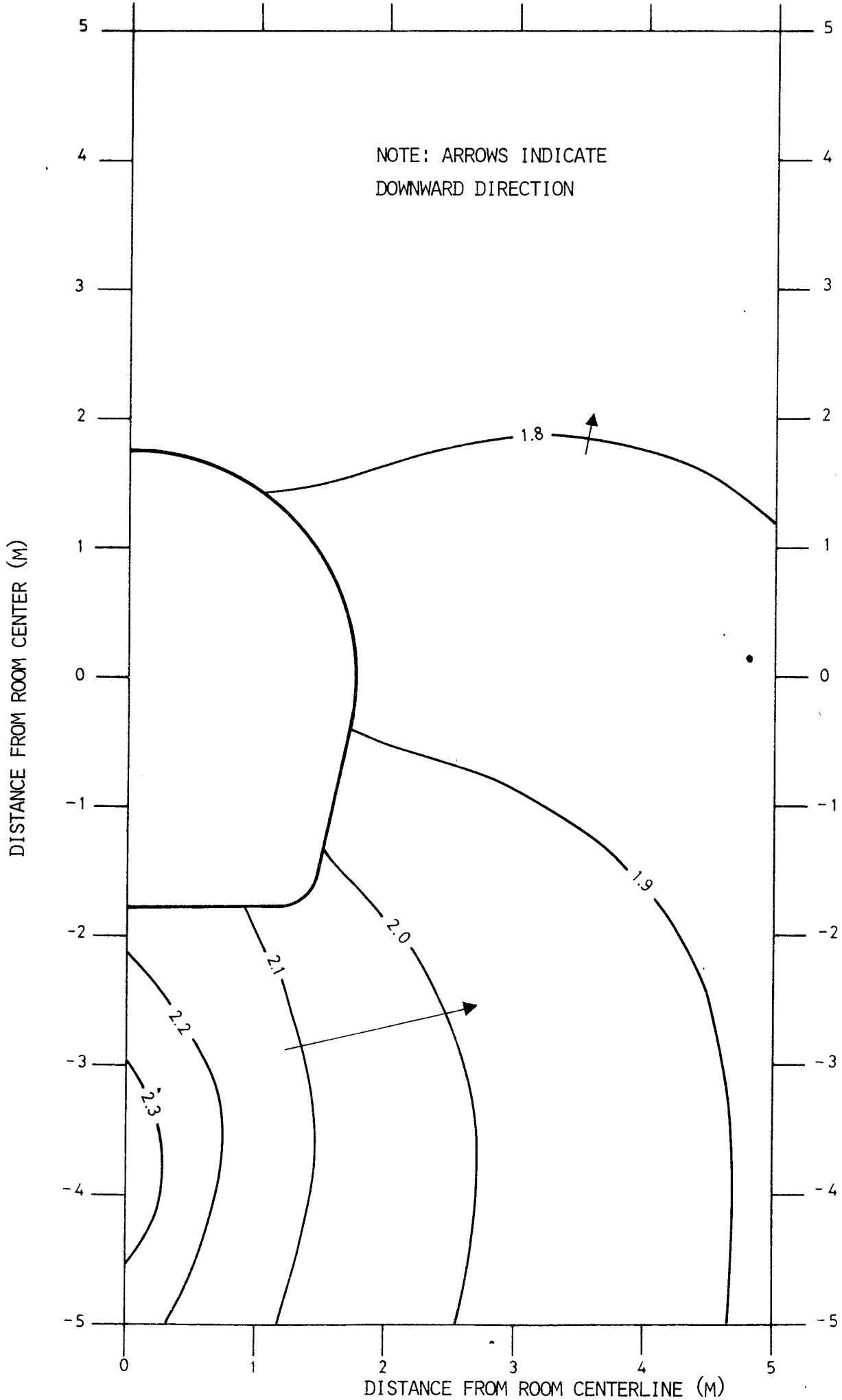
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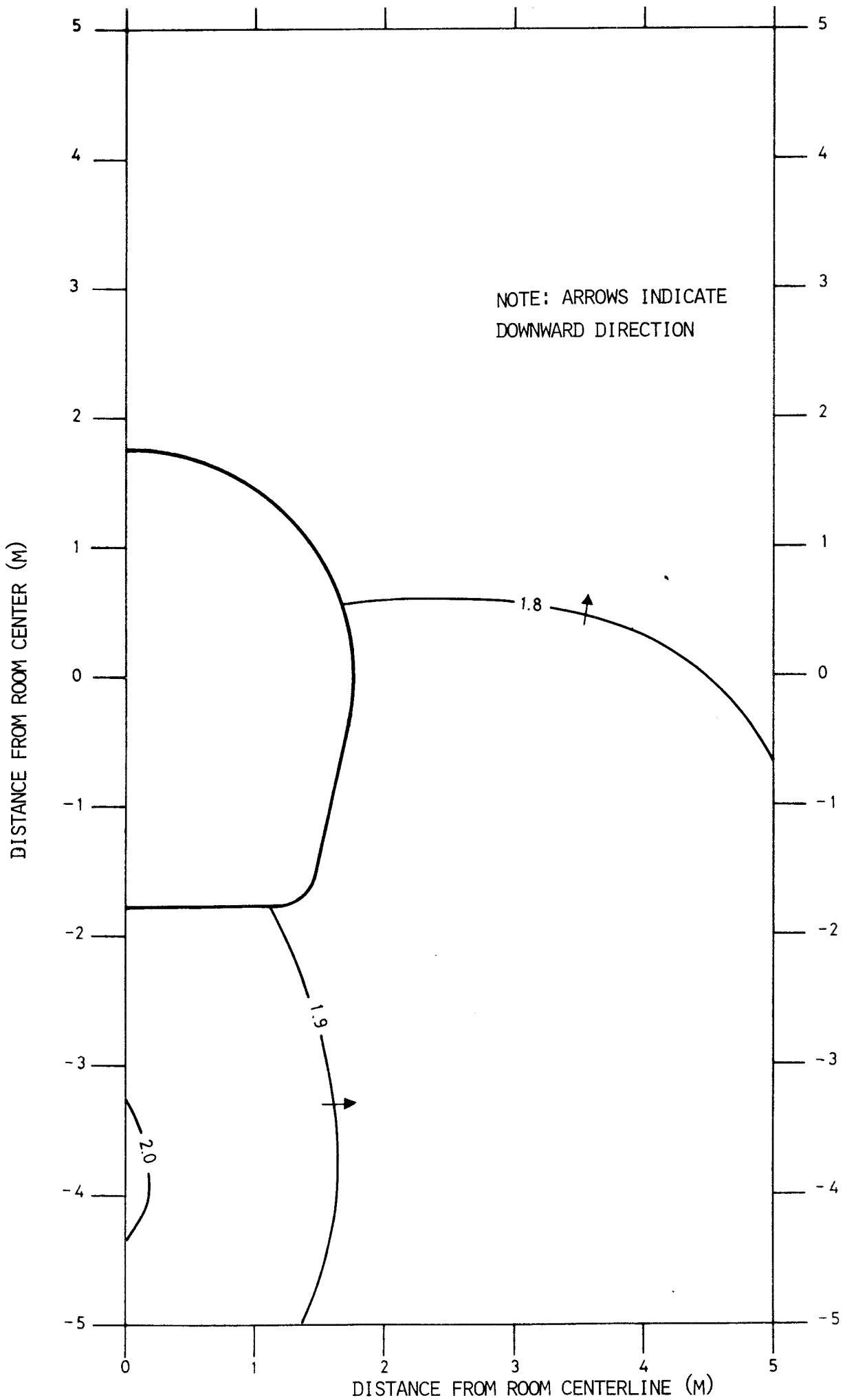
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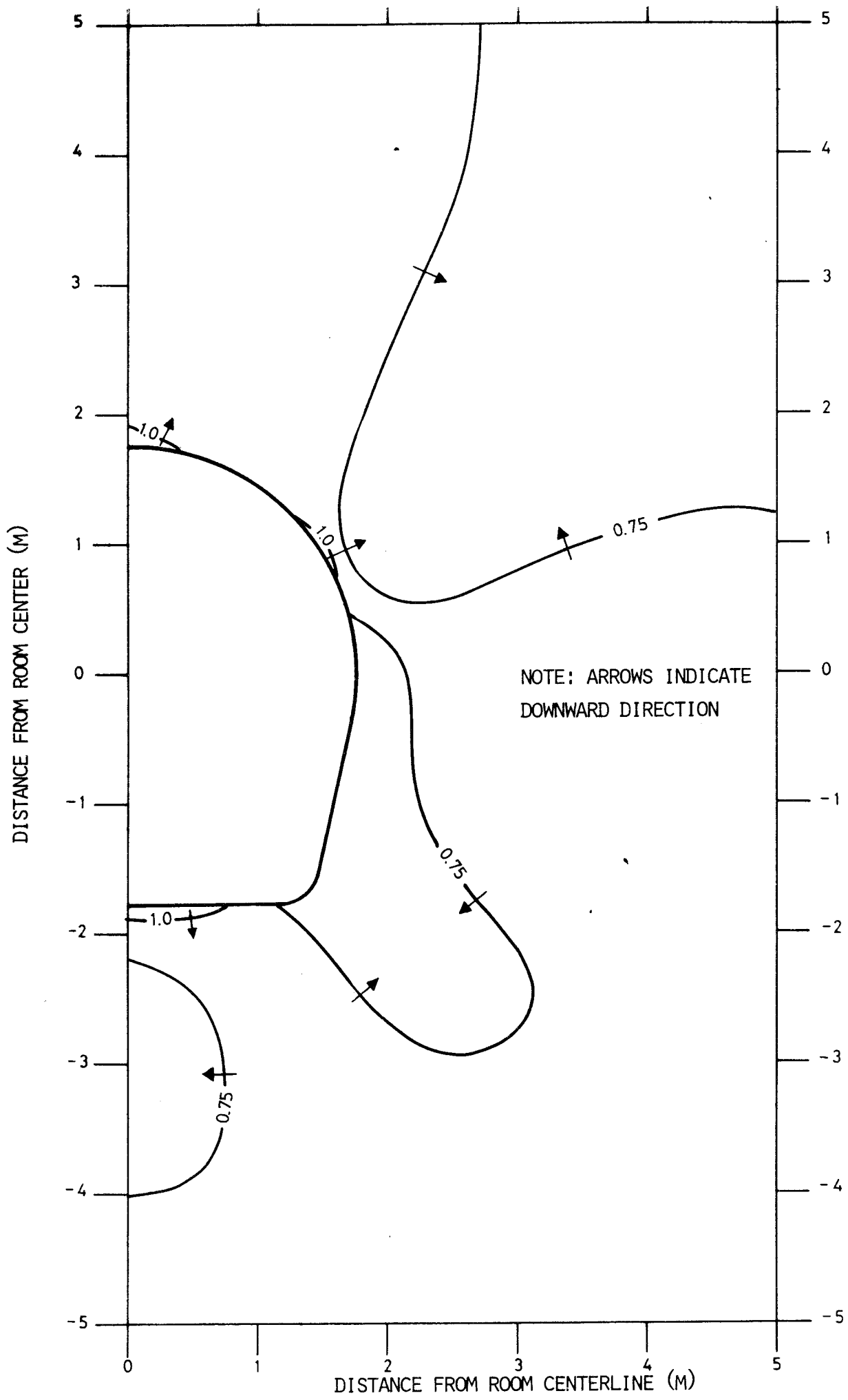
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NOTE: ARROWS INDICATE
DOWNWARD DIRECTION

DISTANCE FROM ROOM CENTER (M)

DISTANCE FROM ROOM CENTERLINE (M)
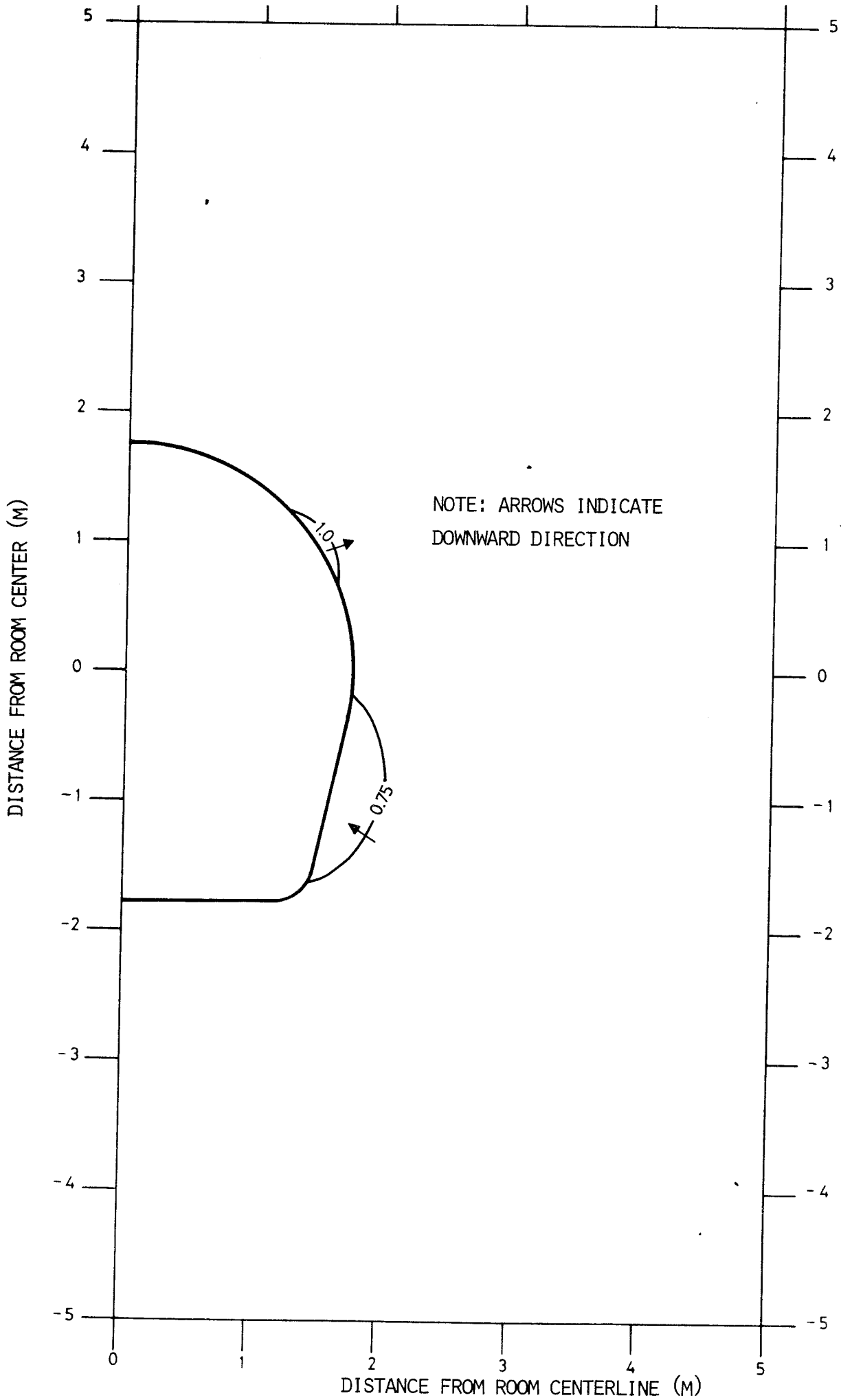
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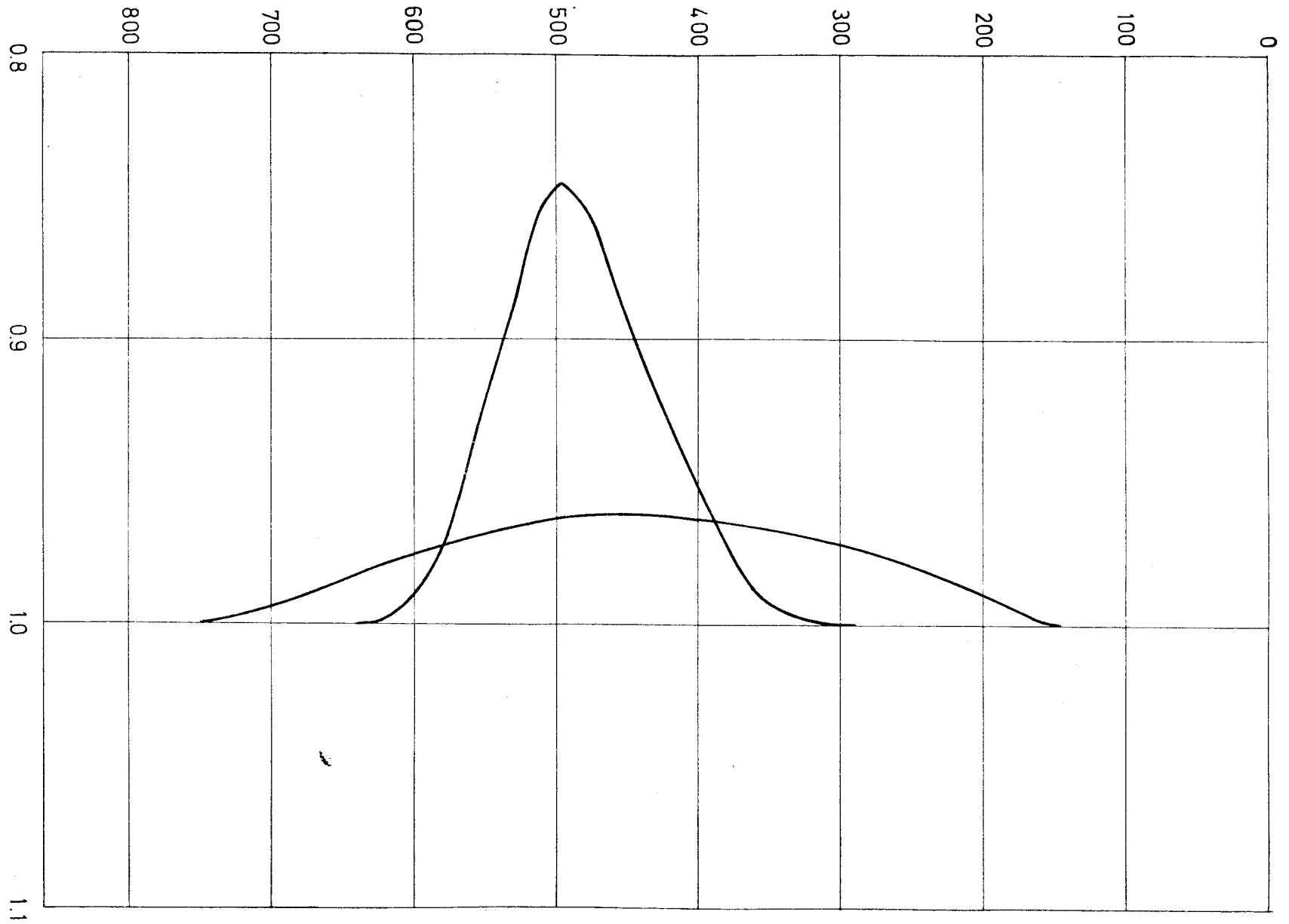
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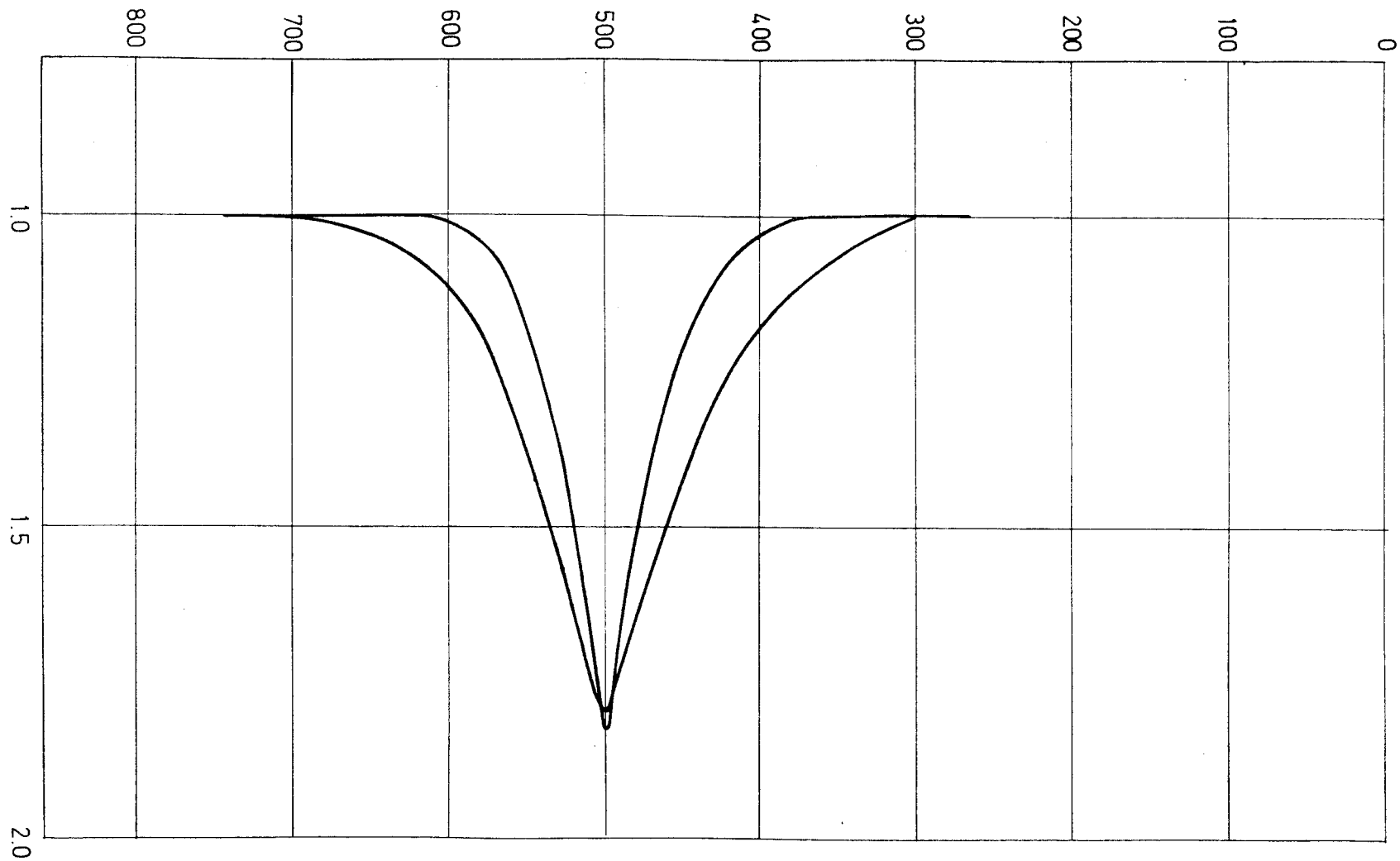
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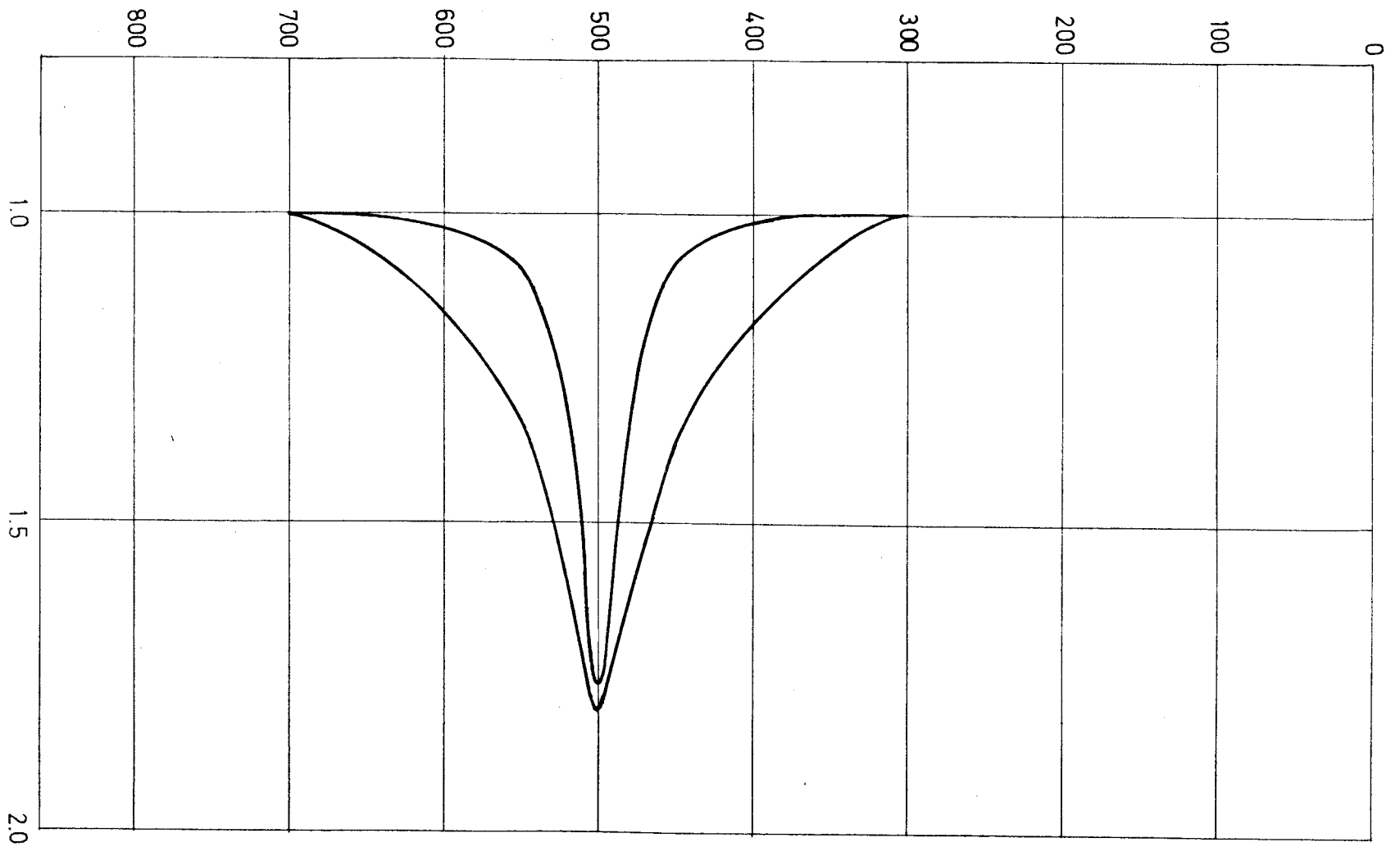
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TECHNICAL REPORT 5
REPOSITORY DOMAIN GROUND-
WATER FLOW ANALYSES
PART II
INFLOW TO REPOSITORY

GROUNDWATER MOVEMENTS AROUND A REPOSITORY

Phase 2.

Technical report 5: Repository Domain Groundwater Flow Analyses
Part II: Inflow to Repository

Hagconsult AB
in association with
Acres Consulting Services Ltd
RE/SPEC Inc.

FOREWORD

This report was prepared as one of a series of technical reports
within a study of the groundwater movements around a repository for
radioactive waste in the Precambrian bedrock of Sweden. It was written
in three parts, (I)Permeability Perturbations, (II) Inflow to the Re-
pository and (III) Thermally Induced Flow. This part is Part II. The
contract for this study was between KBS-Kärnbränslesäkerhet (Project
Fuel Safety) and Hagconsult AB of Stockholm, Sweden. RE/SPEC Inc. of
Rapid City, SD/USA and Acres Consulting Services Ltd of Niagara Falls,
Ontario/Canada acted as subconsultants to Hagconsult AB.

The principal authors of this report are Dr Anthony S. Burgess and
Dr Edward L. Skiba of Acres. Review was provided by Dr Ulf E. Lindblom
of Hagconsult AB and Dr Robin Charlwood of Acres. Input to the study
was provided by Dr Håkan Stille and Mr Joe L. Ratigan of the Study Group
and by other contributors to the KBS project.

The opinions and conclusions in this document are those of the author
and should not be interpreted as necessarily representing the official
policies or recommendations of KBS.

Stockholm September 1977

Ulf E. Lindblom
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1.   SCOPE AND OBJECTIVES

In this report we present the results of a study of groundwater inflow
to the repository, with particular reference to the post-decommission-
ing period. The objective was to determine the time for the repository
to become backflooded (inflow period). After this period, the ground-
water flow will return to the regional conditions with the addition of
any remaining thermal and/or mechanical perturbations. During the inflow
period, groundwater flow by definition is towards the repository. The in-
flow period may therefore be considered as a containment mechanism.

Finite element methods have been used to determine flow patterns. Cou-
pling of the permeability to stress changes has been incorporated and
the effects quantified. The effect of inflow on thermal conditions around
and within the repository has also been examined. Some aspects of con-
struction have been assessed, and their influence treated qualitatively.
The results of the model simulations have been compared with the limited
field data that is available.

For this study, the baseline layout has been used. This consists of rooms
approximately 3.5 m diameter, about 1 km in length and spaced at 25 m
centre to centre. The repository has been assumed to be located at a depth
of 500 m below ground surface.

2.   METHODOLOGY

2.1   General

All inflow periods, t_1, presented below are based upon an unfilled
room, i.e. porosity = 1. The inflow period t_n, for a room backfilled
with material having a porosity n, may be calculated from

$$t_n = n \times t_1$$

The rock mass with fractures has been treated as an equivalent con-
tinuum in the finite element method formulation for flow. To be valid,
this assumption requires that the fracture spacing be small compared
to the domain modelled. The results from the application of the con-
tinuum approximation to room-size models should therefore be considered
only as indicative of effects. However, a more sophisticated simulation
cannot be used until much more extensive field data is available. The
inflow study may be divided into three parts:

- Inflow using estimated initial baseline permeability distri-
  butions.

- Inflow using a permeability distribution determined from
  the insitu stresses and modified by construction stresses.

- The effects of inflow on heat transfer in the repository
  domain.

2.2   Inflow with Baseline permeability Distributions

For this part of the study, two finite element models have been employed:

- Repository domain model (Figure 1a). This represent a two
  dimensional vertical section through the repository. The
  repository centre line has been taken as a line of symmetry.

- Local model (Figure 1b). This also represents a two dimen-
  sional vertical section. The room centre line, and the pillar
  centre line have been taken as lines of symmetry.

The nominal permeability distributions used in the simulations were the same as used for the regional initial conditions. The basis for their selection is described in Technical Report 3 (1) [x).

Case 1.   Isotropic, homogeneous. Permeability $K = 1 \times 10^{-8}$ m/s.

Case 2.   Isotropic, non-homogeneous. Permeability decreasing with depth as per empirical function (1)

Case 3.   Anisotropic, non-homogeneous. Vertical permeability = $= 1.0 \times 10^{-9}$ m/s. Horizontal permeability decreasing with depth as per empirical function (1).

These permeability distributions are shown in Figure 2.

Throughout the analyses, the effect of singular geological features extending into the repository domain have not been considered. If such features exist then

a)   the repository will be sited to avoid them, or

b)   they will be treated by grouting or other techniques to give them material properties similar to the surrounding rock.

For both the repository domain and local model, the vertical boundaries which are lines of symmetry have been taken as flowlines, i.e. impervious boundaries. In addition, the far field vertical boundary in the global model has been considered as an impervious boundary.

For the repository domain model, the groundwater surface has been assumed coincident with the ground surface. The existence of drawdown was checked by a comparison of the boundary flux with the available recharge flux. For the regions of Sweden where the repository may be located, annual precipitation is typically 500 mm. Infiltration coefficients for typical soil conditions vary from 0.11 to 0.37 (2). Using value of 0.2 a recharg of 100 mm/year is available. This is equal to a boundary flux of 3.17×10^{-9} m/s.

x) Numbers in parentheses refer to references at end of text.

In the inflow simulations, boundary fluxes less than the available recharge flux will be satisfied by infiltration alone. If boundary fluxes are higher than the recharge, drawdown of the groundwater surface will result, until an equilibrium is achieved between horizontal inflow, vertical recharge and downward flow. The extent of the drawdown will depend upon the variation of permeability with depth and the degree of anisotropy. In general, drawdown will be least for the aniso-tropic condition where the horizontal permeability is relatively high at the surface.

The potential distributions from the repository domain models were used to determine boundary potentials on the local models. The boundary fluxe; on the local models were then compared with the repository domain fluxes to assess the influence of geometry effects at the repository level.

At the repository itself, two boundary conditions have been considered:

- Free inflow, i.e. any air within the repository escapes or goes into solution at nominal pressures.

- Inflow with backpressure, i.e. any air within the repository is unable to escape or go into solution and causes pressure rises within the repository.

2.3   Inflow with Permeabilities Modified as a Function of Stress

For this part of the study, a finite element model extending from the ground surface to a depth of 1000 m was used (Figures 3a and 3b). The vertical boundaries were lines of symmetry for the room centre-line and the middistance between adjacent rooms respectively. The technique used for the unmodified permeability distributions, i.e. separate repository domain and local models, was found to be unsatisfactory for this part of study. This was because the repository domain mesh was too coarse in the immediate area of the repository and the significant permeability change; could not be modelled. As a result the boundary conditions for the local model could not be accurately determined.

The initial permeability distribution (Case 4) was made compatible with the initial in-situ stresses for a horizontal to vertical stress ratio of 2. The horizontal permeability at depth z was set equal to the horizontal permeability calculated from the empirical function (1). The vertical permeability at depth z was then equated to the horizontal permeability at depth $2z$. The initial permeability values were then modified using the stress-permeability relationship described in part I of this report (3), for the stress changes due to construction. This is referred to as the Case 5 permeability distribution. The ratios of modified to initial permeability values are given in Figures 4a and 4b.

Boundary conditions for this simulation were the same as those discribed in section 2.2., with the exception that the backpressure case was not considered.

2.4   Effect of Inflow upon Thermal Conditions

The advection of heat by inflow was studied using the repository domain model with a thermally induced flow finite element technique. Details of the computational method are given in part III of this report (4).

In order to simulate inflow lasting for a significant part of the thermal cycle, the permeability distribution was taken to be isotropic, homogeneous, with a value of 10^{-10} m/s. The initial gross thermal loading was taken as 5.25 W/m^2, and the thermal conductivity of the rock mass as 2.05 W/m, $^{\circ}$C.

Boundary conditions for this part of the study were the same as detailed in section 2.2.

2.5   Reliability of Inflow Period Predictions

Nominal permeability distributions have been determined by extrapolating to depth permeability values obtained closer to the surface. As a result, apparently short inflow periods result. The inflow period is a function of the permeability in the vicinity of the repository, and of the variation of permeability with depth. In general, the inflow period is

inversely proportional to the permeability in the repository area. Thus an error of one order of magnitude in estimating the permeability would change the inflow period by an order of magnitude.

3. RESULTS

3.1 Free Inflow

The equipotentials for Cases 1 to 5 are shown in Figures 5 to 9 respectively.

The inflow times, inflow rates and upper boundary fluxes are tabulated below. The inflow period is seen to be strongly correlated with the vertical permeability in the vicinity of the repository. The influence of the other factors (anisotropy, horizontal permeability, variation of permeability with depth) on the inflow time can be illustrated by comparing the values of $t_1 \times K_z$ for the different cases. (K_z is the vertical permeability at a depth of 500 m.)

	Flux at ground surface m/s	Inflow rate 1/min, km	K_z m/s	Inflow periods year	$t_1 \times K_z$ m
Case 1	12.13×10^{-9}	28.85	10^{-8}	0.64	.20
Case 2	10.14×10^{-9}	17.97	3.7×10^{-9}	1.02	.14
Case 3	1.77×10^{-9}	4.17	10^{-9}	4.39	.14
Case 4	3.66×10^{-9}	6.28	1.7×10^{-9}	3.17	.17
Case 5	3.69×10^{-9}	6.34	1.7×10^{-9}	3.14	.17

For Cases 1 and 3 where K_z is constant with depth, about 65% of the total flux comes from above the repository and 35% from below. For the Cases 2, 4 and 5 in which the vertical permeability decreases with depth the ratio changes to about 85% from above the repository and 15% from below the repository.

From the global model analyses for Cases 1, 2 and 3 the distribution of inflow fluxes laterally across the repository can be assessed. The inflow flux pattern for the homogeneous isotropic permeability distribution (Case 1) is shown in Figure 10.

The total inflow rate is about 50% higher towards the outer limit of the repository, compared with the inner regions. The assumption of vertical flowlines for the local models of Case 1, 2, 3 and for the models for Cases 4 and 5 are therefore valid only for the central areas of the repository. As a result, the inflow periods based on vertical flowlines vill be somewhat longer than if the complete repository domain flow pattern were to be used. As discussed in sections 2.2 and 2.3, however, modeling limitations for Cases 4 and 5 required that vertical flowline boundaries be adpoted. For comparison proposes, similar conditions have therefore been applied to Cases 1, 2 and 3.

Comparison of inflow periods for Cases 4 and 5 shows that the effect of permeability changes around the room due to construction stresses is very small. This is due to the following reasons:

- The stress changes around the room result in both increases and decreases in the permeabilites in various locations which tend to equalize each other.

- The changes in permeability result in changes in the potential gradients. Thus a decrease in relative permeability results in an increase in gradient in that region.
The effect of the change on the flux is therefore reduced.

Contours of flux magnitudes for Cases 4 and 5 are shown on Figures 11 and 12 respectively. For Case 4, the inflow flux is highest in the roof of the room, and lowest in the floor. The flux distribution for Case 5 is much more variable, reflecting the complex modifications to the insita permeabilities.

3.2   Inflow with Development of Backpressure

The effect of backpressure within the room was determined for Cases 1, 2 and 3. Figure 13 shows the curves representing both free and pressurized inflow results. The percentages of the room flooded as a function of time and the corresponding decrease in flux due to the backpressure effect are also indicated. The time scale is distorted below the graph for the three permeability cases, the times being inversely proportional to the initial

fluxes. It can be seen that there is only a small different between the free inflow and backpressure cases. The flux curve, however, demonstrates that for the latter the flux starts to decrease at roughly the free filling time and decays 5 order of magnitude at 1.26 of the free filling time and 10 orders at 1.54 of the free filling time. The relatively minor nature of the effects of pressurization is a result of the great depth of the repository for which the recharge pressure is nearly 50 atmospheres due to static water head.

3.3   Inflow Effects on Thermal Conditions

The thermal capacity of the tunnels, filled with water by inflow, is 1.61×10^{-6} J /m^2, $^{\circ}$C. Assuming that the water may reach a temperature of the order of 30°C above ambient and using the initial gross thermal loading (GTL) of 5.25 W/m^2, the time required for the initial GTL to raise the room-filling water to 30°C is about 0.3 years. Since a great deal more heat than this is available over the years, the thermal capacity provided by the water in the rooms and hence its effect on temperatures in the rock will be minimal. Only in the event of substantial inflow and pumping could the effect be significant, similar to ventilation.

The conclusion is that inflow to the rooms can have little advective effect on temperatures whatever value of permeability is used. The fluid advects some heat into the rooms but provides extra heat storage capacity as well. The temperatures near the waste can only be expected to drop slightly due to inflow of water into the rooms.

In order to quantify the effect of the advectiv heat transfer due to inflow, a coupled analysis was made. The methodology is described in part III of this report. An arbitrary isotropic homogeneous permeability of 10^{-10} m/s was used in this analysis to provide a case representative of inflow lasting for a significant part of the thermal cycle. For this permeability, the inflow period would be about 65 years. The changes in temperatures due to advection were less than 0.1°C for temperature increases of about 30°C at 70 years. The global water flux associated with this inflow is shown in Figure 14.

4.   DISCUSSION AND CONCLUSIONS

4.1   Comparison with Field Data

Field data on inflow to excavations at depths of hundreds of meters is
primarily from measurements of inflow into mines. A study of inflow to
mines in the Canadian Precambrian Shield is presented by Raven & Gale
(5). The inflow rates were normalised by expressing the inflow as liters
per minute per kilometer of lateral development in the mine (1/min,km).
They concluded that mines with pumped levels within 150 m of the ground
surface showed higher inflow rates than mines where the highest pumped
level was deeper than 150 m. Values for the latter condition, pertiment
to the present study, have been extracted from Raven & Gale and are
given in Table 1.Apart from the value of 61.5 for East Malarctic No 5
shaft, the inflows are between 0 and 3 1/min,km.

As part of the field studies for KBS, measurements of inflow rates will
be performed at Stripa Mine. A rough estimate made by the author during
a site visit indicates an approximate inflow of 3×10^{-2} 1/min,km.
Permeability test indicate that they are of the order of 5×10^{-11} m/s.
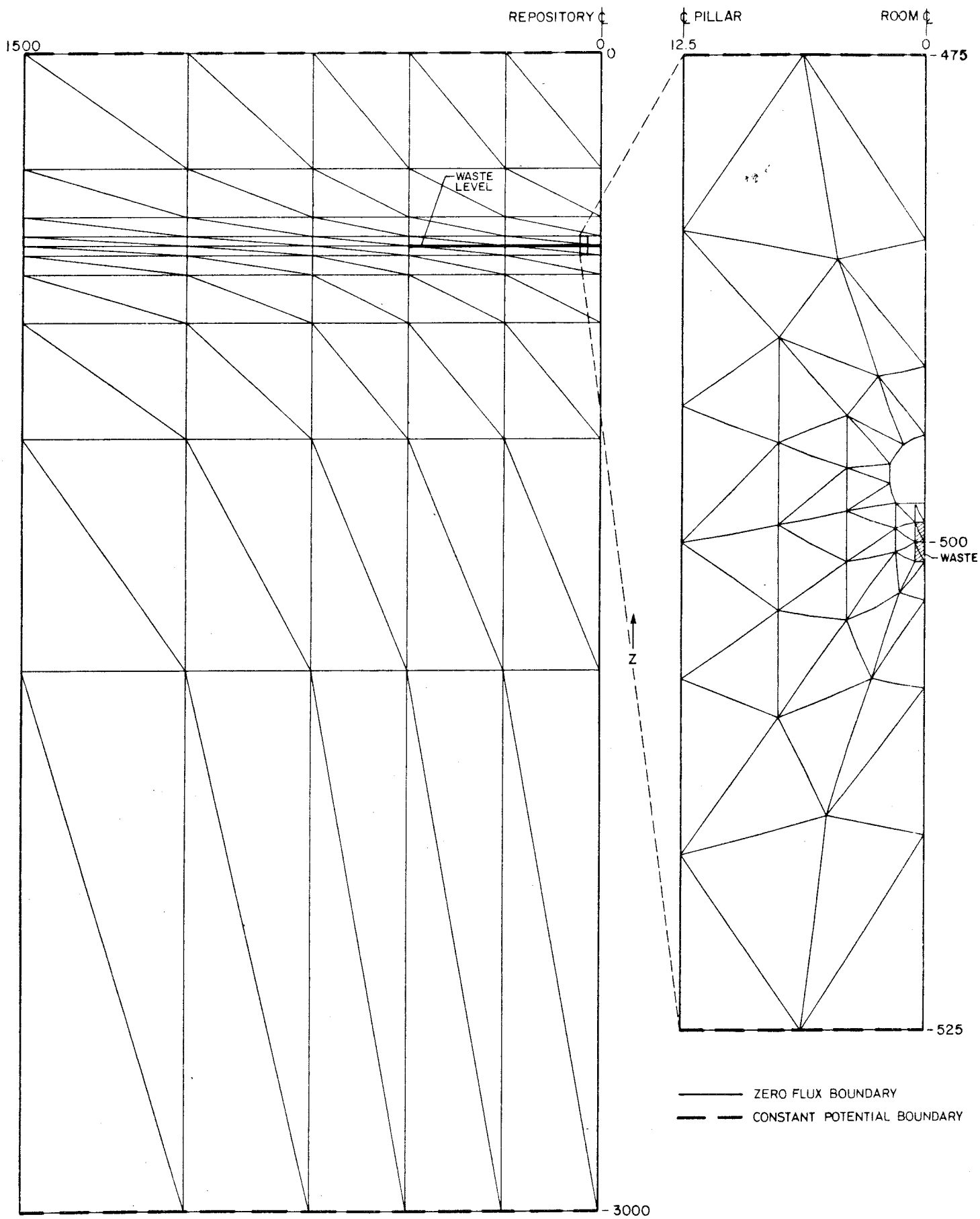
A comparison of these very linutes field data with the inflow rates
from the model simulations indicates that the model values are relatively
high by about an order of magnitude. The final conclusions must await the
results of further field data on the permeability distributions.

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4. Skiba, E.L., 1977, "Groundwater Movements Around a Repository",
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5. Raven, K.G. & Gale, J.E., 1977, "Subsurface Containment of Solid
   Radioactive Waste: a study of the surface and subsurface structural
   and groundwater conditions at selected underground mines and open-
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   740057, EMR/GSC RW Int Rep No 1/77.

TABLE 1.   ESTIMATES OF SEEPAGE WATER PUMPED FROM MINES (5).

MINE & DATE OF SHAFT SINKING	DOMINATE ROCK TYPE(S)	SEEPAGE PUMPED 1/min	PERIOD OF PUMPING	DEPTH OF LEVELS PUMPED m	SEEPAGE PUMPED PER KM OF LATERAL DEVELOPMENT 1/min,km	MAXIMUM DEPTH OF VISIBLE CONTINUOUS SEEPAGE m
Macassa (1917)	Augito Syenite	115	1971	150-2100	1.5	≯915
North (1955)	Norite	3	1975	777-1200	0.2	>1150
New Quirke (1955)	Meta Conglomerate Quartzite	0	1975	260-683	0	520
Strathcona (1964)	Granite Gneiss Norite	0	1975	365-915	0	365
Millenbach (1969)	Andesite, Quartz Feldspar Porphyry	45	1975	716-1224	3.0	355
Langmuir (1971)	Meta Volcanic	6	Oct 1975	154-378	0.6	>378
East Malartic No.5 shaft (1969)	Greywacke Syenite	0 / 680	1975 / April-May 1975	1160-1475	0 / 61.5	<1160

REPOSITORY ℄   ℄ PILLAR   ROOM ℄

1500   0   12.5   0 — 475

WASTE
LEVEL

Z

— 500
— WASTE

— 3000

— 525

—————— ZERO FLUX BOUNDARY
— —— — CONSTANT POTENTIAL BOUNDARY

(a) REPOSITORY DOMAIN HALF MESH   (b) LOCAL HALF MESH
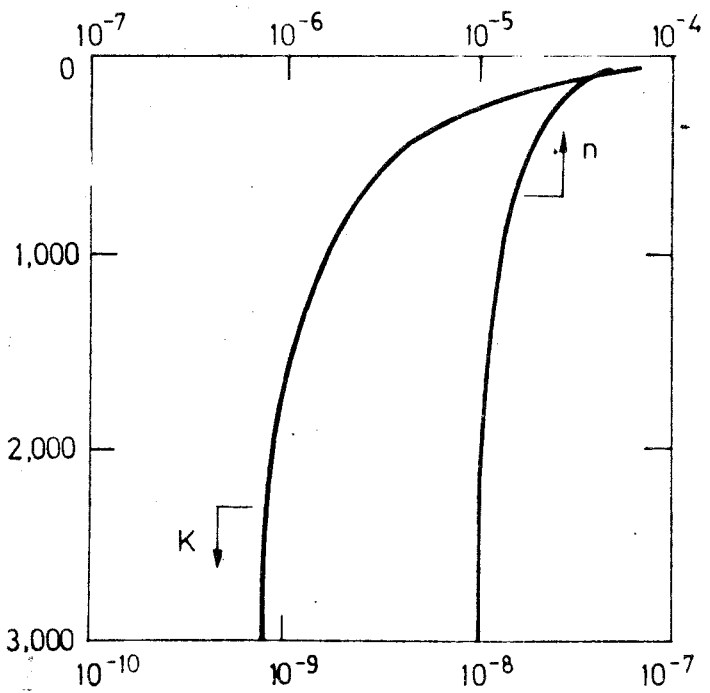
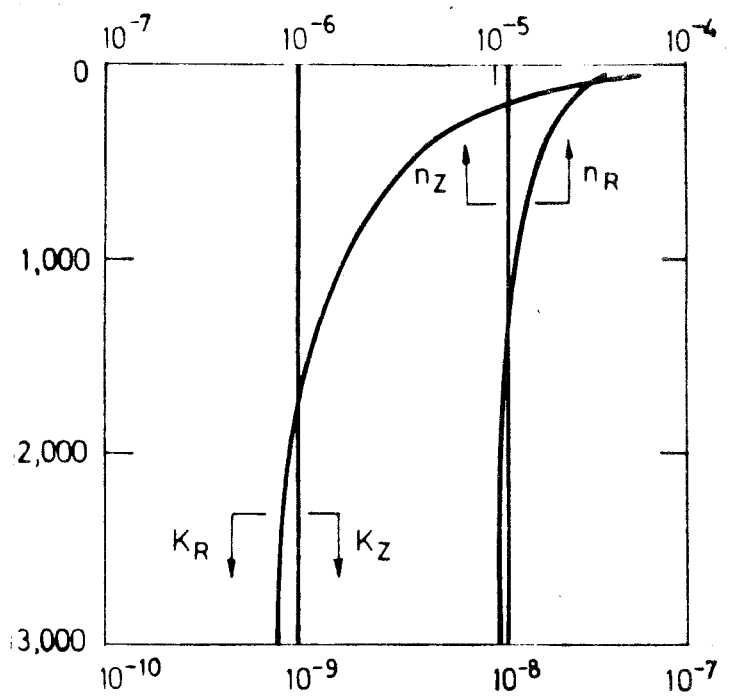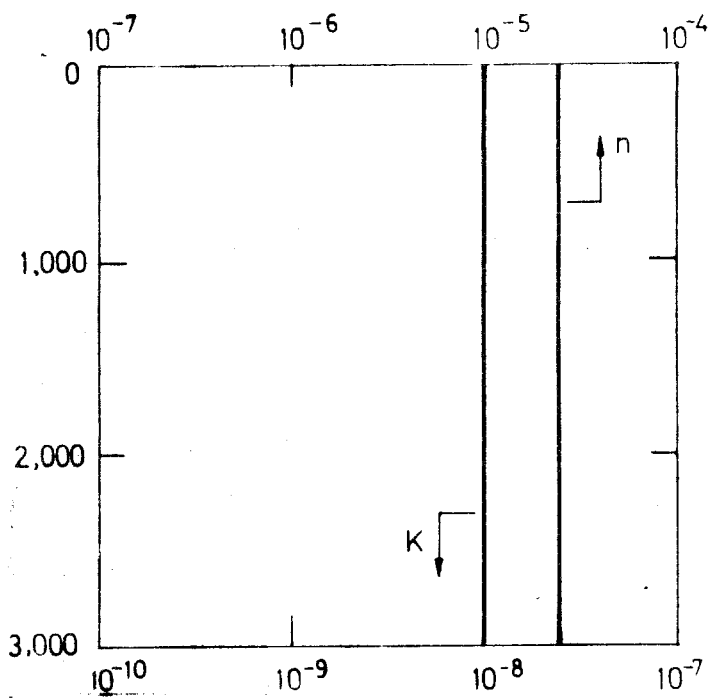FIGURE 1. FINITE ELEMENT MESHES FOR REPOSITORY DOMAIN AND LOCAL MODELS
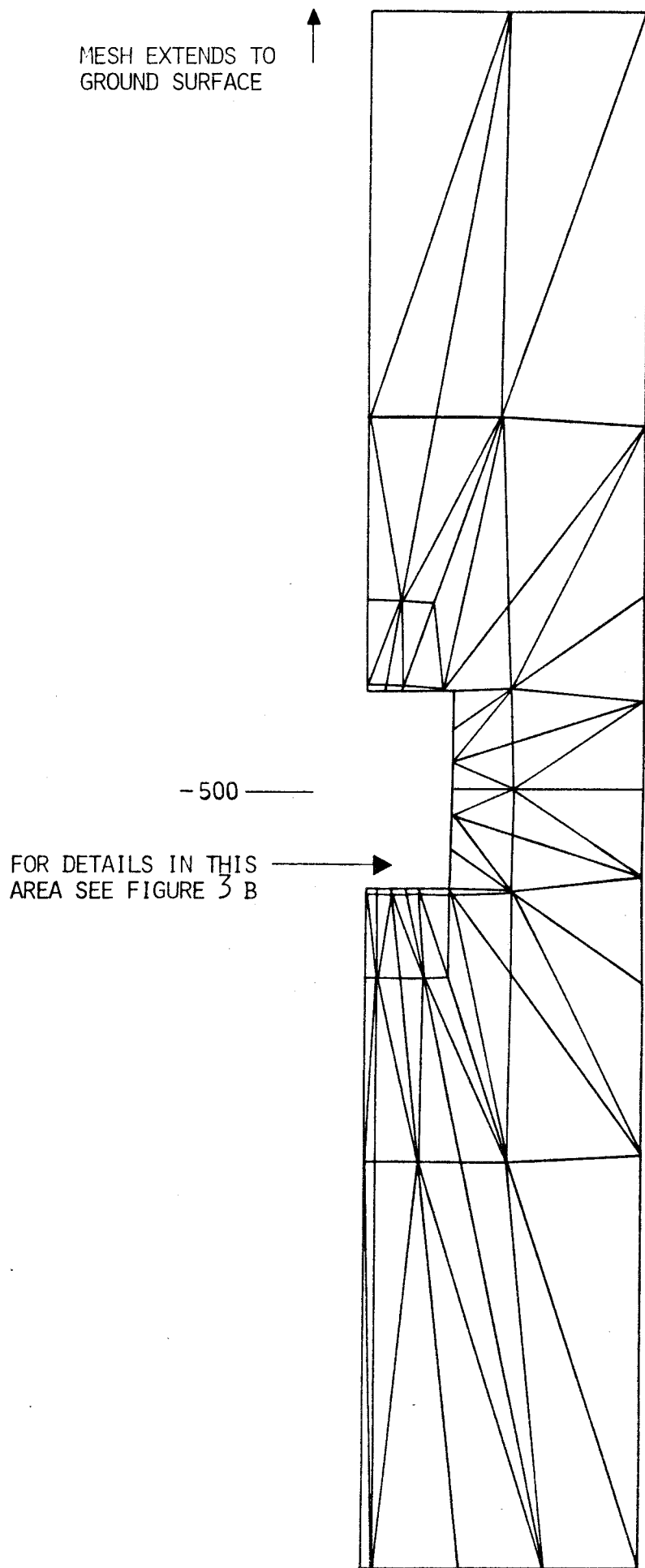
FIGURE 2. NOMINAL PERMEABILITY AND POROSITY DISTRIBUTIONS

MESH EXTENDS TO
GROUND SURFACE

−500 ———

FOR DETAILS IN THIS
AREA SEE FIGURE 3 B

FIGURE 3 A. FINITE ELEMENT MESH FOR STUDY OF INFLOW WITH K AS A FUNCTION OF STRESS.
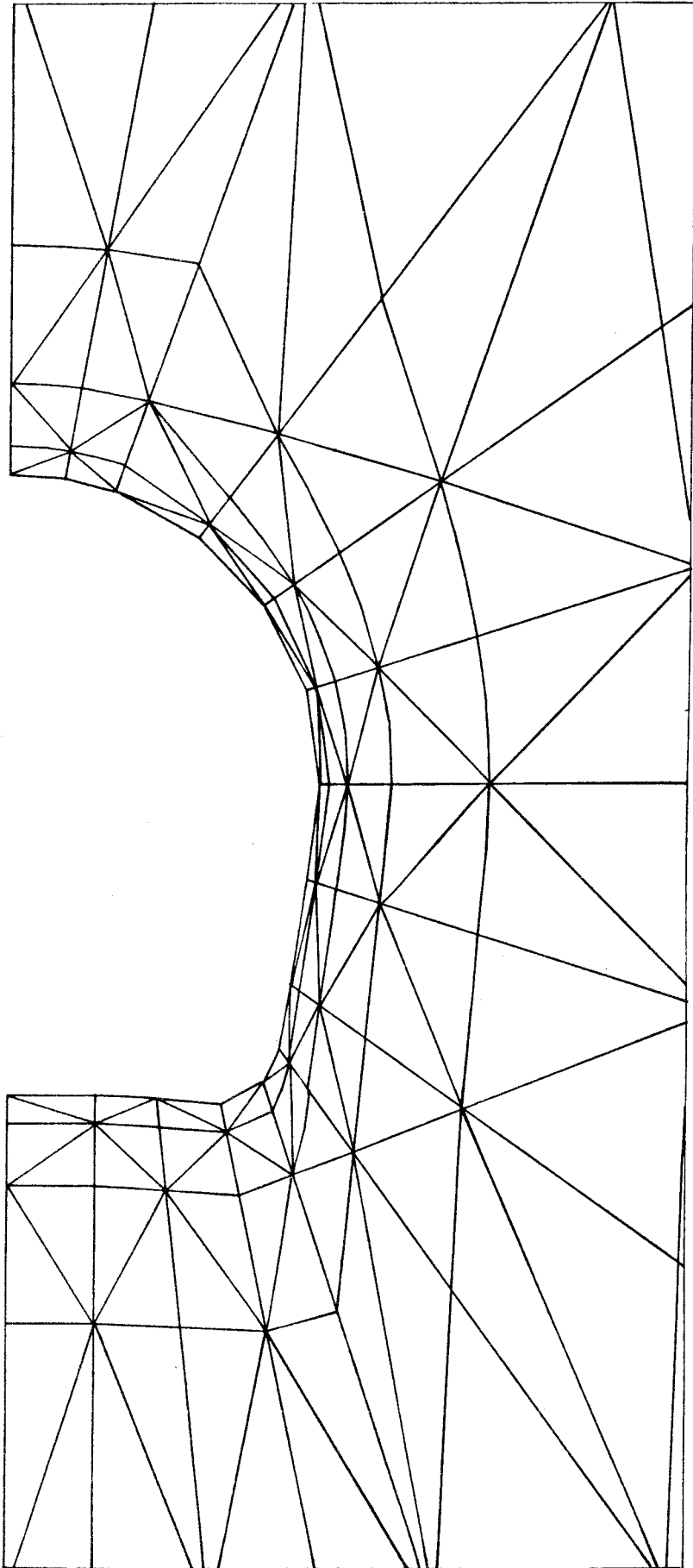
FIGURE 3 B. FINITE ELEMENT MESH FOR STUDY OF INFLOW WITH K AS A
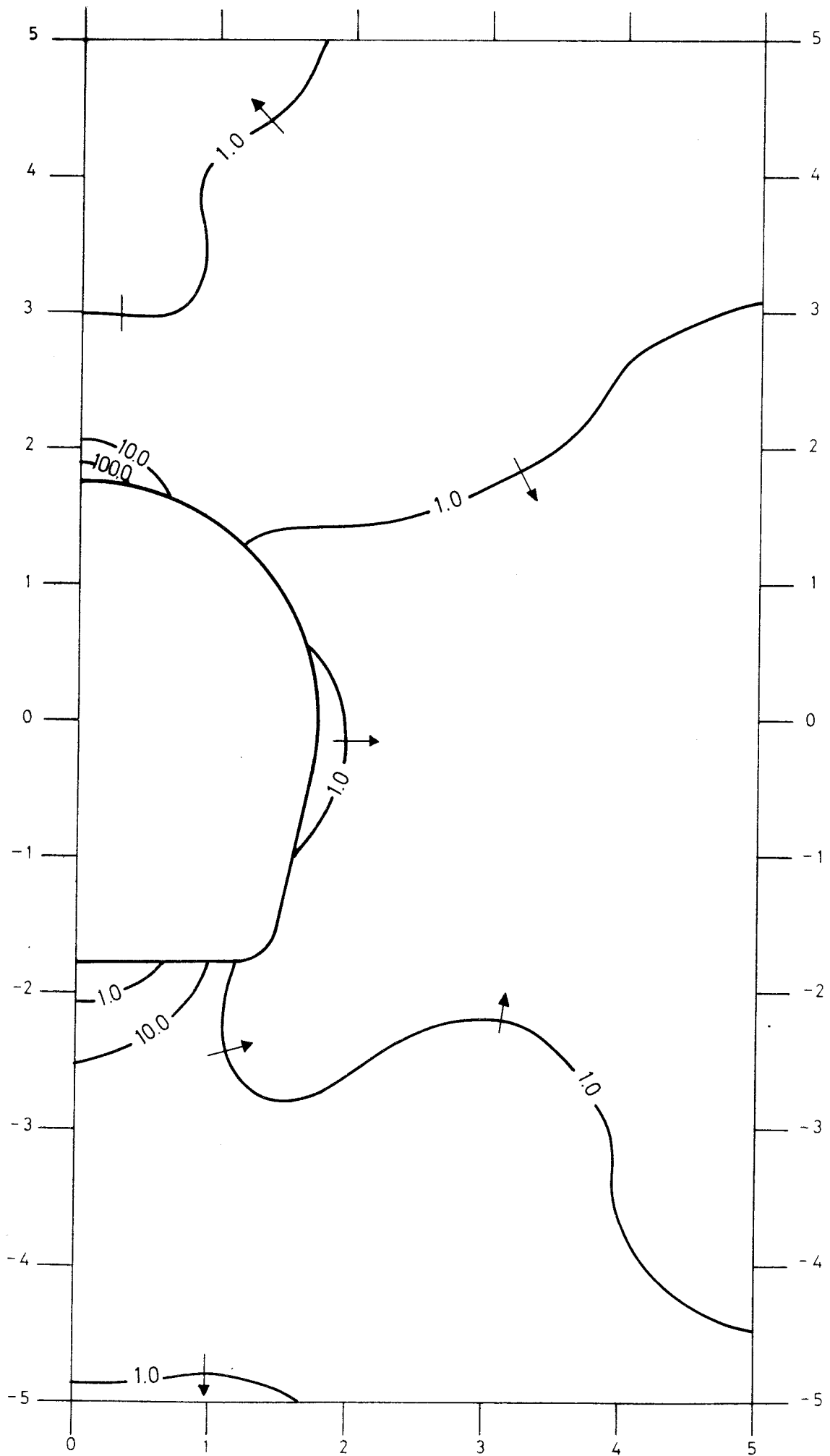FUNCTION OF STRESS.
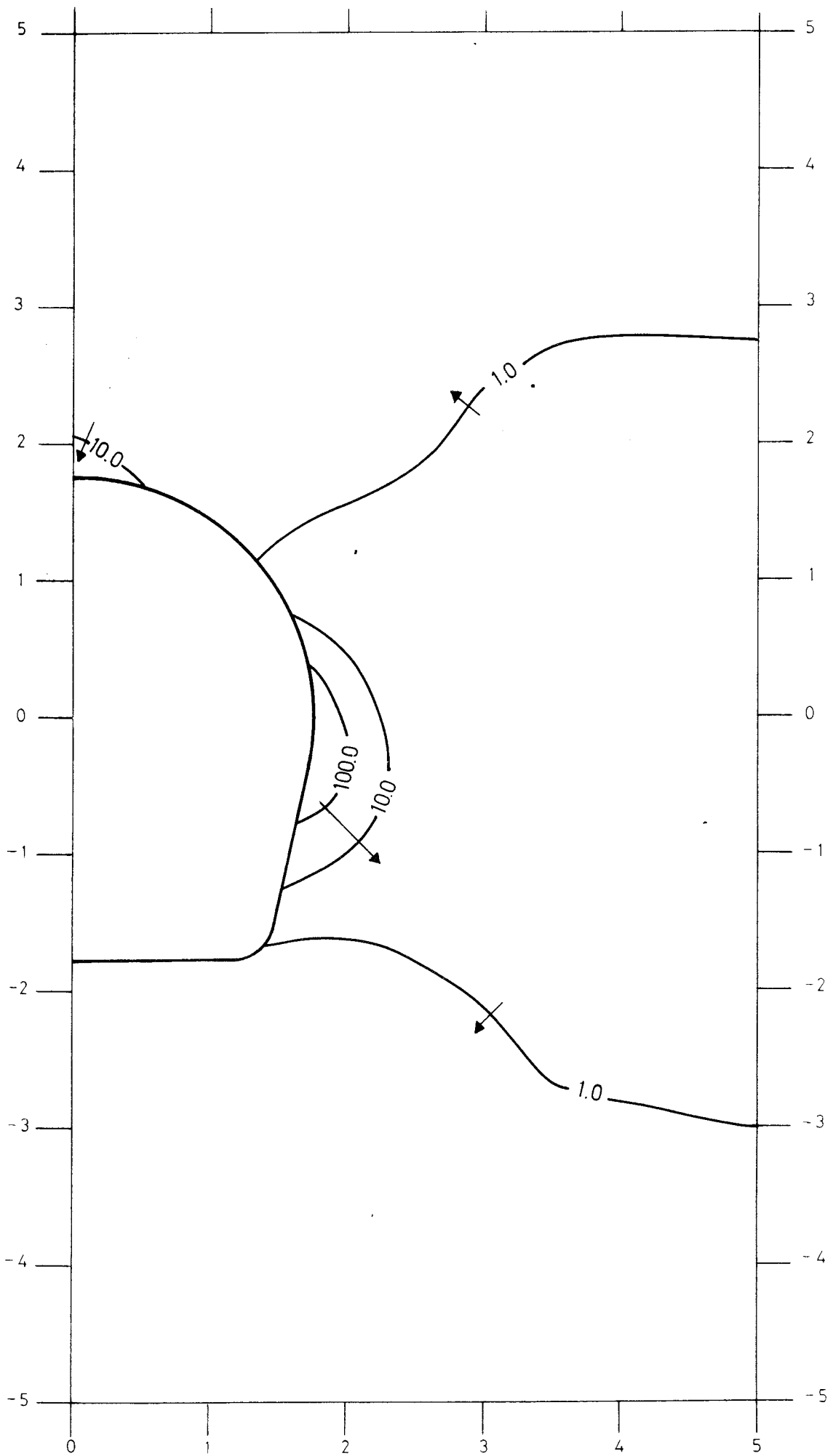
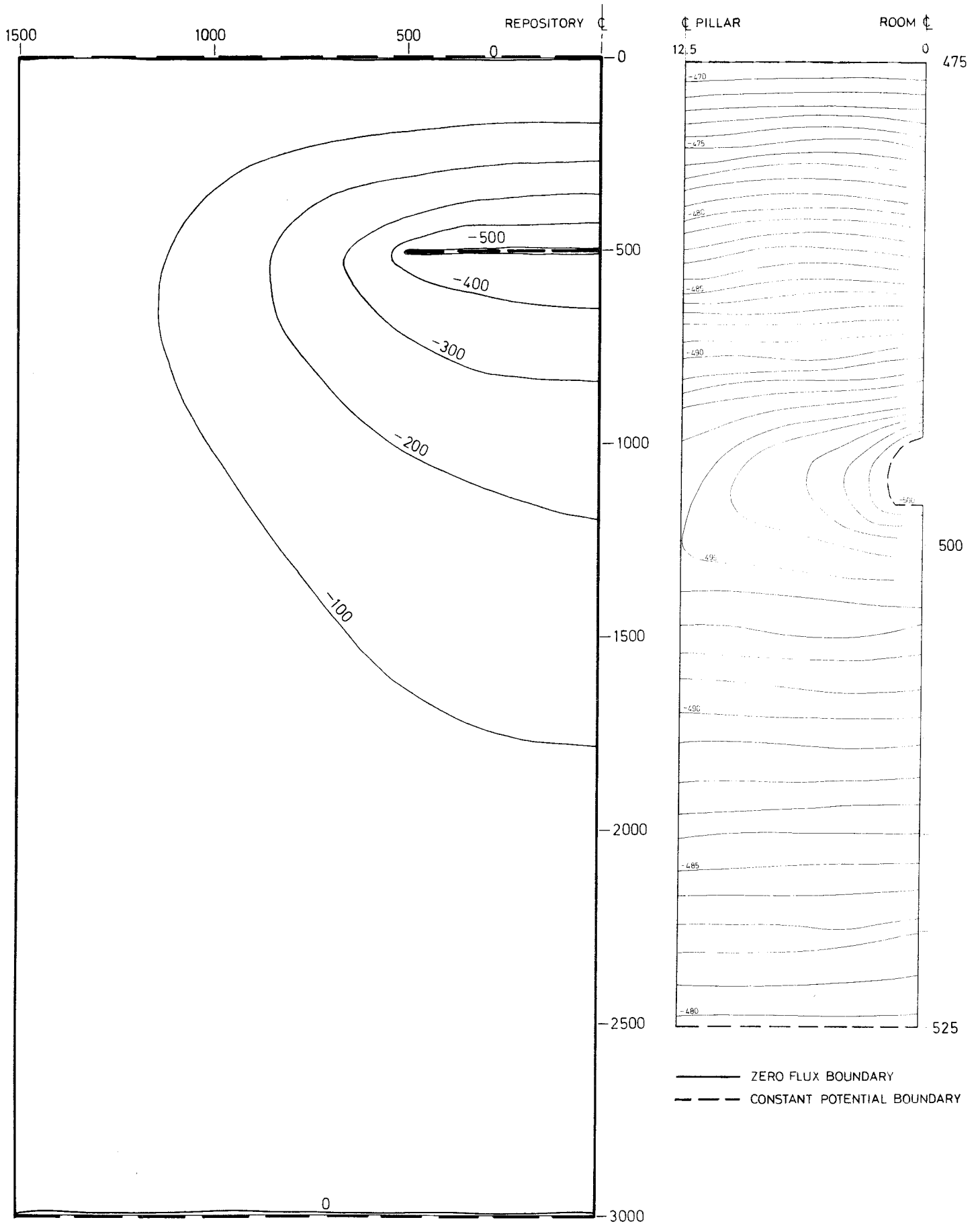FIGURE 4 A. PERMEABILITY RATIOS

FIGURE 4 B. PERMEABILITY RATIOS.

FIGURE 5. CASE 1 EQUIPOTENTIALS
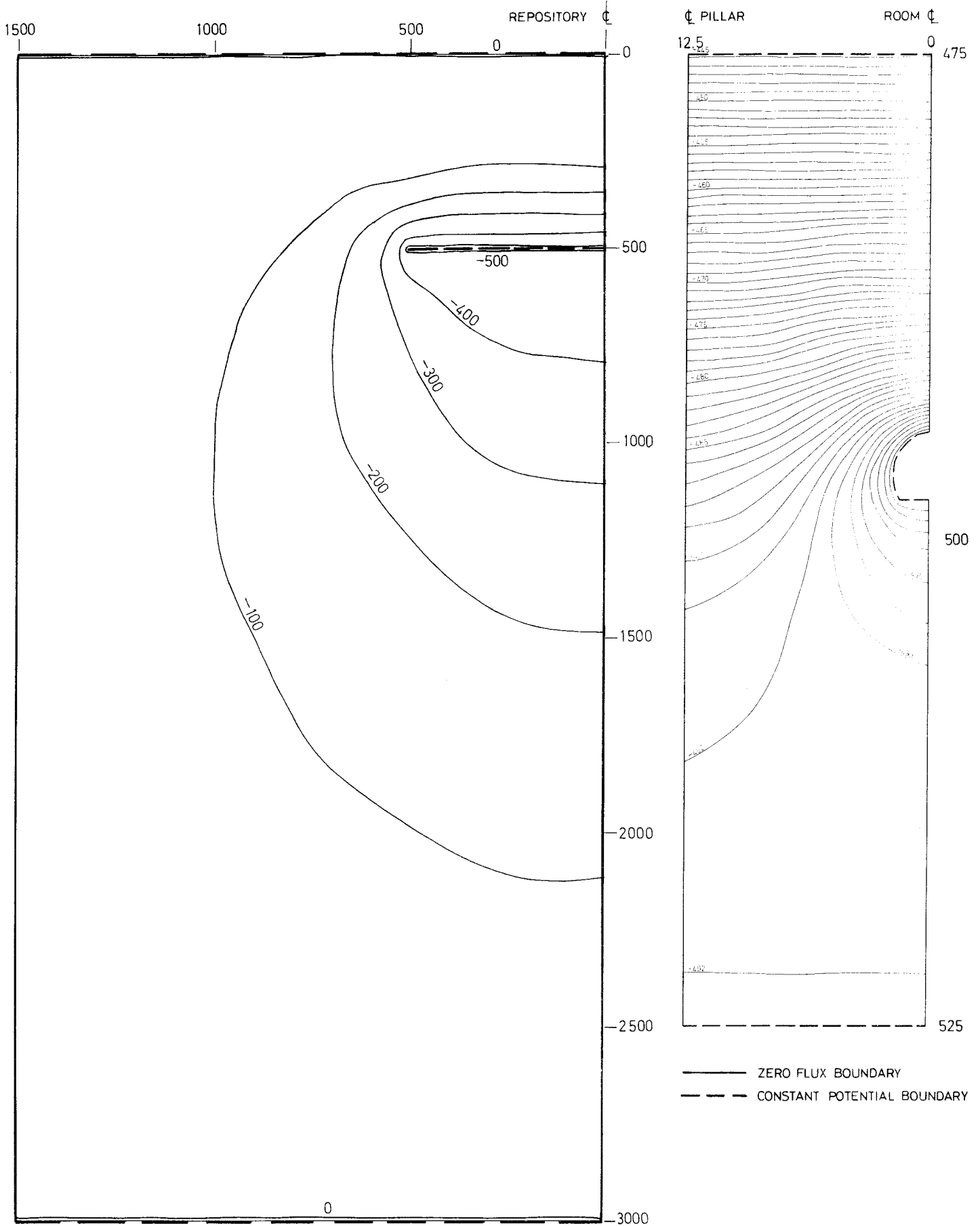
FIGURE 6. CASE 2 EQUIPOTENTIALS

ZERO FLUX BOUNDARY
CONSTANT POTENTIAL BOUNDARY
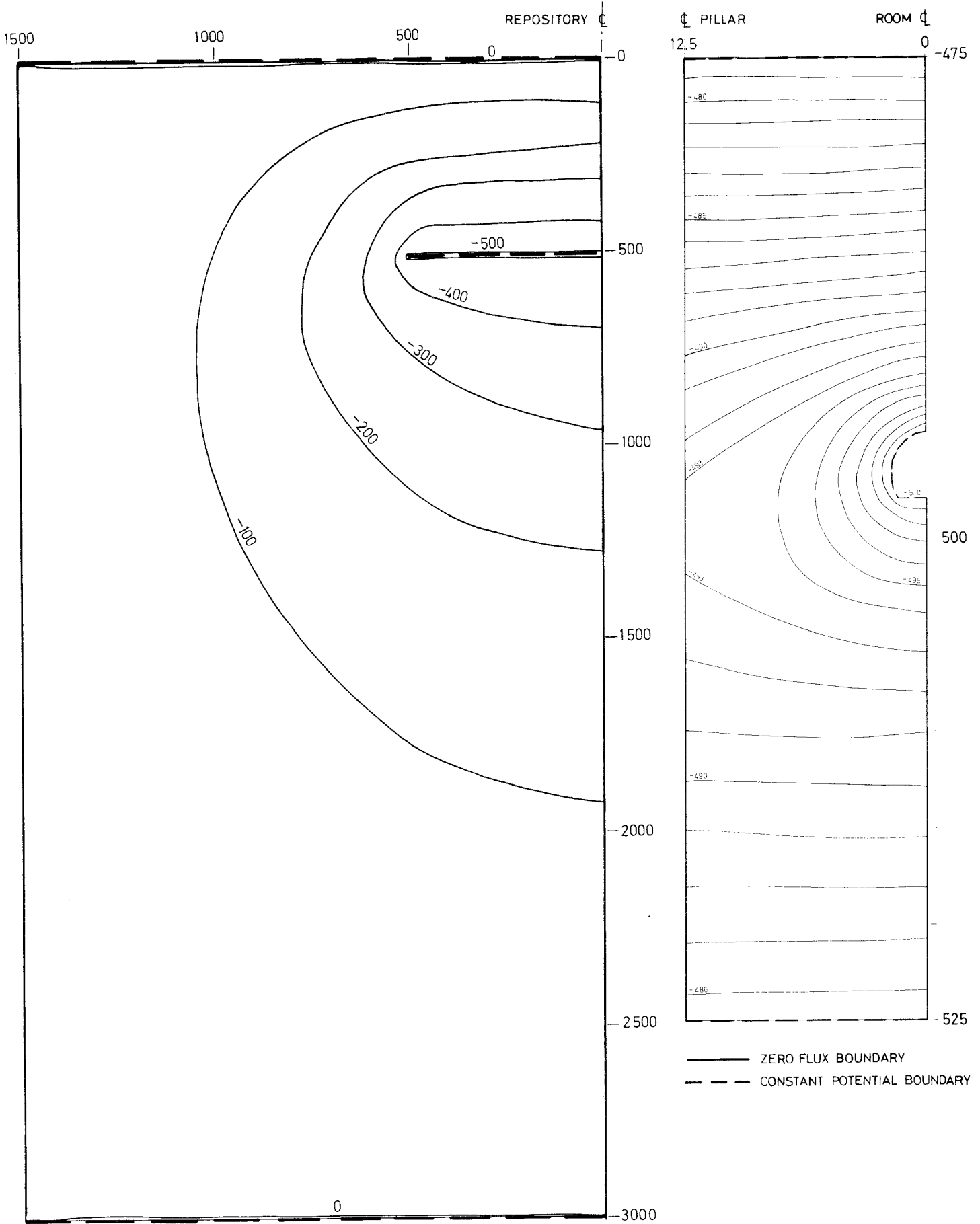
FIGURE 7. CASE 3 EQUIPOTENTIALS

FIGURE 8. CASE 4 EQUIPOTENTIALS

FIGURE 9. CASE 5 EQUIPOTENTIALS
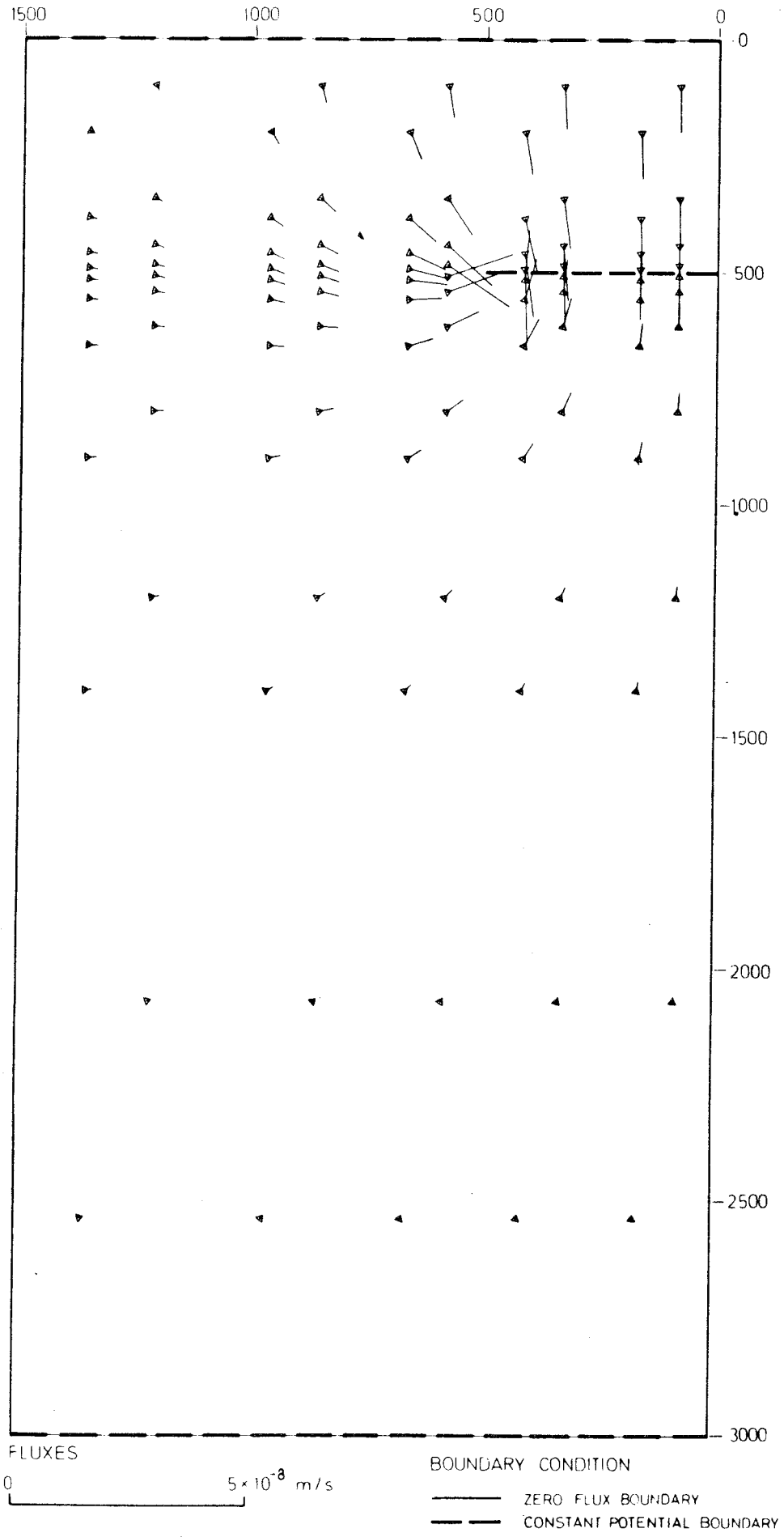
FIGURE 10. CASE 1 FLUX PATTERNS

INFLOW FL5-2 NON PERTURBED K VALUES FLUXES *E-10 M/S

100

50

20

10

NOTE CONTOUR VALUES
ARE FLUX X 10^{-10} M/S
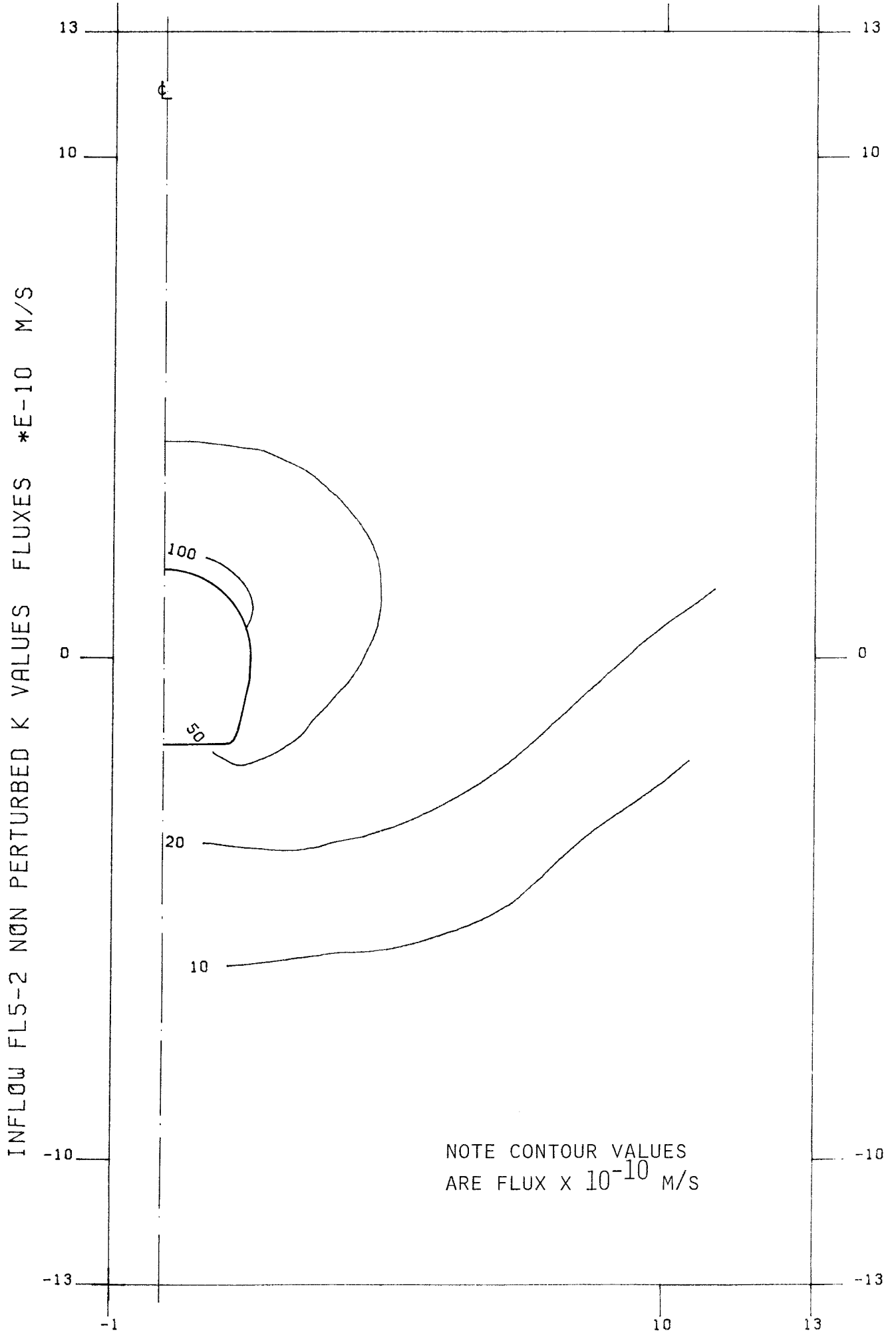
FIGURE 11. CASE 4 CONTOURS OF FLUX MAGNITUDES

INFLOW FLS-4 STRESS MODIFIED K VALUES   FLUXES *E-10 M/S
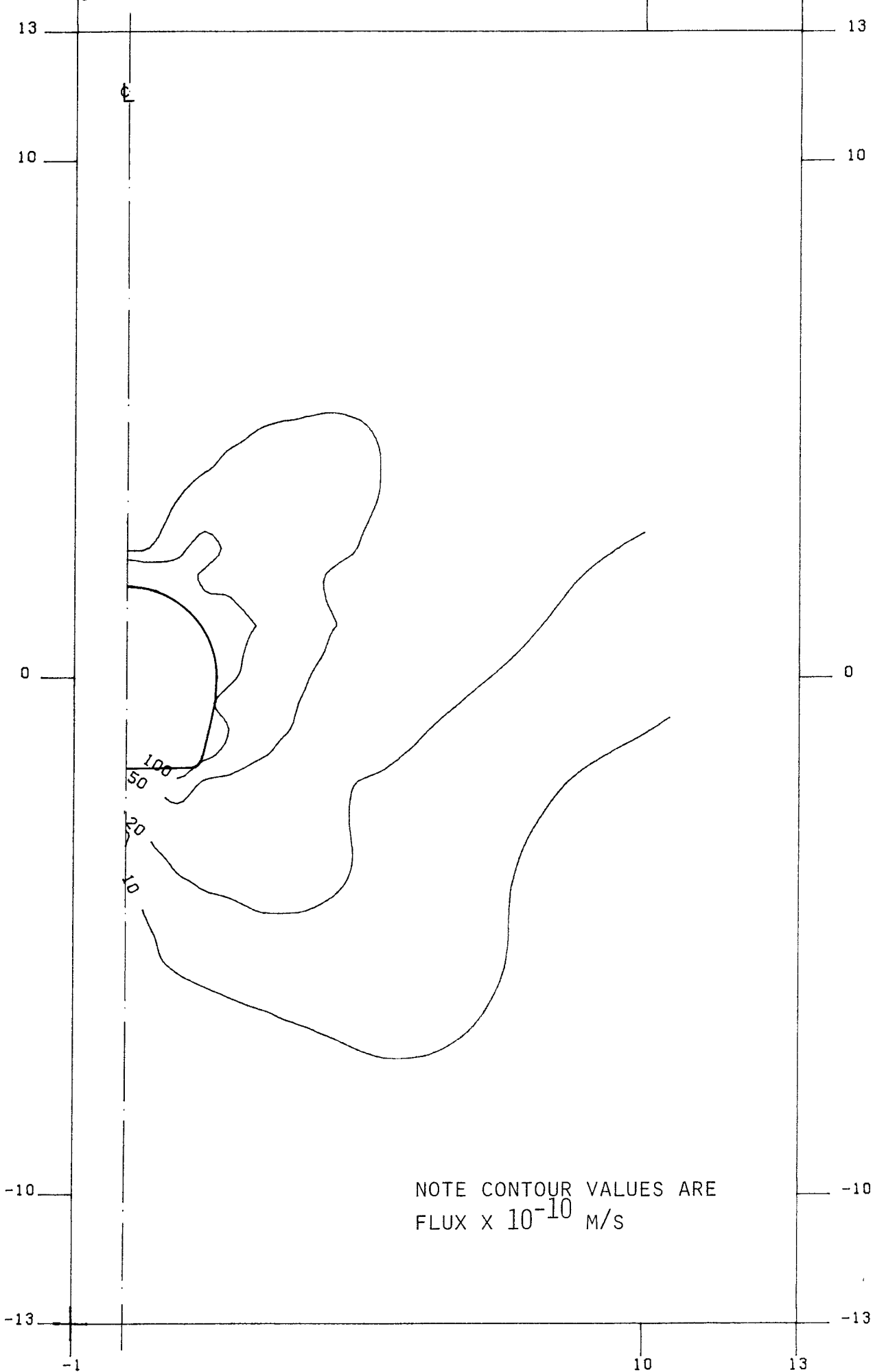
NOTE CONTOUR VALUES ARE
FLUX X 10^{-10} M/S

FIGURE 12. CASE 5 CONTOURS OF INFLOW FLUX MAGNITUDES
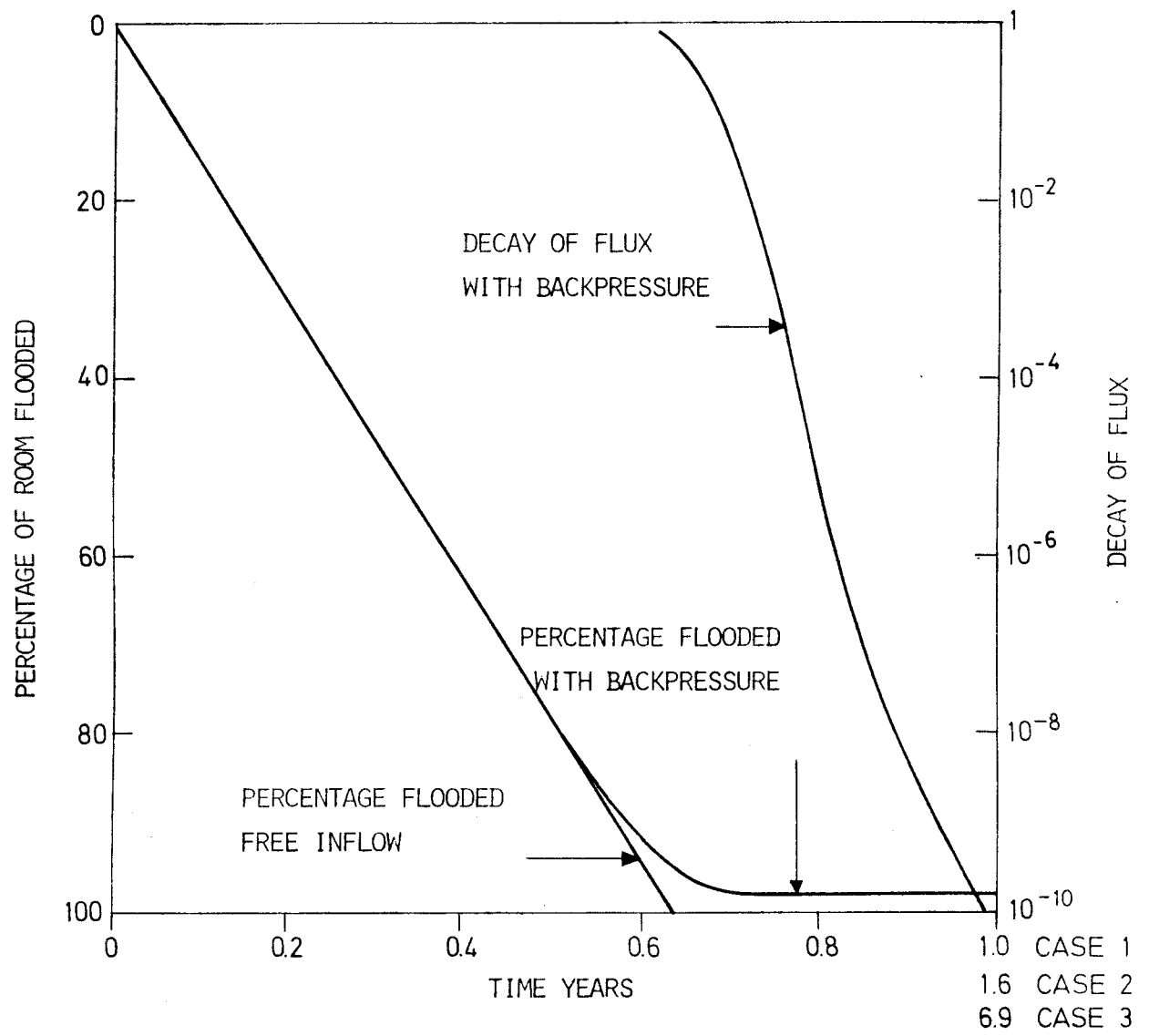
FIGURE 13. EFFECT OF BACKPRESSURE ON INFLOW
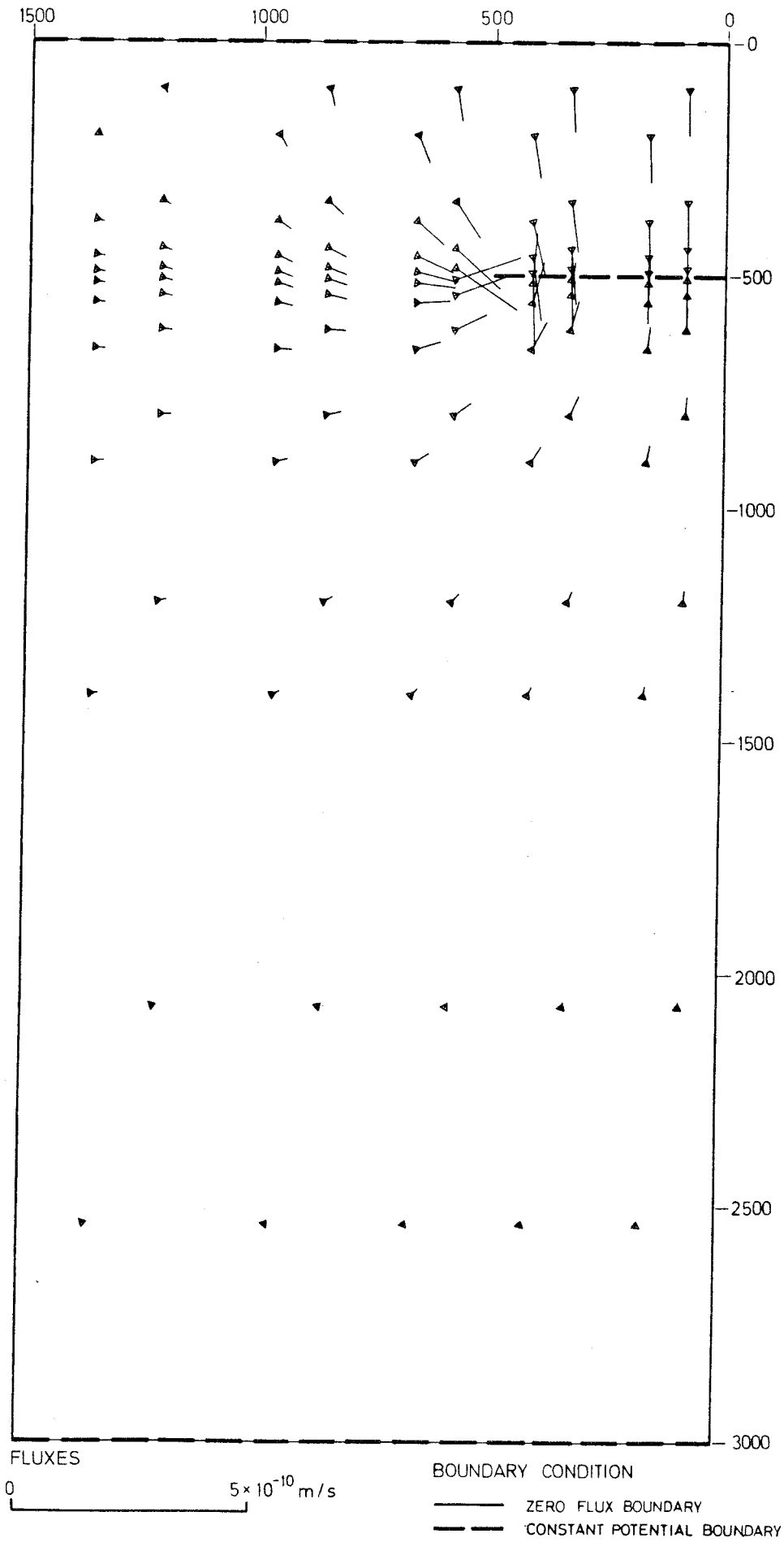
FIGURE 14. CASE 1 ADVECTIVE INFLOW RUN, SEEPAGE FLUXES

TECHNICAL REPORT 5
REPOSITORY DOMAIN GROUNDWATER
FLOW ANALYSES
PART III
THERMALLY INDUCED FLOW

KBS ⌐ Kärnbränslesäkerhet

GROUNDWATER MOVEMENTS AROUND A REPOSITORY

Phase 2. Technical Report 5: Repository Domain Groundwater Flow
        Analyses.

Part III - Thermally Induced Flow

FOREWORD

This report was prepared as one of a series of technical reports
within a study of the groundwater movements around a repository for
radioactive waste in the Precambrian bedrock of Sweden. It was written
in 3 parts, (I) Permeability Perturbations, (II) Inflow to Repository
and (III) Thermally Induced Flow. This is part III. The contract for
this study was between KBS - Kärnbränslesäkerhet (Project Fuel Safety)
and Hagconsult AB of Stockholm, Sweden. Acres Consulting Services
Limited of Niagara Falls, Ontario, Canada and RE/SPEC Incorporated of
Rapid City, South Dakota, U.S.A. acted as subconsultants to Hagconsult AB.

The principal authors of this report are Mr. Edward L. Skiba and
Dr. Robin G. Charlwood of Acres, with contributions by Dr. A.S. Burgess
of Acres and mr. J.L. Ratigan of RE/SPEC. Review was provided by
Dr. Ulf E. Lindblom of Hagconsult AB.

The opinions and conclusions in this document are those of the authors and
should not be interpreted as necessarily representing the official policies
or recommendations of KBS.

Stockholm, October 1977

Ulf E. Lindblom
Study Director
Hagconsult AB

TABLE OF CONTENTS

1.   INTRODUCTION

The overall objective of this study is to assess the groundwater flow
around a radioactive waste repository as produced by various possible
mechanisms. The pre-excavation flow due to the regional gradient is
the subject of a previous report in this series (1)[x)] This report deals
with the time period following loading of the repository, in which
the mechanism of natural convection may produce flows which are tempo-
rary perturbations to the regional flow. The natural convection ends
when the waste-generated temperature perturbation disappears and the
regional geothermal gradient is reestablished.

Inflow into the repository will occur during the construction and
emplacement phases and will continue until the repository zone is
recharged. Inflow,which is discussed seperately (2), is shown to be
much stronger than the thermally induced flows and consequently thermally
induced flows are not significant until the recharge period is over.
The interaction between these mechanisms is discussed in this report.
Following these time periods of transient perturbations,
regional flow again dominates water transport, albeit somewhat in-
fluenced by the residual effects of the repository tunnelling and
associated excavation and fracturing of the rock. This long-
term era flow is the subject of another report in this series (3).

The specific objective of this report is to define flow patterns
during the thermal era, using baseline properties, with the in-
tention of identifying groundwater flow fields which may be of interest
in subsequent safety analyses. These flows and temperature fields are
computed with full coupling between heat transfer and ground-water flow and
are refered to as thermally induced flow analyses. Conduction heat trans-
fer results are also presented for comparison with the thermally induced
flow results and the magnitude of advective heat transfer is discussed.

x)   Numbers in paranthesis refer to references at end of text

After a brief presentation of the assumptions, geometries and material properties used in this short-term study, the bulk of this report is devoted to thermally induced flow and coupled thermally induced and cross flow. Conclusions are presented section by section and summarized later. The appendices contain the development of the theory for the thermally induced flow model and validation runs.

2.   INPUT PARAMETERS AND MODELING

2.1   Layout and Thermal Loading

The repository layout consists of 41 tunnels each 1 km long on 25 m centers, covering an area 1 km square at a depth of 500 m (4). For the study of thermally induced flow, nine thousand canisters of 40 year old highlevel waste (HLW), reprocessed at 10 years, were assumed to be instantaneously emplaced at 4 m spacing along the tunnels. This results in a gross thermal loading (GTL) of 5.25 W/m^2 (525 W/canister divided by 100 m^2/canister), at time of emplacement. The decay of this gross thermal loading is presented in Figure 1 for 1,000 years. After 1,000 years, the thermal loading is almost negligible since it falls to 0.0003 W/m^2 at 10,000 years which is considerably less than the approximate geothermal heat flux of 0.04 W/m^2.

Nonetheless, the theoretical potential for thermally induced flows will remain as long as temperature gradients exist around the repository. The geothermal gradient is not represented in the modeling. Instead, an initial condition of zero temperature everywhere is assumed. The pre-emplacement temperatures in the conduction report were $5^{o}C$ at surface and a $20^{o}C$/km rise with depth giving $15^{o}C$ at the repository depth of 500 m. Since the isotherms due to the geothermal loading are nominally horizontal, they do not directly cause thermally induced flow (5). Horizontal inhomogeneity of the geothermal gradient may be found in thermally active regions but is not considered here as a generator of flow.

The conceptual model considered consists of a large block of "intact" rock bounded by singular features. The intact rock is assumed to be traversed by joint sets and thus permeable and is the subject of these analyses. Flow mechanisms in the singular features are not studied in this report.

2.2   Flow Assumptions

Initially, we assume saturated, one-phase flow with fixed fluid and
permeability parameters. Laminar Darcy flow is assumed valid for an
equivalent porous analogue to the jointed medium. The regional ground-
water flow report (1) in this series describes the details of the deri-
vation of the three permeability and porosity distributions presented
in Figure 2. They are:

Case 1) $K = 10^{-8}$m/s, homogeneous and isotropic

Case 2) $K(Z)$, isotropic, non-homogeneous, decreasing with depth.

Case 3) $K_R(Z)$, anisotropic, non-homogeneous horizontal, decreasing with
        depth, but K_Z, vertical, constant at 10^{-9}m/s

These are used to determine the volumetric flux of fluid, and in con-
junction with their corresponding porosities, the fluid velocity. The
porosities are computed from the permeabilities using a formulation de-
rived by Snow (6) for parallel fracture sets at given spacing of 1.8 m.
Flow patterns derived using any of the above permeability distributions
are also valid for other conditions providing the same relative per-
meability distribution in space is maintained. Fluxes and velocities may
then be scaled homogeneously with the new permeability values. Flux is
scaled linearly with the K scale for given fracture spacing and veloci-
ties scale as $K^{2/3}$. Thus, a decrease by a factor of 1,000 in K decreases
fluxes by 1,000 but velocities only by 100 for the same fracture spacing.

The other material properties used in this report are shown in Figure 3.

2.3   Flow Geometries and Boundary Conditions

The modeling has been divided into two scales for examining global and
local effects. These two model geometries are found in Figure 4. The
two dimensional global model has been constructed with symmetry at the
repository centreline, making the centre boundary impermeable to flow
of heat and fluid. For thermally induced flow modeling the other three
boundaries are also impermeable to fluid flow but are maintained at 0°C,
the initial condition for temperature. The model size is felt to be large
enough to preclude interference in thermal distributions to at least 1,000
years. Beyond this time, growing interference will occur. However, since

near repository temperatures are decreasing anyway and the loading is
negligible, the interference is estimated to be tolerable. The lateral
dimension and depth are also felt to be sufficient to allow ample room
for return flow to complement the thermally induced flow through the repo-
sitory itself, and therefore, to be non-restrictive. The global model
size is representative of the expected intact zone. Any larger model
would have to include singular features, which are considered beyond
the scope of the present study and the state-of-the-art for thermally
induced modeling.

The local model is a half room and pillar section with symmetry boundaries
on both vertical sides, impervious to heat or fluid flow. In all local
model runs, the time dependent boundary conditions on potential and
temperature, required for the top and bottom, are obtained from the
corresponding global model run. The room is assumed to be backfilled
with a saturated impervious material.

For the combination of thermally induced flow and global crossflow con-
sidered later in this report, the thermally induced flow results were
transposed across the symmetry plane and the crossflow applied by super-
position to produce the flow patterns.

3.   THERMALLY INDUCED FLOW

3.1   Conduction Baseline Results

The isotherms predicted with the conduction global (GLOBAL/HT-1) and local
(LOCAL/HT-1) model runs are contained in Figures 5 to 7 at 40, 200 and
1,000 years after emplacement. The isotherms are not plotted for times
greater than 1000 years since there is evidence of boundary interference
with the isotherms. The repository temperature estimates at long times,
however, seem quite good compared with the analyses presented in the
thermal analyses report (4). Figure 8 shows two important temperature
histories with time, the average repository temperature and the drill
hole periphery temperature. These compare well with the temperatrues
attained in the conduction heat transfer report. The only essential
difference in these runs is the backfill in the tunnel of the local model,
but this appears to have little influence on the resulting temperatures.
The runs to 1,000 years were carried out with the same time steps as in
the conduction heat transfer report, and repeated with larger time steps
(3 times larger). The results were comparable to within $0.5^{\circ}C$ at all
locations at all times. The longer time steps were used for subsequent
analyses.

3.2   Thermally Induced Flow Results

In order to assess the relative importance of thermally induced flow in
comparison with regional cross flow, the models were analysed with and
without regional cross flow. All thermally induced flow analyses for
zero cross flow showed no significant deviation from the conduction tempera-
tures discussed previously. All temperatures were within $0.1^{\circ}C$ of the con-
duction temperatures at all times, thus demonstrating that advective heat
transport is negligible as far as temperature predictions are concerned.

Velocity plots are contained in Figures 9 to 13 for the GLOBAL/HT-2 and
LOCAL/HT-2 runs using the Case 1 permeability, K homogeneous and isotropic
at 10^{-8}m/s, at 40, 200, 1,000, 10,000 and 100,000 years after emplacement.
Similar results are presented in Figures 14 to 18 for the GLOBAL/HT-3 and

LOCAL/HT-3 runs using the Case 3 permeabilities, K(Z) horizontally, and K uniform vertically. For the global thermally induced flow runs, Figure 29 displays the average vertical transport velocities above the repository. Local velocities are larger near the repository. Global runs were attempted for the Case 2 permeabilities, K(Z) isotropic, but the vertical permeability was so large near the surface that numerical instability developed at the large time steps used in the runs and the results were invalidated. All the results show a mild degree of numerical instability at low times (40 and 200 years for Case 1 and 40 years for Case 3) which damps out for longer times and yields results which are stable. The runs to 1,000 years were performed with one iteration per time step, but the run extensions to 100,000 years used two iterations per time step, allowing the use of large steps while preserving stability. Since heat advection is so low in this problem, it would seem adviseable to neglect it in further runs, which would increase the stability of the buoyant flow produced. This numerical stability, or instability, is not to be confused with the physical instability of vertically stratified systems (Bénard cells). Since a horizontal gradient of temperature and density always exists in our system, there is no minimum temperature difference which must be exceeded in the classical sense for thermally induced flow to commence (5). Other inhibiting mechanisms, due to the fractured nature of our medium or partial saturation, may exist, but are not obvious and have not been considered here. Experimental and numerical comparisons of fractured and analogous porous media undergoing thermal flow loading must be conducted to answer the many questions which may arise on this matter.

The nature of the flows themselves warrant some comment. Globally, a cell, rising through the repository and descending around it, develops at short times and persists, only dropping an order of magnitude over 100,000 years. A local cell also develops beside the waste canister and tunnel but it begins to weaken substantially by 1,000 years. The inherent problem with these flows is that the local cell channels flow over the canister itself and the global cell provides direct transport to the surface. For the nominal conductivities used herein the expected travel times from the repository to the surface are presented below. It must be noted that these are for the conditions of no regional crossflow. This condition is likely only beneath a large body of water well away from the shore.

Table 2.

Thermally induced flow - travel time to surface assuming no cross flow

Travel time	starting at 1,000 years	starting at 100,000 years
Case 1 ($K_z = 10^{-8}$) isotropic homogeneous	200 years	2,000 years
Case 3 ($K_z = 10^{-9}$) anisotropic non-homogeneous	1,000 years	10,000 years

If these permeabilities are decreased by 3 orders of magnitude, the times increase by a factor of 100 assuming a similar fracture frequency. These decreased permeabilities are quite limiting to the flow. Note that the increased horizontal permeability of Case 3 near the surface appears to have little influence on the travel time to the surface. The expected increase in travel time for a one order decrease in K is $10^{2/3} = 4.64$, close to that found numerically.

An interesting effect of the anisotropy and non-homogeneity is noted in the location of the eye of the global cell. For both cases, the eye moves outward from the repository at long times. In Case 1 (isotropic), the eye drops below the repository several hundreds of meters. The increased horizontal permeability of the anisotropic Case 3, near the surface, appears to permit the eye to remain at repository level at long times, in contrast to the isotropic case.

3.3   Effect of Inflow on Thermally Induced Flow

Inflow is expected to be capable of dominating hydrothermal flow for a considerable period of time. This could be evaluated by comparing downward inflow velocities (2) and upward thermally induced flow velocities directly above the global repository level, see below.

Relative strenghts of inflow and thermally induced flow

	Flux (m/s)	Velocity (m/s)
Case 1 Inflow	1.213×10^{-8}	4.89×10^{-4}
Case 1 T/I flow	1.463×10^{-11}	5.90×10^{-7} (100 years)
Case 3 Inflow	1.77×10^{-9}	1.08×10^{-4}
Case 3 T/I flow	2.165×10^{-12}	1.32×10^{-7} (100 years)

According to the table, the inflows are stronger than the thermally induced flows by a factor of 830 for Case 1 and 820 for Case 3. In both cases, inflow and thermally induced flows are equally controlled by the vertical hydraulic permeability and scale equally as K is scaled in value. It is anticipated that for any K and n distributions, the ratio of inflow to thermally induced flow will be near to 800 or roughly 3 orders of magnitude. These results are for free inflow and initial pressurizing inflow as well. On the tail of the pressurizing inflow will still dominate hydrothermal flow for a period roughly equal to its free inflow time without pressurizing. This domination time could extend to hundreds of years (640 for $K_Z = 10^{-11}$ m/s), but the value of K required for domination of inflow to 100,000 years would be incomprehensibly low (0.64×10^{-13} m/s). This result would indicate that thermally induced flow could be expected to occur, despite inflow, at some time during the life of the repository.

4.   THERMALLY INDUCED FLOW

Regional gradient flow and thermally induced flow can only be expected to
interact significantly if the two flows are of comparable magnitude. Figures 19
to 23 present the interaction for the Case 1 K distribution with a regional
gradient of $2 \cdot 10^{-3}$ (1). Since the hydrothermal and cross flow are essentially
non-advective of heat, superposition of solutions was used to produce these
plots. At 1,000 years, the cross flow is about one order of magnitude stronger
than the average thermally induced flow; however, at 100,000 years, this in-
creases to two orders of magnitude. The time for flow due to cross flow from
the edge of the repository to the edge of the geometry (1,000 m) is about
40 years. Thus, it is evident that the cross flow, while it cannot completely
suppress the hydrothermal flow, sweeps the flow horizontally and may reduce
or eliminate the possibility of direct surface discharge of contaminants due
to vertical hydrothermal flow, even for small regional gradients.

Figures 24 to 28 present similar plots for the Case 3 anisotropic K distri-
bution in which the horizontal permeability is greater than the vertical. In
fact, it is evident that the regional flow dominates even more here since it
is determined by the horizontal permeability, not the vertical. Time to exit
at the vertical boundary is 70 years since the regional gradient velocity is
about 0.45×10^{-6} m/s as compared to 0.8×10^{-6} m/s for Case 1 at the repository.
The velocities may be scaled by $K^{2/3}$ and the times by $K^{-2/3}$ if scaling of K
is desired. This will not effect the relative strenghts of the regional and
thermally induced flows, however, if the same isotropy or distribution of
anisotropy with depth is maintained during scaling.

Quantity of water flowing past canister is = (flux x cross section area of
canister perpendicular to flux). The highest quantity will be for horizontal
flow. Flow through pillars unimportant for leaching. Either needs more explana-
tion or should be deleted.

5. SUMMARY AND CONCLUSIONS

Thermally induced flows have been predicted to occur using a fully coupled thermal/flow finite-element program. It is assumed that flow in the host rock in the region of the repository can be modelled by Darcy's law with a bouyant term under fully saturated conditions.

Thermally induced flow does not cause significant advective heat transfer for expected values of rock mass permeabilities. In the future thermally induced flow analyses may be done with bouyant coupling only using temperature fields computed on the basis of conduction only if the gross thermal loading is not significantly increased. This will enchance analytical stability and output accuracy.

If regional flow is not present, natural convection cells will develop. There will be global cells which could conceivably transport contaminants to the surface. Figure 29 displays the calculated average vertical transport velocities above the repository assuming no cross-flow. The predicted travel times are in the range of 1,000 years depending on the rock mass permeability. Local cells may also develop which will have the effect of channelling the flow close to the waste cannisters.

The theoretical model indicates that in the absence of cross flow natural convection cells can be initiated with small temperature gradients. The concept of a critical gradient for flow initiation does not apply in this problem. If partially saturated conditions exist, these may inhibit thermal flows. The fully saturated case analysed here is thought to give higher velocities and hence conservative results for safety analyses.

The inflow period may last for many years, depending on the nature of flow around and into the repository. While inflow lasts, it is expected that the inflow velocities will be 2 to 3 orders of magnitude larger than the velocities caused by natural convection alone. Consequently, inflow would prevent thermally induced flows from moving upwards to the surface during the recharge period. Natural convection flows can, however, start once inflow is essentially complete.

It is expected that a regional groundwater gradient of $2 \cdot 10^{-3}$ will exist which will cause the convection flows to be swept almost horizontally, indicating that the most likely point of exit from the host rock is into a singular feature at depth and not up to the surface above the repository.

The quantity of groundwater flowing through the repository due to regional gradients is increased by the thermally induced flow effects. These flow quantities may affect leach rate predictions depending on whether it is determined by concentration in the water or by the diffusion rate in the glass matrix.The local thermally induced flows may tend to channel flow near the canisters, although this tendency may be somewhat reduced by the regional flow.

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APPENDICES

A.1 INTRODUCTION

This appendix describes FINI 520, a two-dimensional, planar and axi-
symmetric, transient (or steady-state) finite element saturated thermally
induced flow code developed at Acres Consulting Services Limited,
Niagara Falls, Ontario, Canada, in 1977.

The available elements are an anisotropic isoparametric quadratic
head/temperature triangle with 6 nodes (12 degrees of freedom) and a
line loading element with 3 nodes (6 degrees of freedom).

Matrices and vectors are computed by numerical quadrature. Conductivity
Material properties are principal values of anisotropic thermal and
flow permeability and principal angles, bulk density and bulk specific
heat, fluid density and fluid specific heat, thermal expansion coefficient
and porosity.

The model starts from a specified set of initial conditions and
proceeds stepwise through a number of solutions during the course
of a transient run. Several iterations must be performed to get
steady-state solutions.

Boundary conditions include specified head/temperature at nodes,
nodal loads (fluid/heat source), element internal loads and surface
loads. The surface loads are specified normal seepage velocity/heat
flux and specified head/temperature outside a film. All specified
values and loads are variable linearly over time spans of one to
several time steps and may be started and stopped independently
whenever desired.

Pore velocities are computed in a post-processor attached to the
main program itself and plotting of meshes, isotherms, equipotentials
and pore velocities is accomplished by a follow-up program, FINI 511,
that accesses several of the disc files produced by the main code.

The program is divided into several overlay segments to minimize
core storage requirements. Within each overlay the coding is broken
into subroutines containing logical segments. In addition, extensive

A.2 FORMULATION AND DEVELOPMENT

The basic equations for porous thermally induced flow are the motion equation for fluid flow and the heat transport equation, as discussed separately below. Later in this section, they are combined to form the system matrix equations and boundary condition compatibility is described.

A.2.1 Groundwater Flow

The equation of motion for the liquid phase in a nonisothermal, single-component, groundwater flow system may be written as:

$$\frac{\partial}{\partial x_p} \left(\frac{\rho g k_{pq}}{\mu} \left(\frac{\partial P}{\partial x_q} + \rho g \delta_{q1} \right) \right) - \rho g \left(n\beta^P + \alpha^P \right) \frac{\partial P}{\partial t}$$

$$- n\rho g \beta^T \frac{\partial T}{\partial t} + \rho g Q^\phi = 0 \tag{1}$$

where x = linear dimension,

ρ = fluid density,

g = gravitational constant,

k = intrinsic permeability tensor,

μ = dynamic viscosity,

P = pressure,

δ = Kronecker delta and direction 1 is upward,

n = porosity,

β^P = fluid compressibility,

α^P = matrix compressibility,

t = time,

β^T = fluid thermal expansion coefficient (negative),

T = temperature, and

Q^ϕ = volumetric rate of fluid source.

The equation of state of the fluid is

$$\rho = \rho_0 + \rho_0 \beta^T \left(T - T_0 \right) + \rho_0 \beta^P \left(P - P_0 \right) \tag{2}$$

where the subscript o denotes evaluation at the reference state.

We use the Boussinesq approximation (7) in which properties are assumed at the reference condition (T_o, P_o), except in the temperature buoyancy term, and neglect pressure buoyancy. Since time dependence in this problem may be relegated mainly to heat transport, the motion equation may be considered quasi-steady. We get:

$$\frac{\partial}{\partial x_p} \left(K^\phi_{pq} \left(\frac{\partial P}{\rho g \, \partial x_q} + \delta_{q1} + \beta^T (T - T_o) \delta_{q1} \right) \right) + Q^\phi = 0 \quad (3)$$

where $K^\phi = \rho g \frac{k}{\mu}$ = hydraulic permeability tensor, and $\quad$ (4)

$$\phi = \frac{P}{\rho g} + z = \text{hydraulic potential or head.} \quad (5)$$

$$\left(\text{Note: } \frac{\partial \phi}{\partial x_q} = \frac{\partial P}{\rho g \, \partial x_q} + \delta_{q1} \right)$$

The final equation becomes:

$$K^\phi_{pq} \frac{\partial^2 \phi}{\partial x_p \partial x_q} + K^\phi_{pq} \beta^T \frac{\partial T}{\partial x_p} \delta_{q1} + Q^\phi = 0 \quad \cdot \quad (6)$$

The seepage flux is defined by:

$$q_p = -K^\phi_{pq} \left(\frac{\partial \phi}{\partial x_q} + \beta^T (T - T_o) \delta_{q1} \right) \quad (7)$$

and the pore velocity by:

$$v_p = q_p / n \quad \cdot \quad (8)$$

In sealed flow cases, T_o for neutral buoyancy cannot be determined a priori and must be calculated as the average temperature in the mesh. With substantial throughflow, T_o is fixed at the inflow supply temperature.

A.2.2 Heat Transport

The energy equation describing heat transport in a porous medium containing only a solid and one liquid phase is:

$$\frac{\partial}{\partial x_p} \left(K^T_{pq} \frac{\partial T}{\partial x_q} \right) - n\rho c_v v_p \frac{\partial T}{\partial x_p} - \bar{\rho} \, \bar{c}_v \frac{\partial T}{\partial t} + Q^T = 0 \qquad (9)$$

where $K^T = (nK^T_f + (1 - n)K^T_s) =$ thermal conductivity tensor with fluid and solid components,

c_v = fluid specific heat,

$\bar{\rho}$ = bulk density,

$\bar{c}_v$ = bulk specific heat,
    such that $\bar{\rho} \, \bar{c}_v = (n\rho_f c_{vf} + (1 - n) \rho_s c_{vs})$, and

$Q^T = (nQ^T_f + (1 - n)Q^T_s) =$ volumetric heat source
    including $\rho c_v Q^\phi T_b$ for source of fluid at temperature T_b.

All properties are rigorously defined at reference conditions.

A.2.3 Galerkin Functionals

In the finite element discretization of the modeled region, the physical space is broken up into elements, in each of which the potential and temperature fields are described by a simple relationship:

$$\phi = \phi_i N_i, \quad T = T_i N_i \qquad (10)$$

The ϕ_i and T_i are solution values at specified discretization points or nodes and the N_i are distribution functions associated with each node attached to the element.

The T_i are also assumed to vary linearly over time steps so that accounting for the time variation expands equation 10b into:

$$T = T^k_i M_k N_i \qquad (11)$$

in which the M_k are time distribution functions associated with the old and new solutions spanning the time interval in question.

A weighted-residual method may be used with finite elements to formulate a set of linear algebraic equations representative of the modeled system. In this model, we use Galerkin weighting in which the residuals, multiplied by the distribution functions previously mentioned, are integrated over time and space and set equal to zero.

The form of the flow Galerkin pseudo-functional is:

$$\chi_\phi = \int_R (K_{pq}^\phi \frac{\partial^2 \phi}{\partial x_p \partial x_q} + K_{pq}^\phi \beta^T \frac{\partial T}{\partial x_p} \delta_{q1} + Q^\phi)^\dagger \phi \, dR \quad . \tag{12}$$

† The residual is not varied in the derivation of the matrix equations below.

Integration-by-parts and the divergence theorem yield a quadratic formulation:

$$\chi_\phi = \int_R (-K_{pq}^\phi \frac{\partial \phi^\dagger}{\partial x_p} \frac{\partial \phi}{\partial x_q} + (K_{pq}^\phi \beta^T \frac{\partial T}{\partial x_p}^\dagger \delta_{q1} + Q^\phi) \phi) \, dR$$

$$+ \int_S (K_{pq}^\phi \frac{\partial \phi^\dagger}{\partial x_p} i_q) \phi \, dS \tag{13}$$

where the second integral is a boundary flux term used to express natural boundary conditions (impermeable) or imposed seepage fluxes and specified head outside a film. i_q is the outward-facing normal to the surface S. This term is omitted for impermeable boundaries or those with fixed potential and is replaced for other boundaries by:

$$\int_S (q^\phi - \alpha^\phi (\phi - \phi_a))^\dagger \phi \, dS \tag{14}$$

where the imposed seepage flux q^ϕ is positive for fluid inflow.
α^ϕ is the film transfer coefficient and ϕ_a is the ambient potential
outside the film. The impermeable condition implies that the
boundary is a streamline.

The energy Galerkin pseudo-functional is:

$$\chi_T = \int_R (K^T_{pq} \frac{\partial^2 T}{\partial x_p \partial x_q} - n\rho c_v v_q \frac{\partial T}{\partial x_q} - \overline{\rho}\,\overline{c}_v \frac{\partial T}{\partial t} + Q^T)^\dagger T \; dR \; . \quad (15)$$

Integrating-by-parts and using the divergence theorem we get:

$$\chi_T = \int_R (-K^T_{pq} \frac{\partial T^\dagger}{\partial x_p} \frac{\partial T}{\partial x_q} + n\rho c_v v_q T^\dagger \frac{\partial T}{\partial x_q} - \overline{\rho}\,\overline{c}_v \frac{\partial T^\dagger}{\partial t} T + Q^T T) \; dR$$

$$+ \int_S (K^T_{pq} \frac{\partial T}{\partial x_p} - n\rho c_v v_q T)^\dagger i_q T \; dS \quad \cdot \quad (16)$$

Surface integrals drop out if surface throughflow is zero or if T
is specified on the boundary. Thermal load boundaries may exist
in which the first surface integral term is replaced by:

$$\int_S (q^T - \alpha^T (T - T_a))^\dagger T \; dS \quad (17)$$

analogous to the fluid case.

In cases of fluid inflow with specified temperature T_b, the second
term should be replaced by:

$$\int_S (-n\rho c_v v_q T_b i_q)^\dagger T \; dS \quad (18)$$

to account for the heat brought in by the forced inflow of fluid.

The boundary conditions on potential/fluid flow and temperature/ heat
flux are only compatible in certain combinations. In general, if
there is fluid throughflow, the temperature at the boundary must be
specified. Otherwise, any temperature boundary condition is permitted.

Seepage fluxes and pore velocities are by equation (7) a function of the choice of T_0, the nonbuoyant reference temperature. A bad choice of T_0 can produce velocities that are totally unrealistic.

A good choice of T_0 is the boundary temperature if it is known. If it is not known, or highly variable over the boundary, it may not be possible to maintain the impervious nature of boundaries in the model run and the results may be invalidated. The model permits the use of a specified single value for T_0 or for the use of the average temperature in the modeled region, calculated in the program automatically. More involved determinations of T_0 may be required, e.g. when the neutral buoyancy temperature is a function of position in the mesh, and this must be added in special subroutines created for that purpose.

A.2.4 Matrix Equation

When the first variations of these pseudo-functionals are taken as:

$$\partial \chi_\phi / \partial \phi_i = 0 \;\; , \;\; \partial \chi_T / \partial T_i = 0 \tag{19}$$

The following matrix equations result at time ℓ, where Δt = time step length:

(a) for fixed α^ϕ and α^T:

$$
\left(\frac{2}{3} \left[\begin{pmatrix} \overline{K}^T + \overline{\alpha}^T & 0 \\ 0 & \overline{K}^\phi + \overline{\alpha}^\phi \end{pmatrix} + \frac{1}{\Delta t} \begin{pmatrix} \overline{C}^T & 0 \\ 0 & 0 \end{pmatrix} \right] \right) \begin{pmatrix} T_j \\ \phi_j \end{pmatrix}^\ell =
$$

$$
\left(-\frac{1}{3} \left[\begin{pmatrix} \overline{K}^T + \overline{\alpha}^T & 0 \\ 0 & \overline{K}^\phi + \overline{\alpha}^\phi \end{pmatrix} + \frac{1}{\Delta t} \begin{pmatrix} \overline{C}^T & 0 \\ 0 & 0 \end{pmatrix} \right] \right) \begin{pmatrix} T_j \\ \phi_j \end{pmatrix}^{\ell-1}
$$

$$
+ \frac{1}{3} \begin{pmatrix} \overline{Q}^T + \overline{q}^T + \overline{\overline{\alpha}}^T T_a \\ \overline{Q}^\phi + \overline{q}^\phi + \overline{\overline{\alpha}}^\phi \phi_a \end{pmatrix}^{\ell-1} + \frac{2}{3} \begin{pmatrix} \overline{Q}^T + \overline{q}^T + \overline{\overline{\alpha}}^T T_a \\ \overline{Q}^\phi + \overline{q}^\phi + \overline{\overline{\alpha}}^\phi \phi_a \end{pmatrix}^\ell
$$

$$
\begin{pmatrix} nTv_q \overline{V}_q \\ K^\phi \frac{\partial T}{p1 \partial x_p} \beta^T \beta \end{pmatrix}^{\ell-1}
\tag{20a}
$$

(b) for variable α^T and α^ϕ:

   delete $\overline{\alpha}^T$ and $\overline{\alpha}^\phi$ on L.H.S. and
   replace $\overline{\alpha}^T$ and $\overline{\alpha}^\phi$ on R.H.S. by $3\overline{\alpha}^T$ and $3\overline{\alpha}^\phi$.   (20b)

(c) for steady-state solutions:

$$
\begin{pmatrix} \overline{K}^T + \overline{\alpha}^T & 0 \\ 0 & \overline{K}^\phi + \overline{\alpha}^\phi \end{pmatrix}
\begin{pmatrix} T_j & \\ & \\ \phi_j \end{pmatrix}^\ell =
$$

$$
\begin{pmatrix} \overline{Q}^T + \overline{q}^T + \overline{\overline{\alpha}}^T T_a \\ \\ \overline{Q}^\phi + \overline{q}^\phi + \overline{\overline{\alpha}}^\phi \phi_a \end{pmatrix}^\ell
+
\begin{pmatrix} nTv_q \overline{V}_q \\ \\ K^\phi_{p1} \dfrac{\partial T}{\partial x_p} \beta^T \overline{\beta} \end{pmatrix}^{\ell-1}
$$ (20c)

where ℓ denotes the ℓth iteration for steady-state runs. At least
three iterations must be performed for steady-state runs in order
to load the buoyancy and advection terms.

The buoyancy and advection terms are written out fully above to
indicate that T, v_q and $\partial T/\partial x_p$ must be evaluated for each element
from the old solution before these terms are added into the load
vector. All other loads are specified and not solution dependent.
This solution dependence of advection and buoyancy requires the use
of an iterative process in solution of steady-state cases, but for
transient problems, the nonlinearity can be assimilated into the normal
time-stepping solution procedure.

The component matrices are:

$$\overline{K} = \overset{e}{\underset{e}{\Sigma}} \int_e K_{pq} N_{j,x_p} N_{i,x_q} \, dR, \text{ where } \overset{e}{\Sigma} \text{ is a sum over elements,}$$

$$\overline{\alpha} = \overset{e}{\underset{e}{\Sigma}} \int_e \alpha N_j N_i \, dS, \text{ where } S \text{ is the external boundary,}$$

$$\overline{\overline{\alpha}} = \overset{e}{\underset{e}{\Sigma}} \int_e \alpha N_i \, dS \text{ ,}$$

$$\bar{C} = \sum_e \int_e \bar{\rho}\, \bar{c}_v\, N_j N_i \; dR \quad ,$$

$$\bar{Q} = \sum_e \int_e QN_i \; dR + Q_i \quad , \text{ where } Q_i \text{ is a nodal load,}$$

$$\bar{q} = \sum_e \int_e qN_i \; dS \quad ,$$

$$\bar{V}_m = \sum_e \int_e \rho c_v N_{i,x_m} \; dR \quad , \text{ and}$$

$$\bar{\beta} = \sum_e \int_e N_i \; dR \quad .$$

These matrices are presented for the general case. In 2-dimensional planar geometry (dR) is (dxdy). For axisymmetry, replace (dR) by $(2\pi r dz dr)$, (dS) by $(2\pi r dS)$ and (Q_i) by $(2\pi r Q_i)$.

A.2.5 Units

The program is written in dimensionless form so that any compatible set of units for the input data will be acceptable.

As an example consider the following: In metric units, the desireable units of length and time may be the metre and second respectively. The values of K^T, $\bar{\rho}$ and $\bar{c}_v$ (ρ and c_v, as well) must be given in the input data so that the thermal diffusivity $K^T/\bar{\rho}\,\bar{c}_v$ has units of (m^2/sec). K^T in $(\text{W/m}^\circ\text{C})$, $\bar{\rho}$ in (kg/m^3) and $\bar{c}_v$ in $(\text{W sec/kg}^\circ\text{C})$ will achieve this and are compatible. It is important to note that the nodal locations must be given in the length unit desired, (m) in this example. Also, thermal loading units must be compatible: volumetric heat generation Q^T in (W/m^3), boundary heat flux q^T in (W/m^2), thermal convection coefficient α^T in $(\text{W/m}^2 \; {}^\circ\text{C})$, ambient temperature T_a in $({}^\circ\text{C})$ and nodal load Q_i^T in (W). For flow parameters, K^ϕ is in (m/sec), n is dimensionless and β^T is in $({}^\circ\text{C}^{-1})$. The constituent units for K^ϕ are k in (m^2), μ in (kg/m sec), ρ in (kg/m^3) and g in (m/sec^2). The loading units are: volumetric source Q^ϕ in $(\text{sec}^{-1}$, actually m^3/m^3 sec), boundary normal seepage flux q^T in (m/sec), film coefficient α^ϕ in (sec^{-1}), film external potential ϕ_a in (m) and nodal source Q_i^ϕ in (m^3/sec).

use is made of disc storage and all data passed from overlay to overlay is written to disc and reread in the newly loaded overlay. This not only minimizes storage but facilitates restarts during long runs or multiple runs and simplifies debugging.

APPENDIX B

Model Validation

B.1 Thermal validation

Validation of the model for heat transport was carried out with an analytical solution to a 1-dimensional problem. The initial conditions are zero temperature everywhere in the positive x region at time $t = 0$ and the boundary conditions are $T(0) = 1$ and $T(\infty) = 0$ for $t>0$. The analytical solution is:

$$T = \frac{1}{2} \left[\; erfc \left(\frac{x - vt}{(4Kt)^{\frac{1}{2}}} \right) \; + \; exp \left(\frac{xv}{K} \right) \; erfc \left(\frac{x + vt}{(4Kt)^{\frac{1}{2}}} \right) \right]$$

where v is the velocity and K is the thermal conductivity. The problem was solved for $v = 5$, $K = 1$, $0 \le x \le 1$, and with 25 time steps of 0.002 to 0.05. The boundary condition $T(1) = 0$ was an approximation of $T(\infty) = 0$ so the solution is reported here to $x = 0.5$ Table B.1 gives the analytical and model results for various x values at $t = 0.05$. The agreement is excellent.

Table B.1 Thermal validation results

x	T analytical	T model
0	1.0	1.0
0.05	0.9565	0.9532
0.10	0.9037	0.9052
0.15	0.8422	0.8380
0.20	0.7732	0.7747
0.25	0.6988	0.6931
0.30	0.6212	0.6211
0.35	0.5422	0.5347
0.40	0.4646	0.4631
0.45	0.3912	0.3823
0.50	0.3228	0.3191

This problem is strongly advective and adequately tests thermal capacity, conductivity and advection terms, all the transport terms in the heat transfer equation.
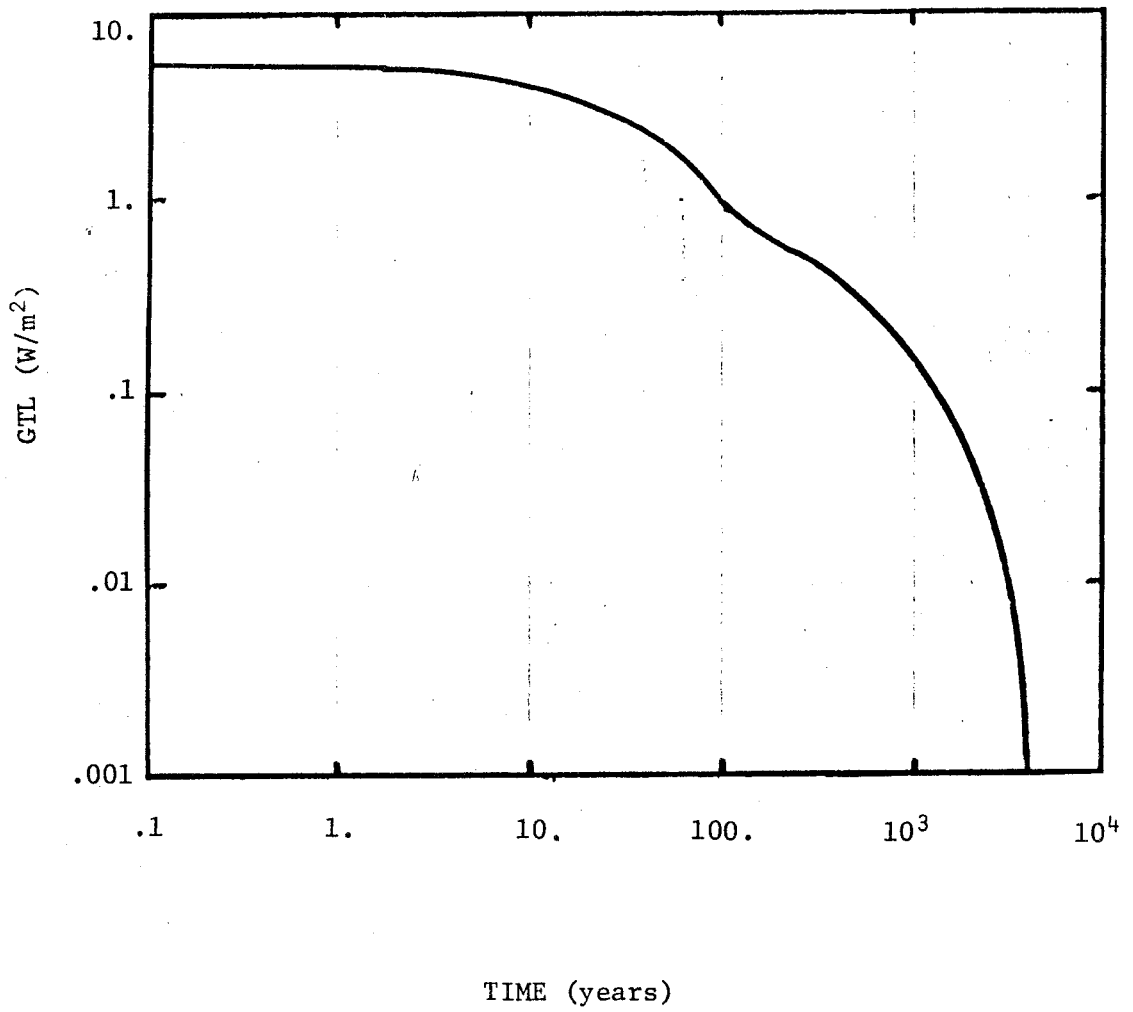
Figure   1 - Decay of GTL for 40 year old HLW
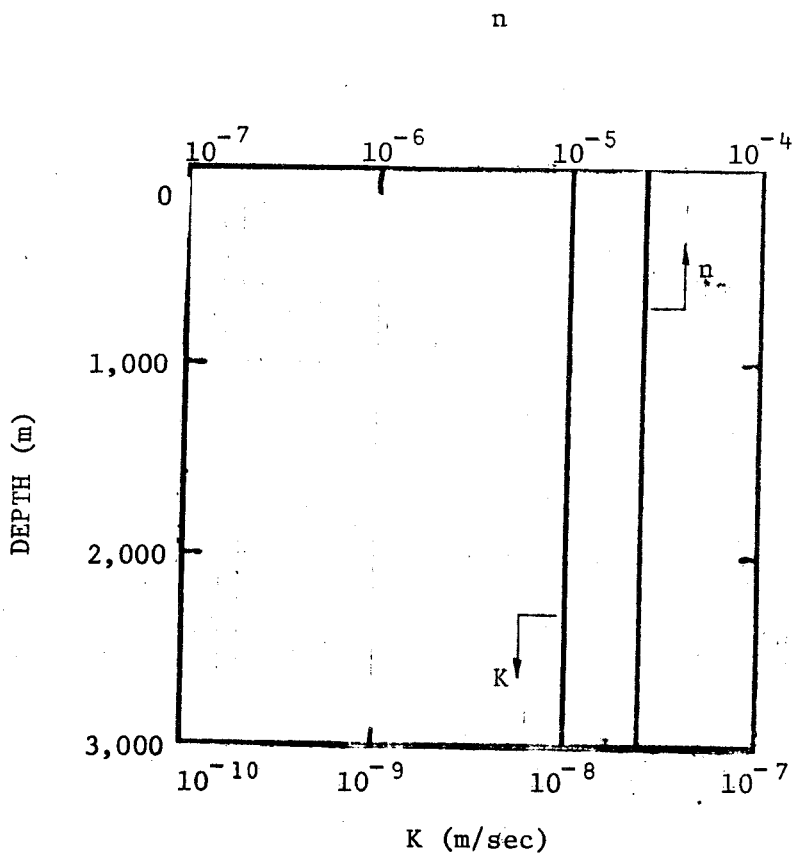
Case 1: K homogeneous, isotropic

Case 2: K(Z)

Isotropic, non-homogeneous
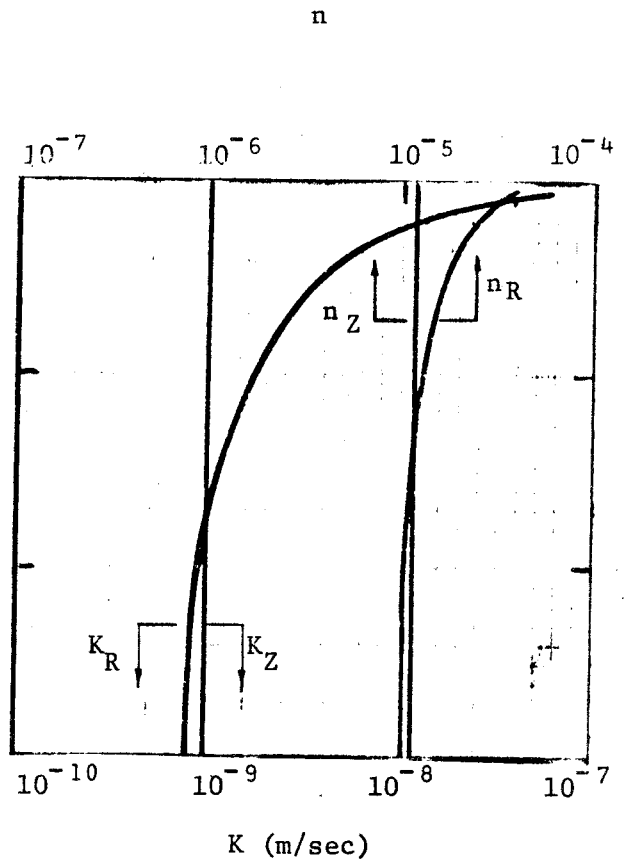
Anisotropic, non-homogeneous

Case 3: $K_R(Z)$, K_Z uniform

R = horizontal
Z = vertical
S = Fracture spacing = 1.8 m
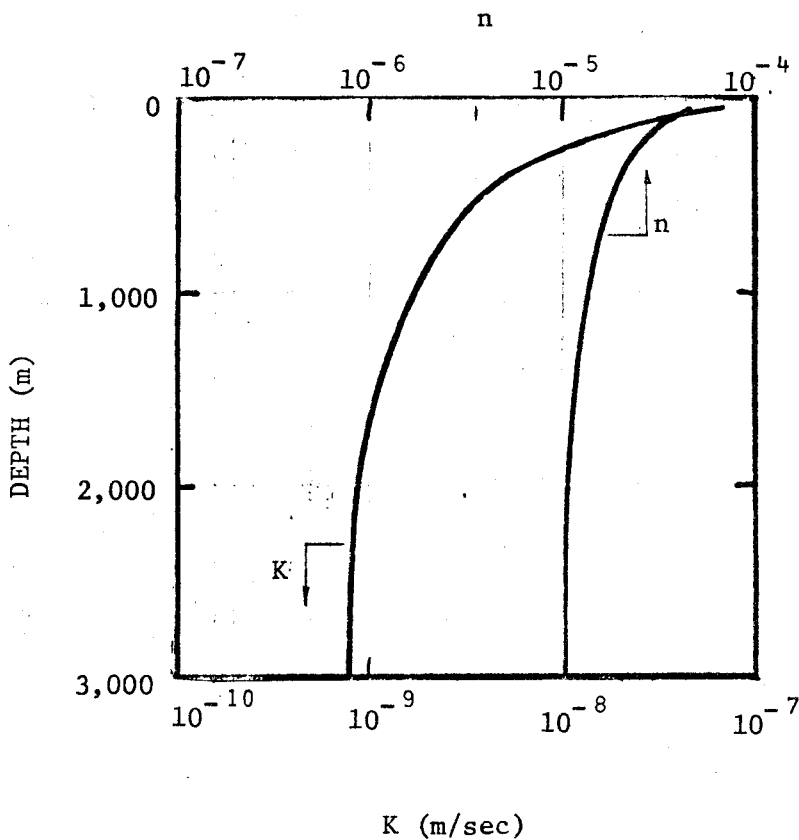
FIGURE 2: Distributions of Permeability (K) and Porosity (n) with depth (Z)

Figure 3 : Material Properties

Property	Symbol	Units	Host rock	Backfill
Thermal				
Thermal Conductivity (bulk)	K^T	$W/m\,^{\circ}C$	2.05	1.5
Density - water	ρ	kg/m^3	1,000	1,000
- bulk	$\bar{\rho}$		2,800	2,070
Specific Heat - water	c_v	$J\,/kg\,^{\circ}C$	4,180	4,180
- bulk	$\bar{c}_v$		735	936
Flow				
Permeability	K	m/s	*	10^{-10}
Porosity	n	—	*	0.37
Volumetric Expansion coefficient of water at $20^{\circ}C$	$-\beta^T$	$^{\circ}C^{-1}$	-2.07×10^{-4}	-2.07×10^{-4}

* Varies See Section 2.2 of this report.
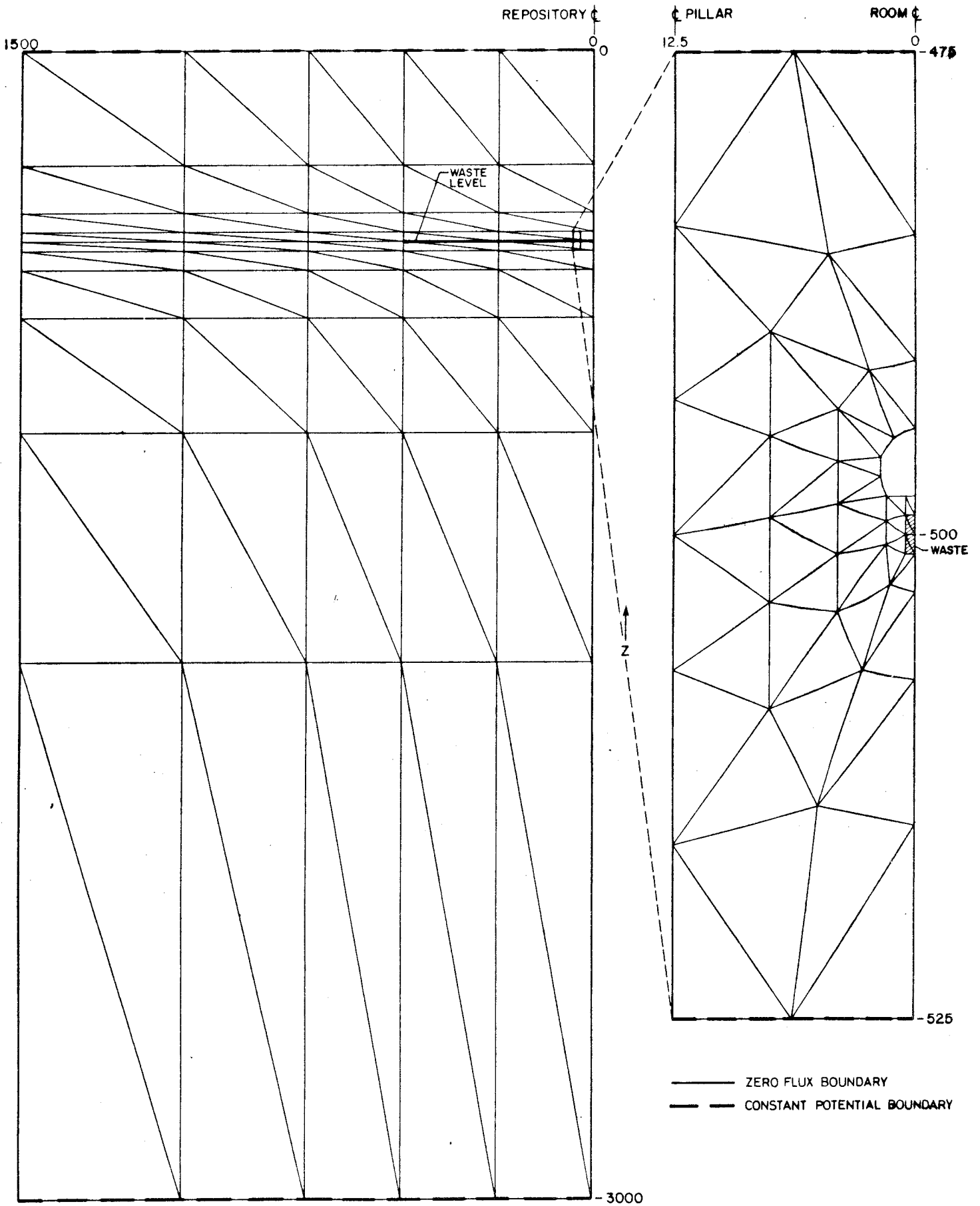
FIGURE 4. FINITE ELEMENT MODELS FOR FLOW ANALYSES
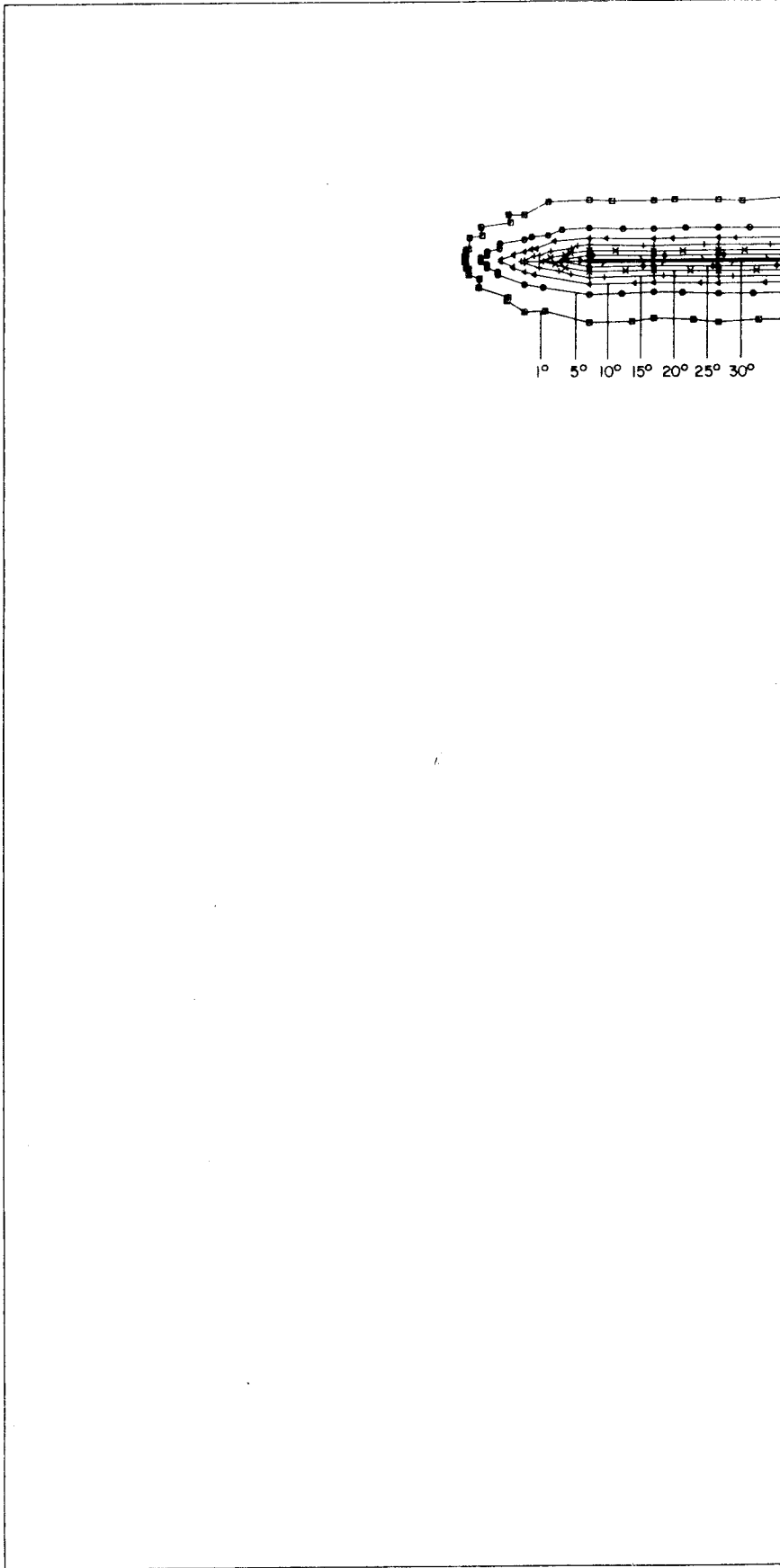
KBS HYDROTHERMAL RUN GLOBAL I
CONDUCTION BASELINE

CONTOUR PLOT

GEOMETRY SCALE   0   500 M

TEMPERATURE CONTOURS IN DEGREES CELSIUS
40 YEARS
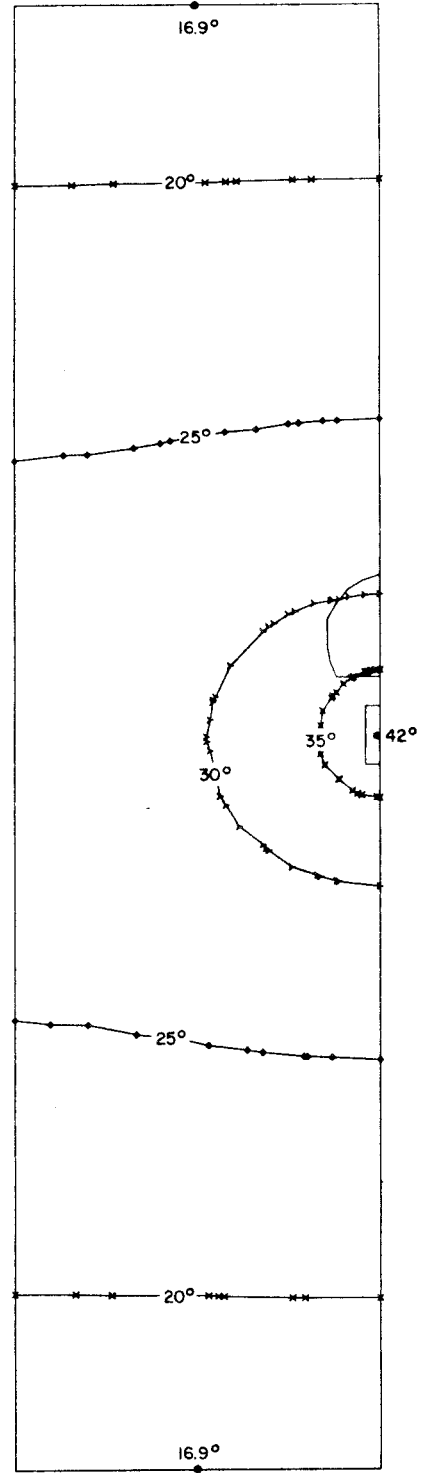
LOCAL I

0   5   10 M

FIGURE   5
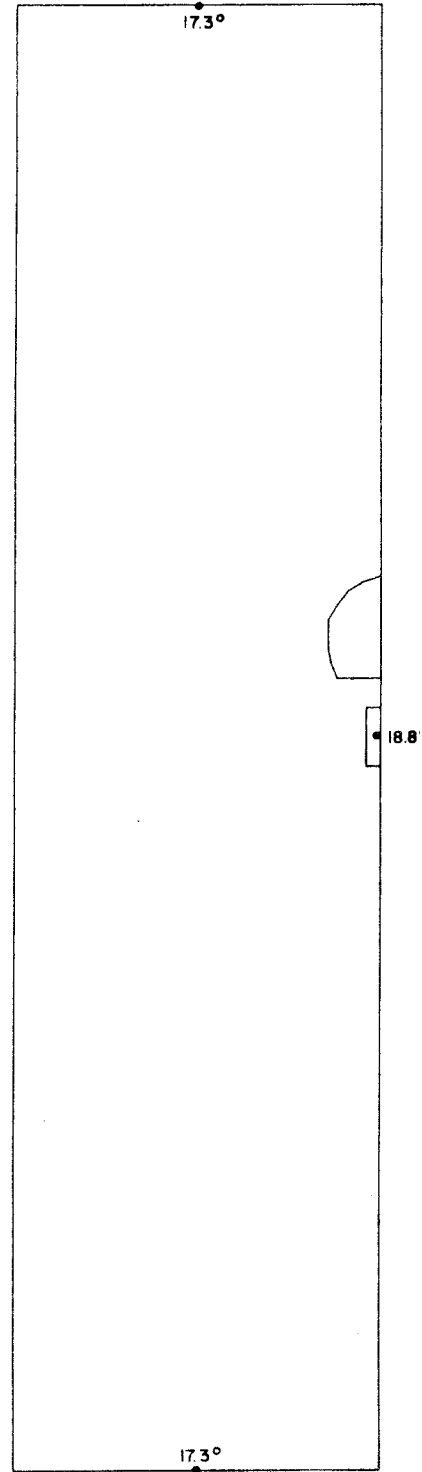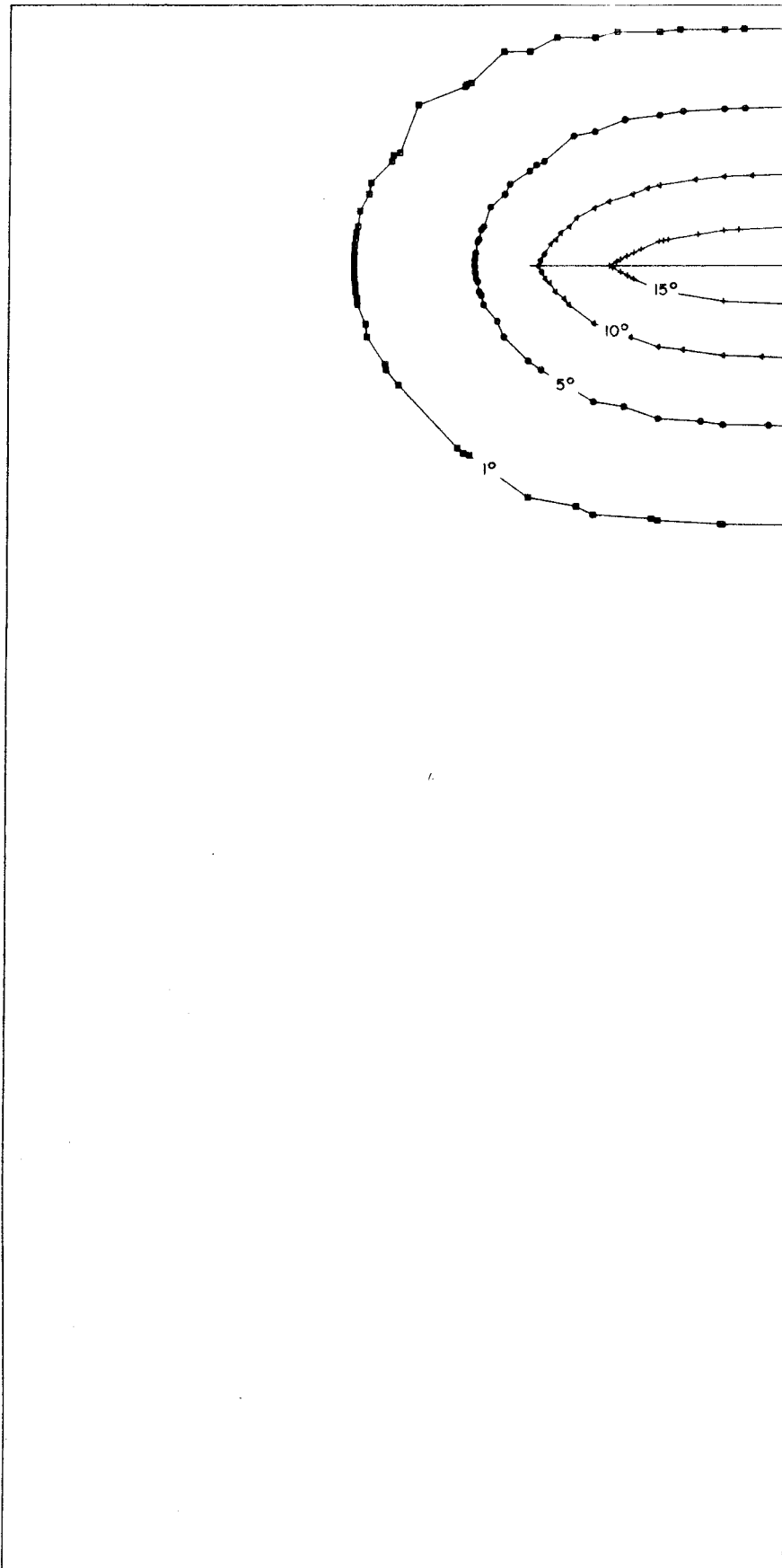
KBS HYDROTHERMAL RUN GLOBAL I
CONDUCTION BASELINE

CONTOUR PLOT

GEOMETRY SCALE

TEMPERATURE CONTOURS IN DEGREES CELSIUS

200 YEARS

LOCAL I

FIGURE 6

KBS HYDROTHERMAL RUN GLOBAL I
CONDUCTION BASELINE

CONTOUR PLOT

GEOMETRY SCALE

TEMPERATURE CONTOURS IN DEGREES CELSIUS

1000 YEARS
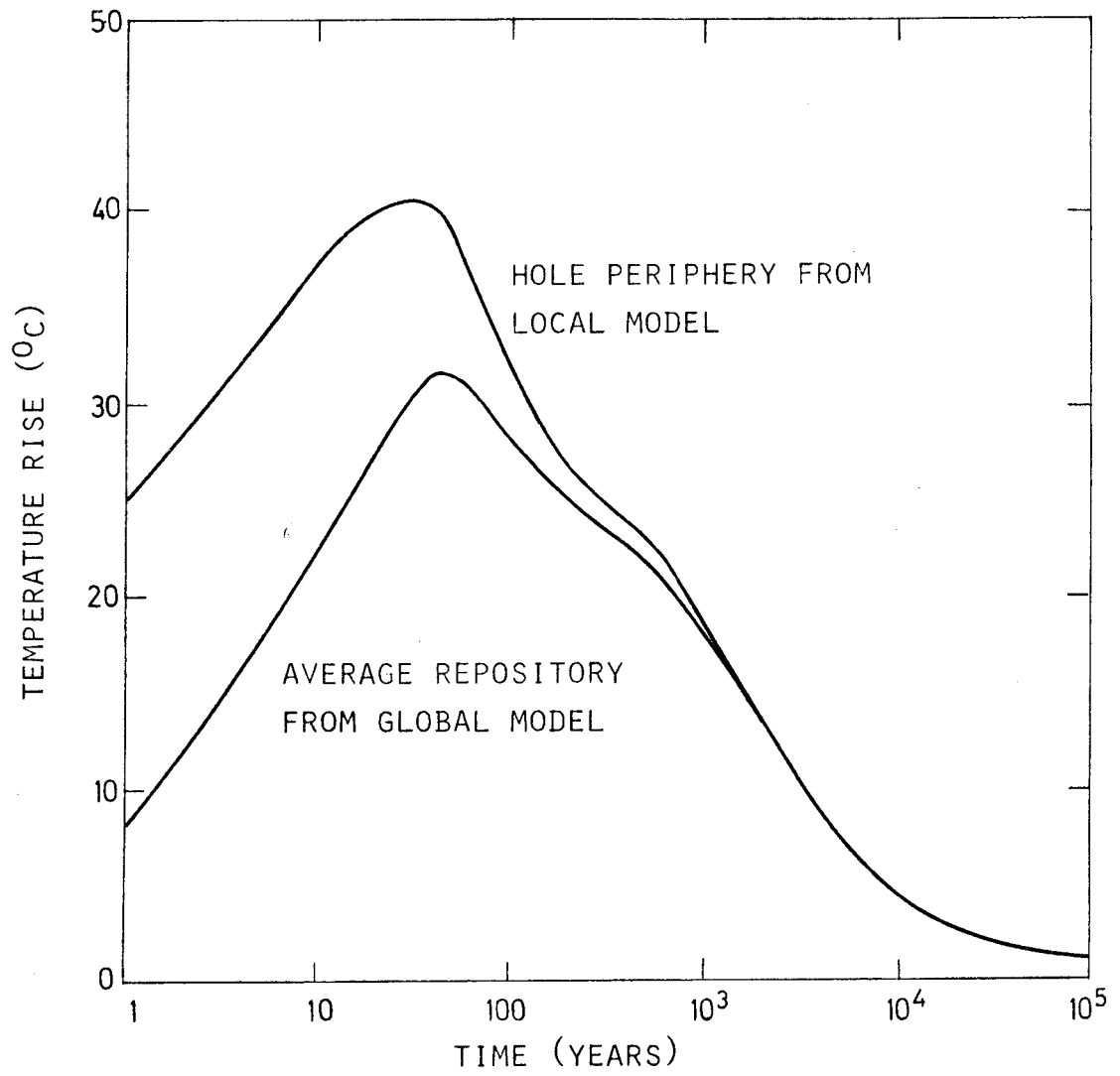
LOCAL I

FIGURE 7

FIGURE 8. CONDUCTION TEMPERATURE HISTORY IN
GLOBAL AND LOCAL MODELS
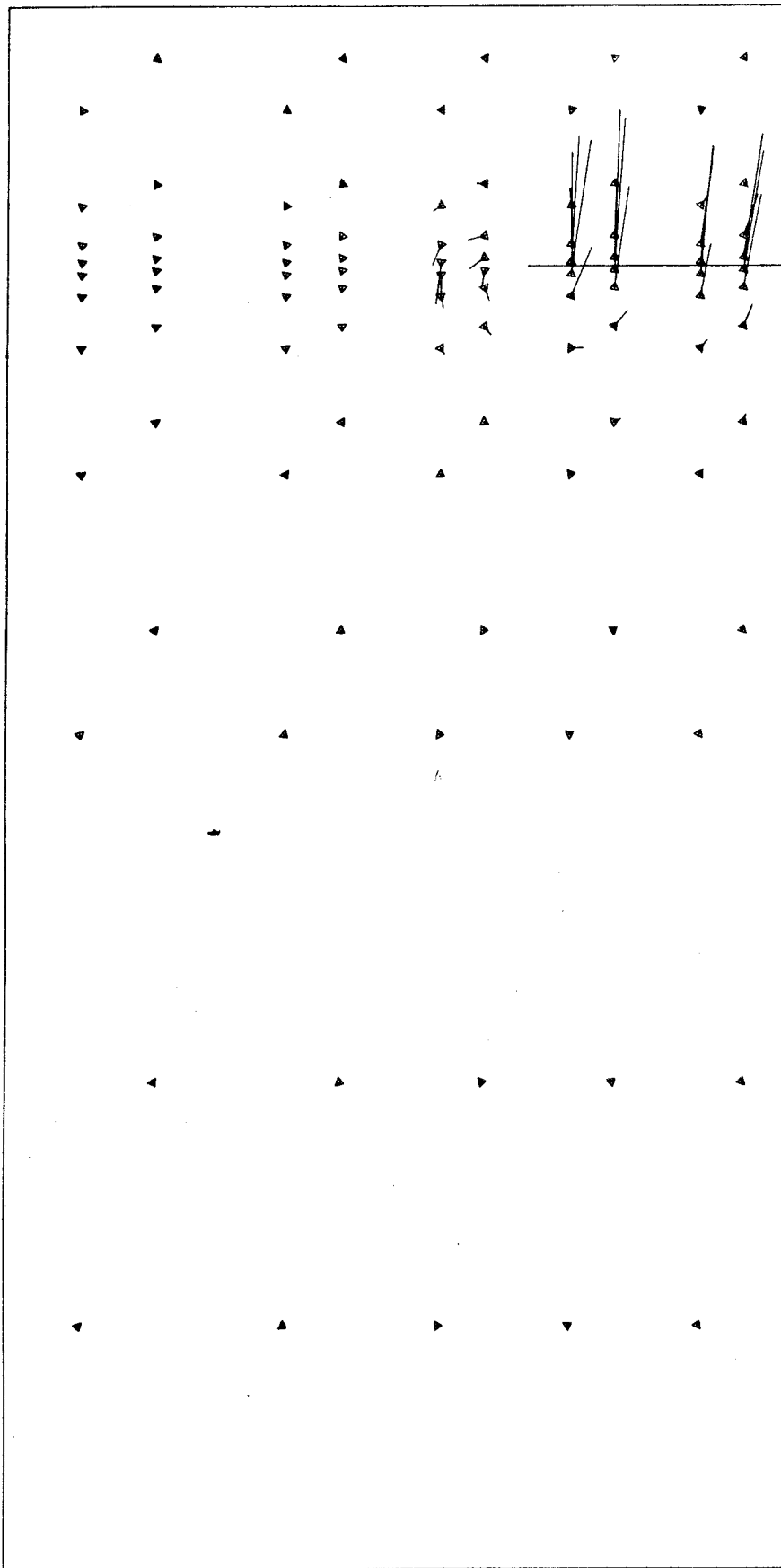
KBS HYDROTHERMAL RUN GLOBAL 2
K HOMOGENEOUS ISOTROPIC

LOCAL 2

VELOCITY PLOT

GEOMETRY SCALE |⊢___0_____500 M|

VELOCITY SCALE ⊢——⊣ = 0.4 ✳ 10 ✳ ✳ – 6 M/SEC
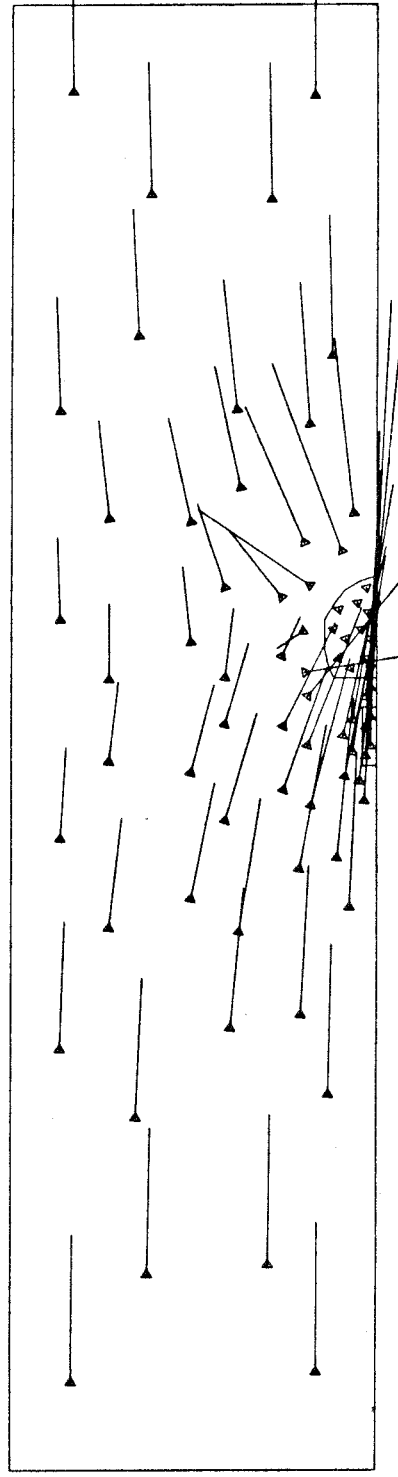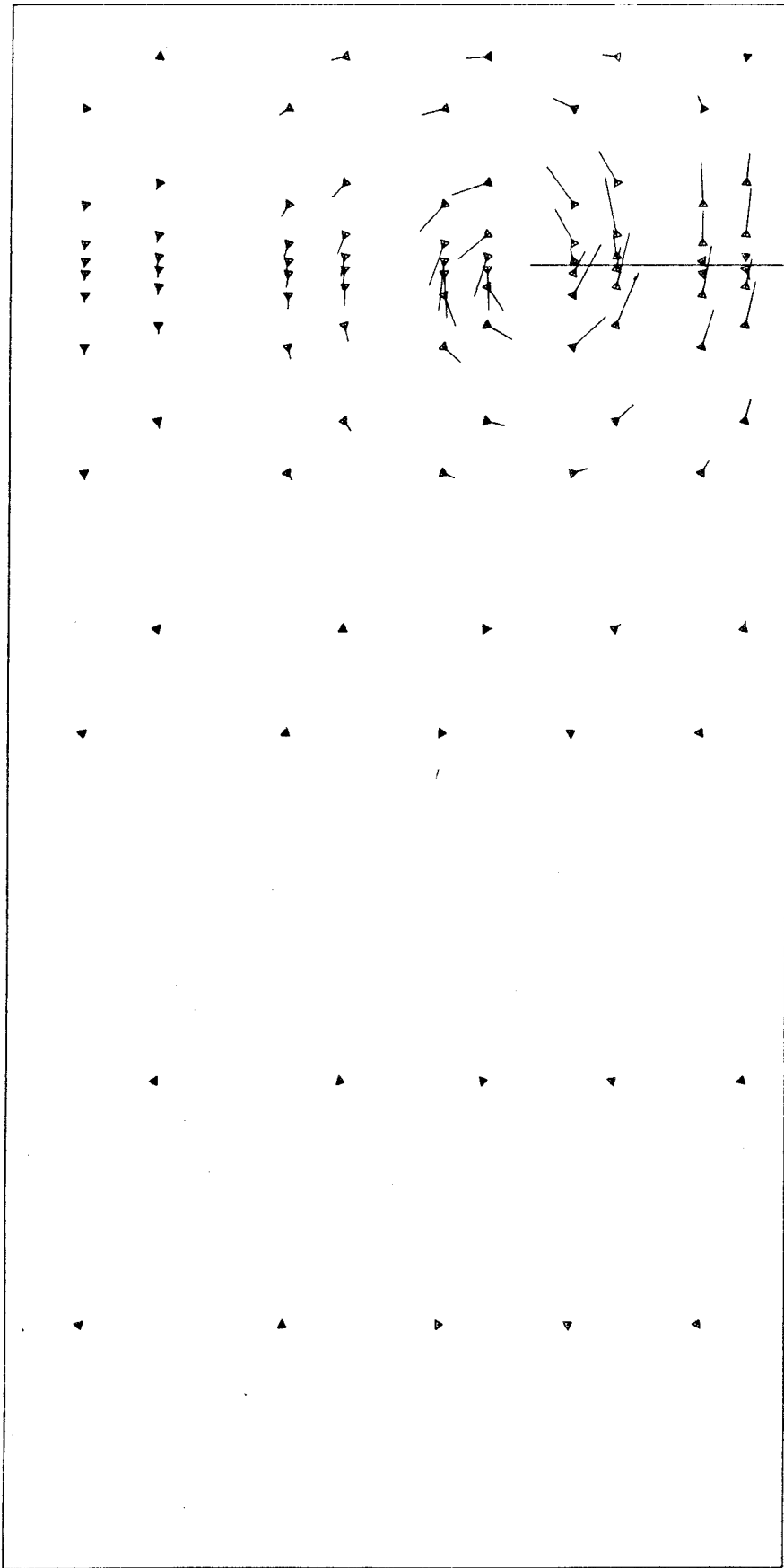
0   5   10 M

40 YEARS

FIGURE 9

KBS HYDROTHERMAL RUN GLOBAL 2
K HOMOGENEOUS ISOTROPIC

LOCAL 2

VELOCITY PLOT

GEOMETRY SCALE 0 ⊢————————⊣ 500 M

VELOCITY SCALE ⊢————⊣ = 0.4 ✱ 10 ✱ ✱ -6 M/SEC
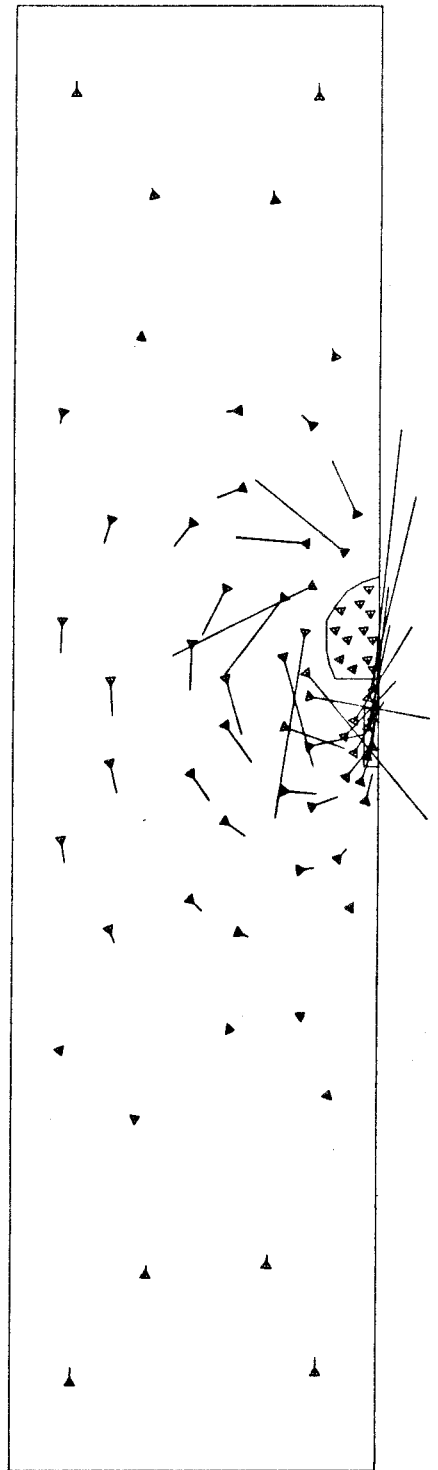
0   5   10 M
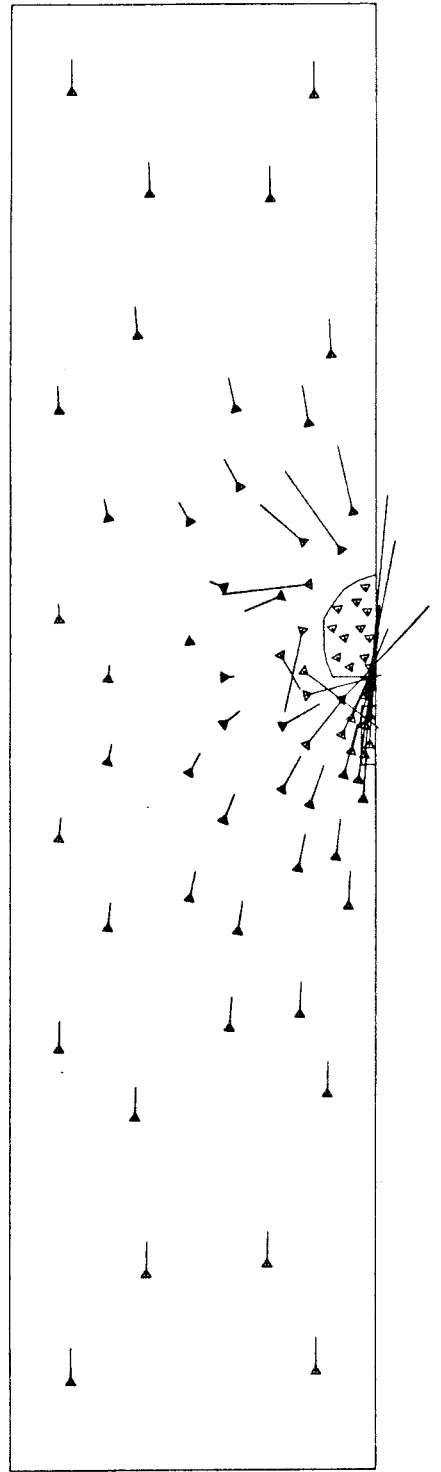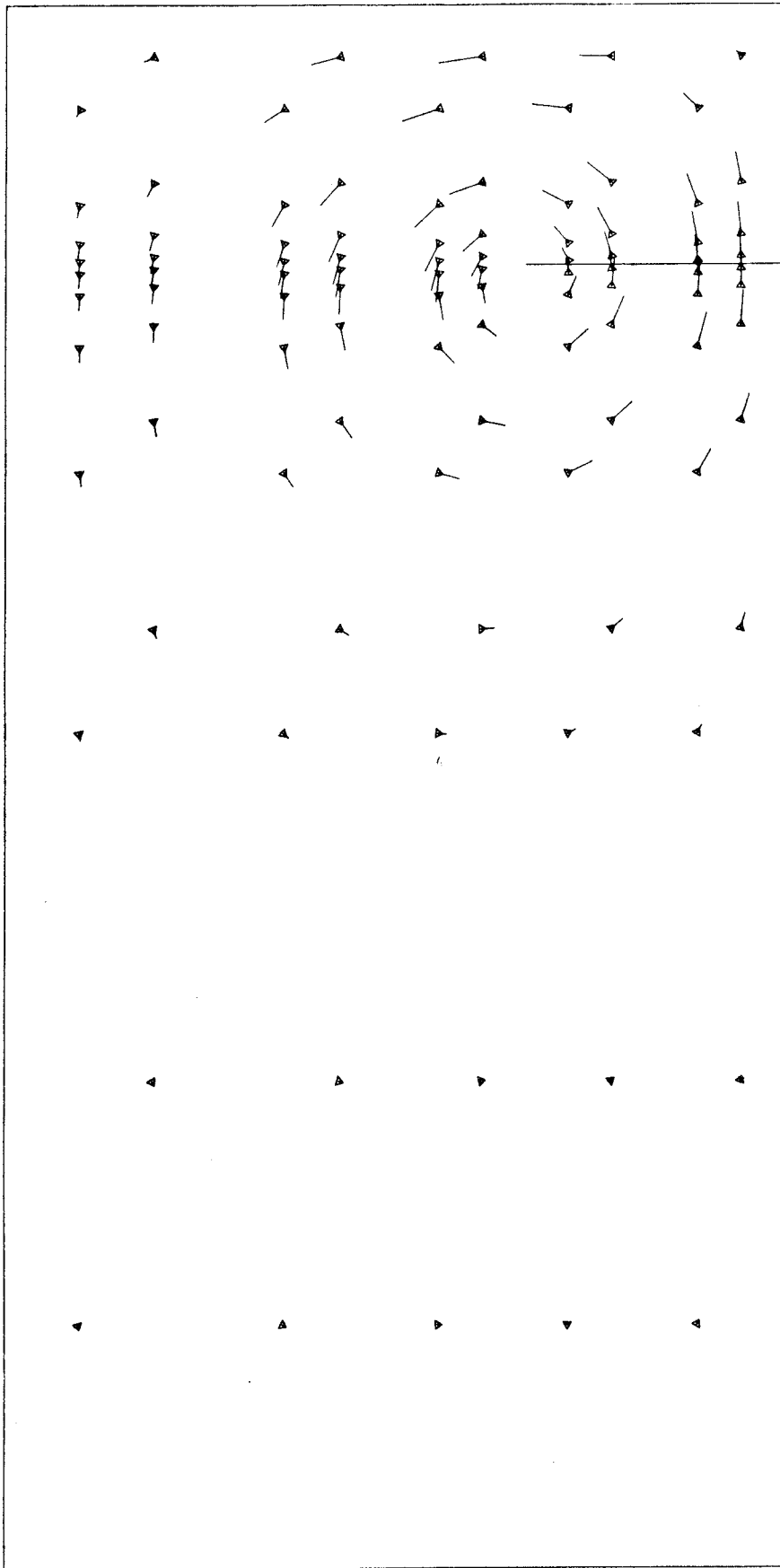
200 YEARS

FIGURE 10

KBS HYDROTHERMAL RUN GLOBAL 2
K HOMOGENEOUS ISOTROPIC

LOCAL 2

VELOCITY PLOT

GEOMETRY SCALE $\vdash$———$\dashv$ 0   500 M
VELOCITY SCALE $\vdash$———$\dashv$ = 0.4 ✳ 10 ✳ ✳ − 6 M / SEC

0   5   10 M

1000 YEARS

FIGURE 11
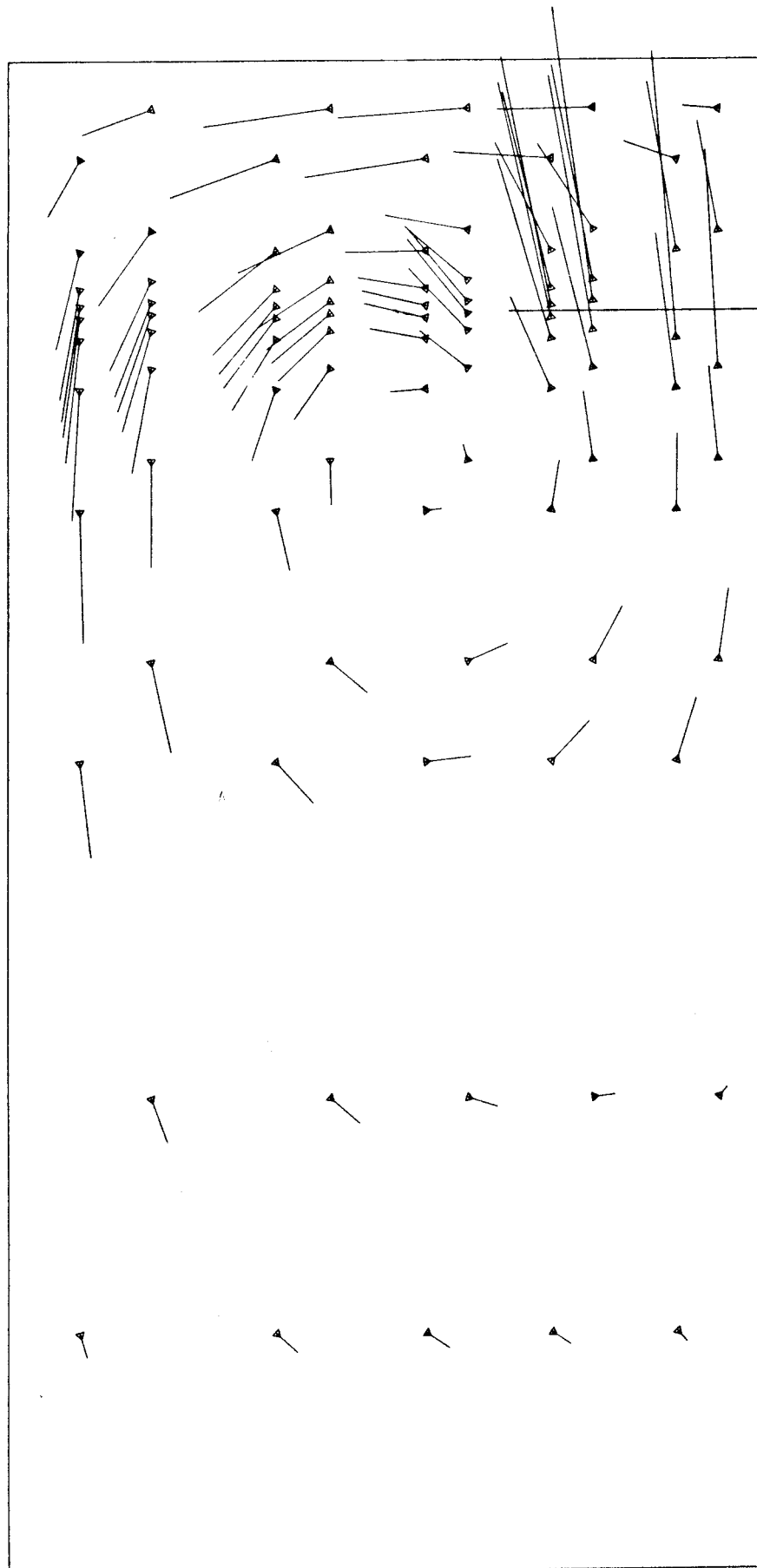
KBS HYDROTHERMAL RUN GLOBAL 2
K HOMOGENEOUS ISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE |—————| 0   500 M

VELOCITY SCALE |———| = 0.4 * 10 ** - 7 M/SEC

10 000 YEARS

FIGURE 12

KBS HYDROTHERMAL RUN GLOBAL 2
K HOMOGENEOUS ISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE 0 |_____| 500 M

VELOCITY SCALE |———| = 0.4 ∗ 10 ∗ ∗ - 7 M/SEC

100 000 YEARS
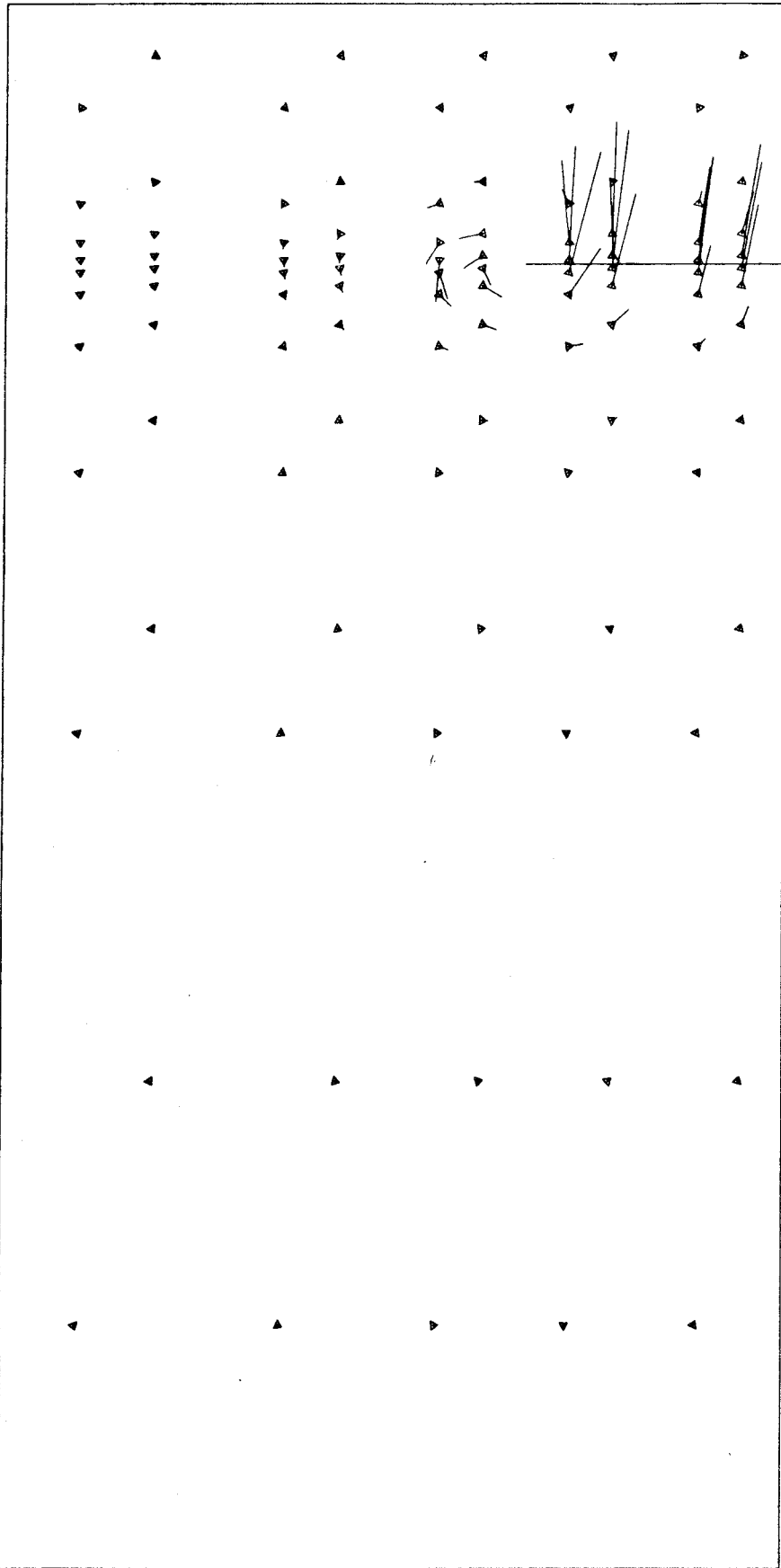
FIGURE 13

KBS HYDROTHERMAL RUN GLOBAL 3
K (Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE |0 —————— 500 M|

VELOCITY SCALE |———| = 0.8 * 10 ** –7 M / SEC

LOCAL 3

|0 —————— 5 —————— 10 M|

40 YEARS

FIGURE 14

KBS HYDROTHERMAL RUN GLOBAL 3
K(Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE |———————— 500 M
0

VELOCITY SCALE ⊢————⊣ = 0.8 ✳ 10 ✳✳ - 7 M/SEC

200 YEARS
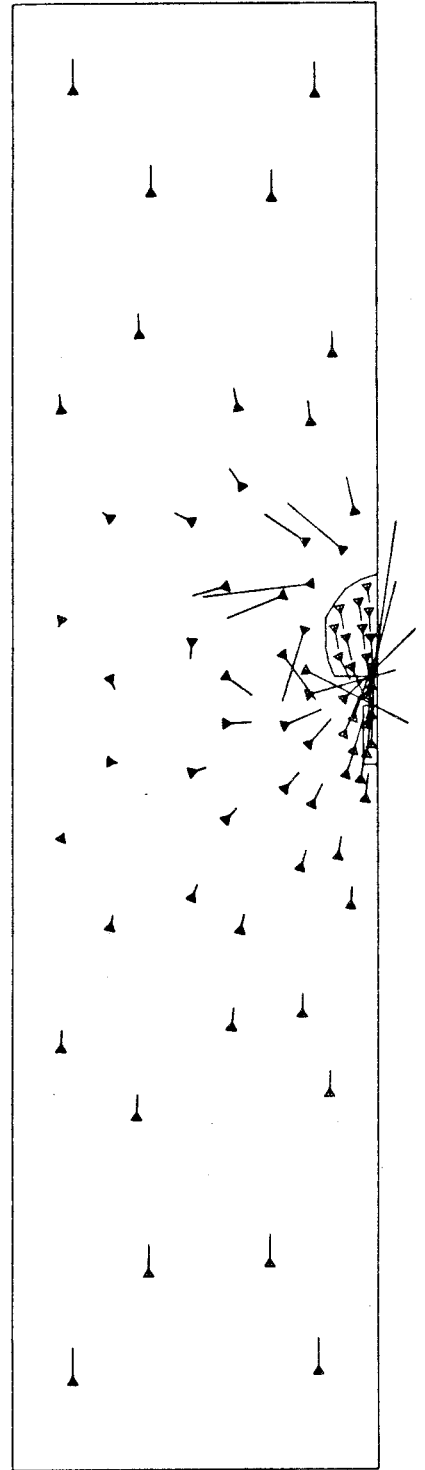
LOCAL 3

0   5   10 M

FIGURE 15
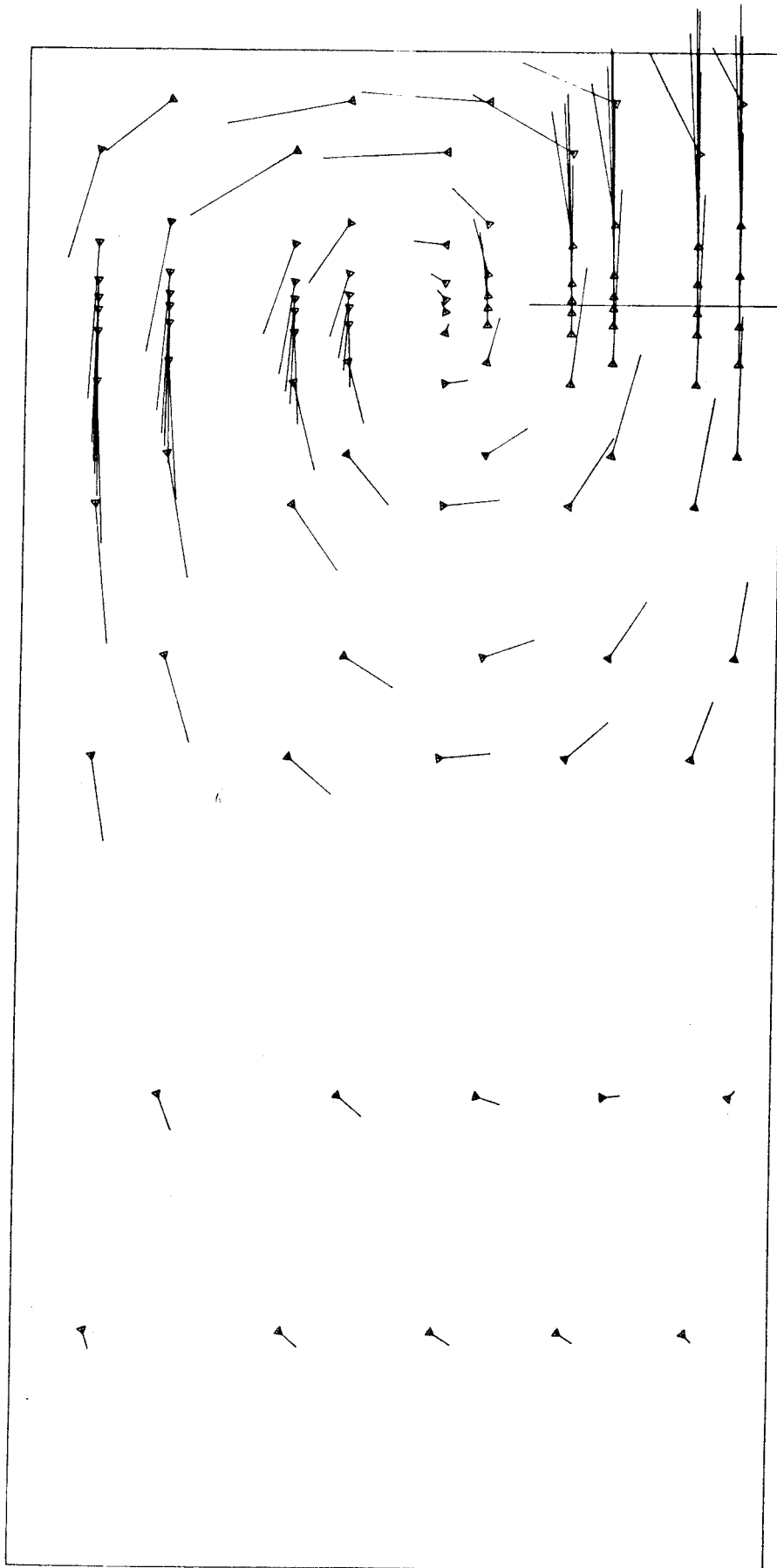
KBS HYDROTHERMAL RUN GLOBAL 3
K (Z) ANISOTROPIC

LOCAL 3

VELOCITY PLOT

GEOMETRY SCALE |0————————500 M
VELOCITY SCALE ├————┤ = 0.8 * 10 * * − 7 M/SEC

0   5   10 M

1000 YEARS

FIGURE 16

KBS HYDROTHERMAL RUN GLOBAL 3
K(Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE $\vdash\!\!\!\!\longrightarrow\!\!\!\!\dashv$ 0 ———————— 500 M

VELOCITY SCALE $\vdash\!\!\!\!\longrightarrow\!\!\!\!\dashv$ = 0.8 * 10 ** - 8 M/SEC

10 000 YEARS
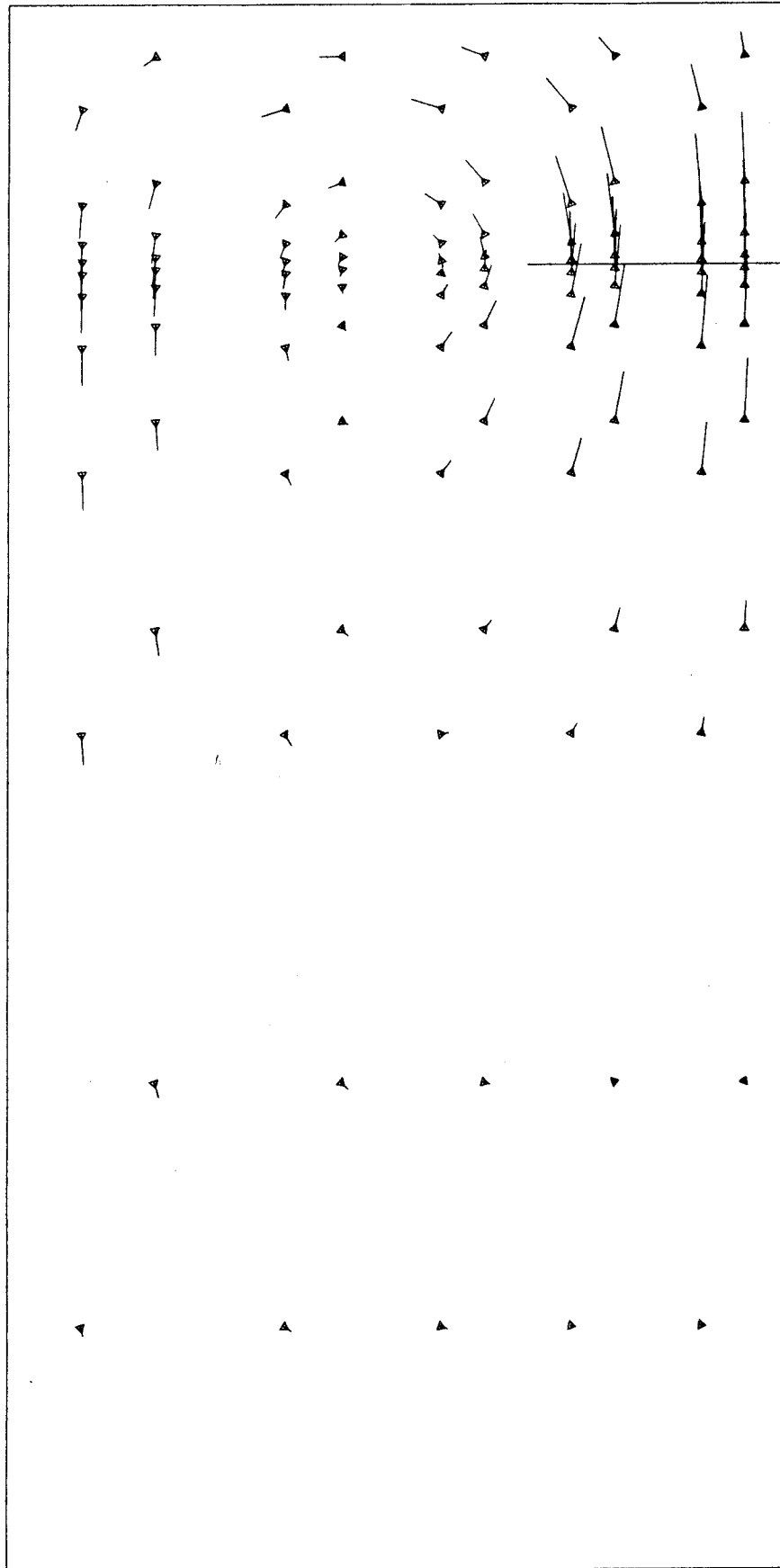
FIGURE 17

KBS HYDROTHERMAL RUN GLOBAL 3
K(Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE $\vdash$————$\dashv$ 0   500 M

VELOCITY SCALE $\vdash$————$\dashv$ = 0.8 $*$ 10 $**$ - 8 M/SEC

100 000 YEARS
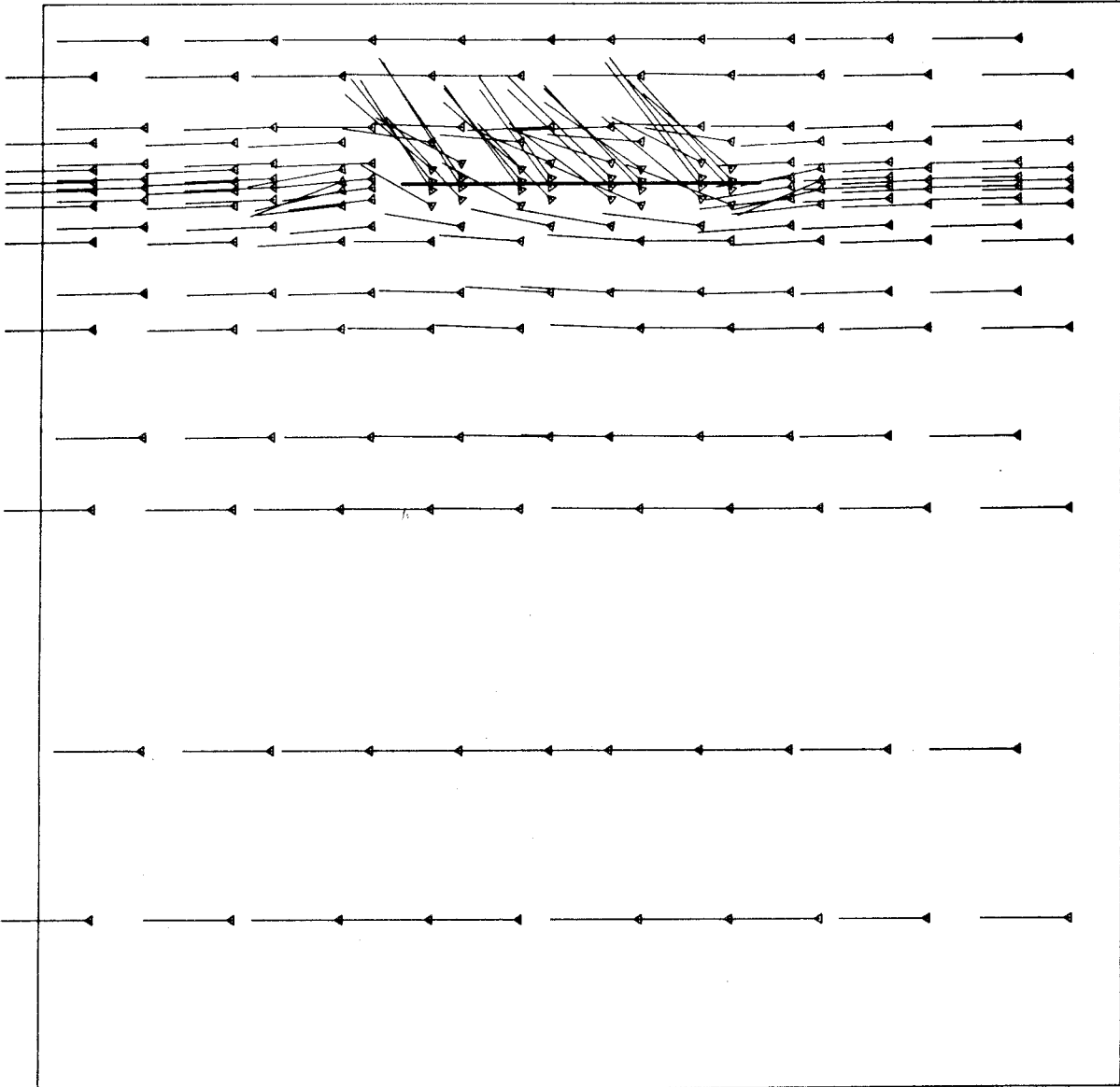
FIGURE 18

KBS HYDROTHERMAL RUN GLOBAL 2 WITH CROSSFLOW
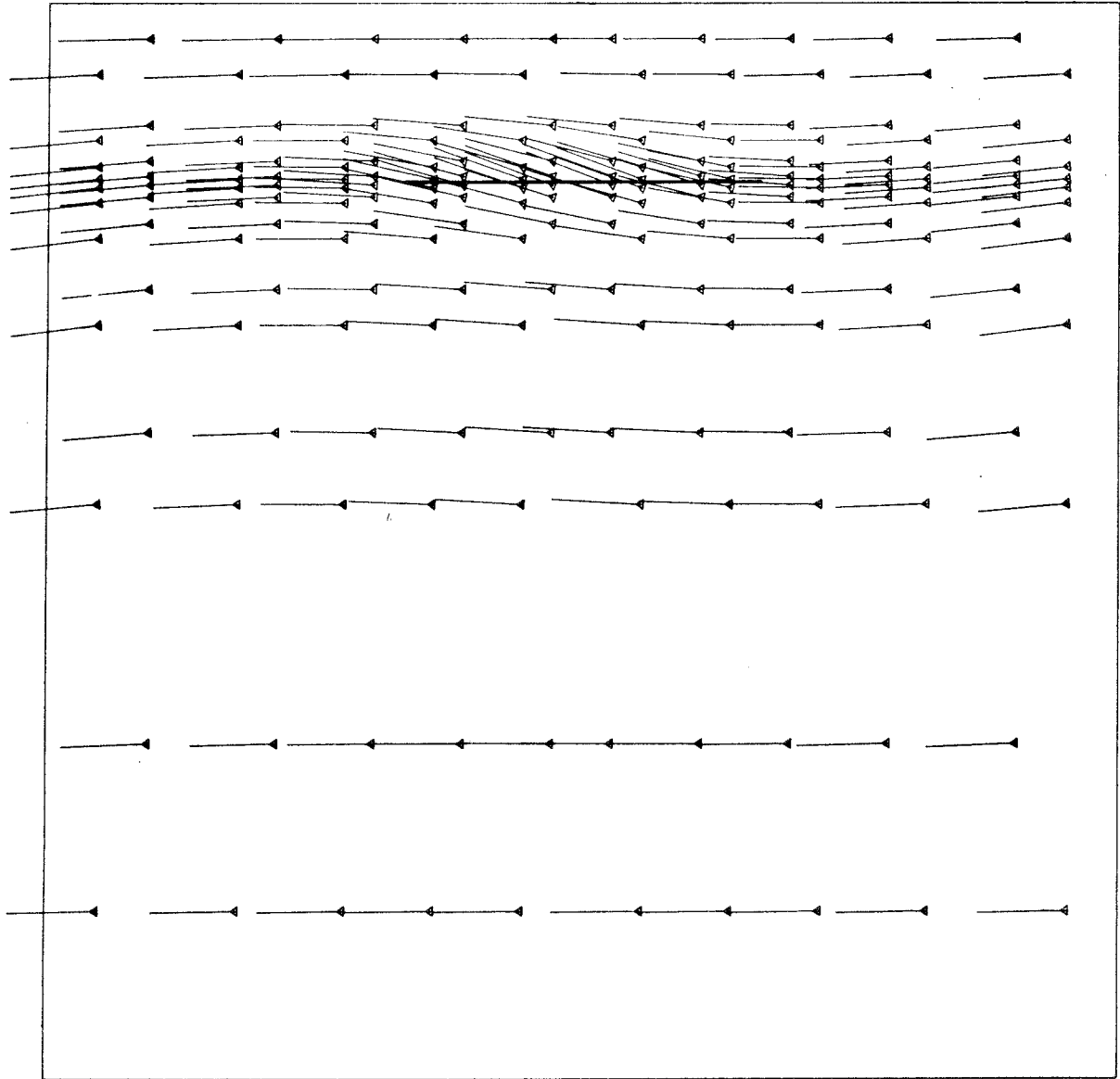K HOMOGENEOUS ISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE   0   500 M

VELOCITY SCALE ⊢——⊣ = 0.5 ✳ 10 ✳✳ −6 M/SEC

40 YEARS
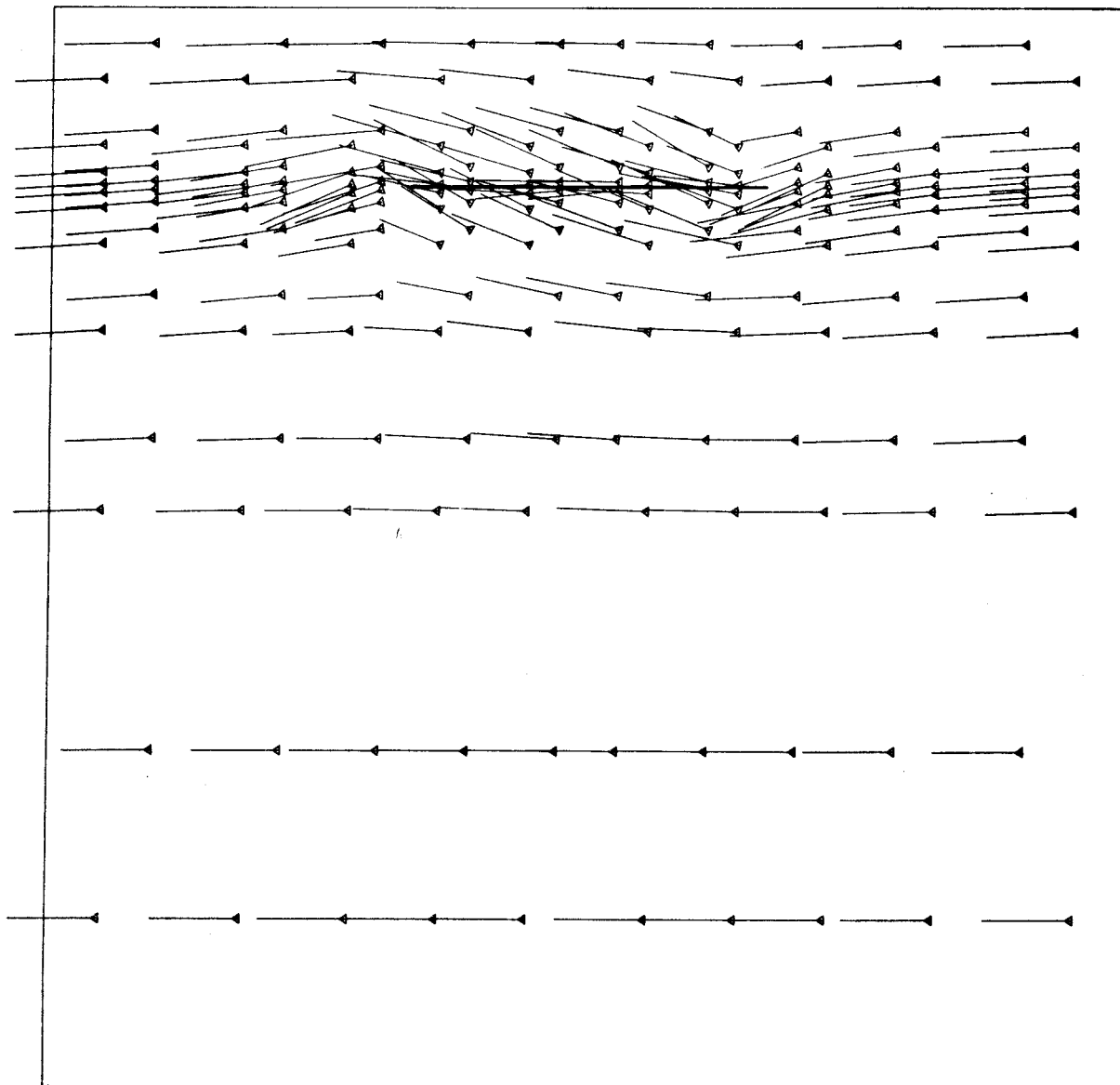
FIGURE 19

KBS HYDROTHERMAL RUN GLOBAL 2 WITH CROSSFLOW
K HOMOGENEOUS ISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE   0 |___ _ _ _| 500 M

VELOCITY SCALE |———| = 0.5 ✱ IO ✱ ✱ -6 M/SEC
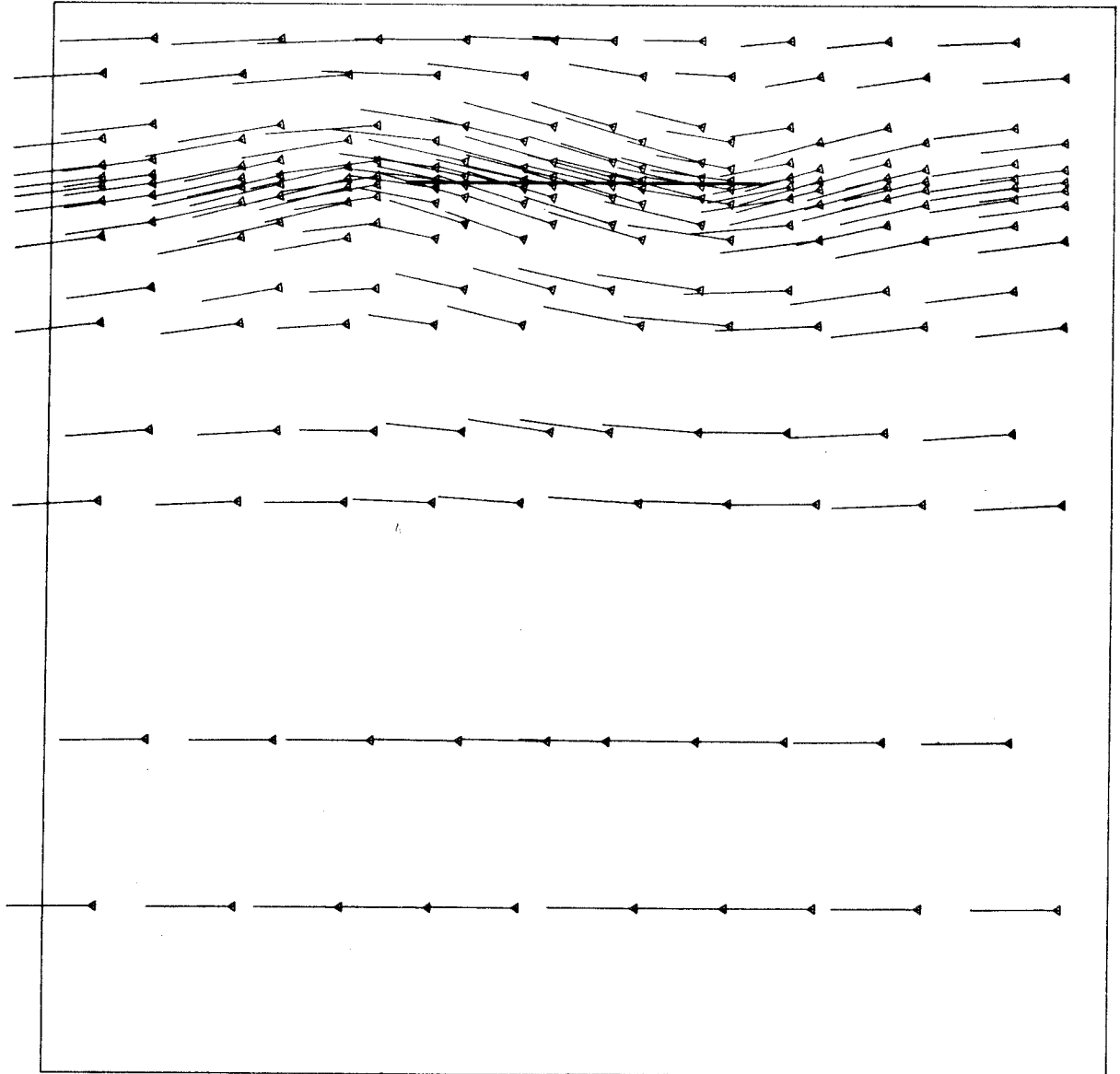
200 YEARS

KBS HYDROTHERMAL RUN GLOBAL 2 WITH CROSSFLOW
K HOMOGENEOUS ISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE   0 ⊢———————┤ 500 M
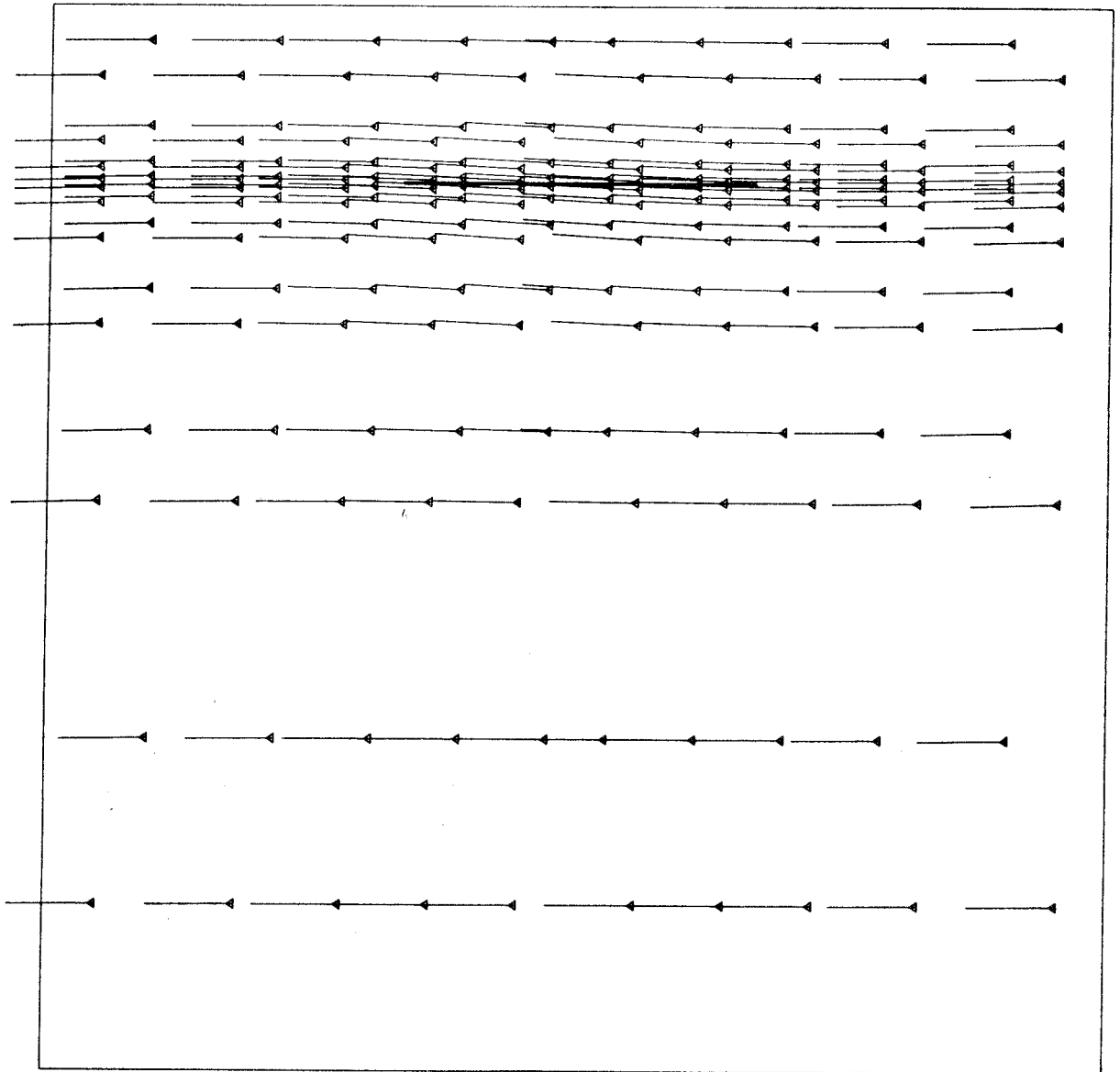
VELOCITY SCALE ⊢———┤ = 0.5 * 10 ** - 6 M/SEC

1000 YEARS

FIGURE 21

KBS HYDROTHERMAL RUN GLOBAL 2 WITH CROSSFLOW
K HOMOGENEOUS ISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE   0 |___.___.___| 500 M

VELOCITY SCALE |————| = 0.5 ✱ IO ✱ ✱ -6 M/SEC

IO OOO YEARS

FIGURE 22

KBS HYDROTHERMAL RUN GLOBAL 2 WITH CROSSFLOW
K HOMOGENEOUS ISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE   0 |___ ___ ___ ___| 500 M
VELOCITY SCALE ├────┤ = 0.5 ✱ 10 ✱✱ - 6 M/SEC

100 000 YEARS
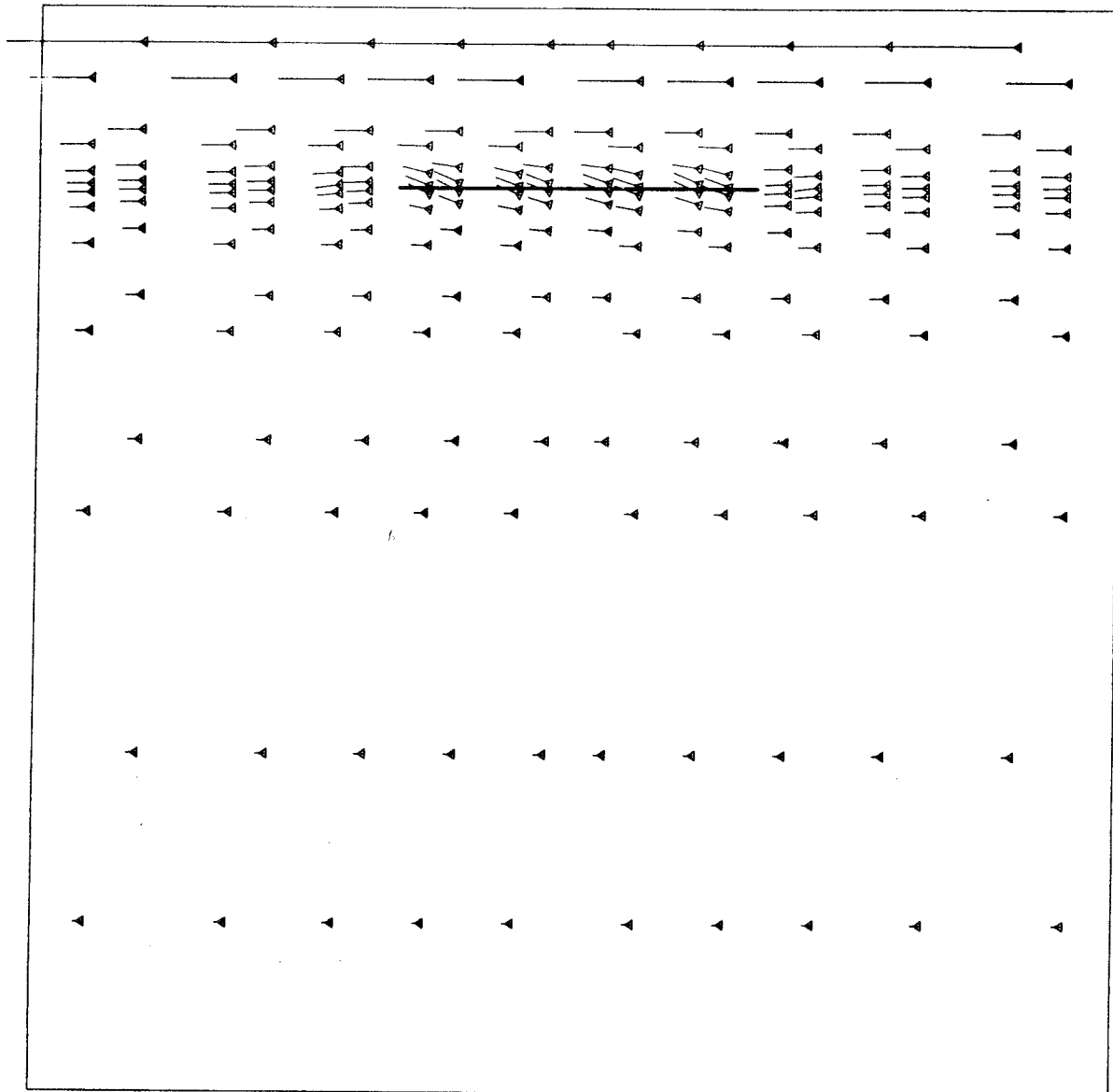
FIGURE 23

KBS HYDROTHERMAL RUN GLOBAL 3 WITH CROSSFLOW
K(Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE   0 _____ 500 M

VELOCITY SCALE ⊢————⊣ = 1.0 ✳ 10 ✳✳ -6 M/SEC

40 YEARS
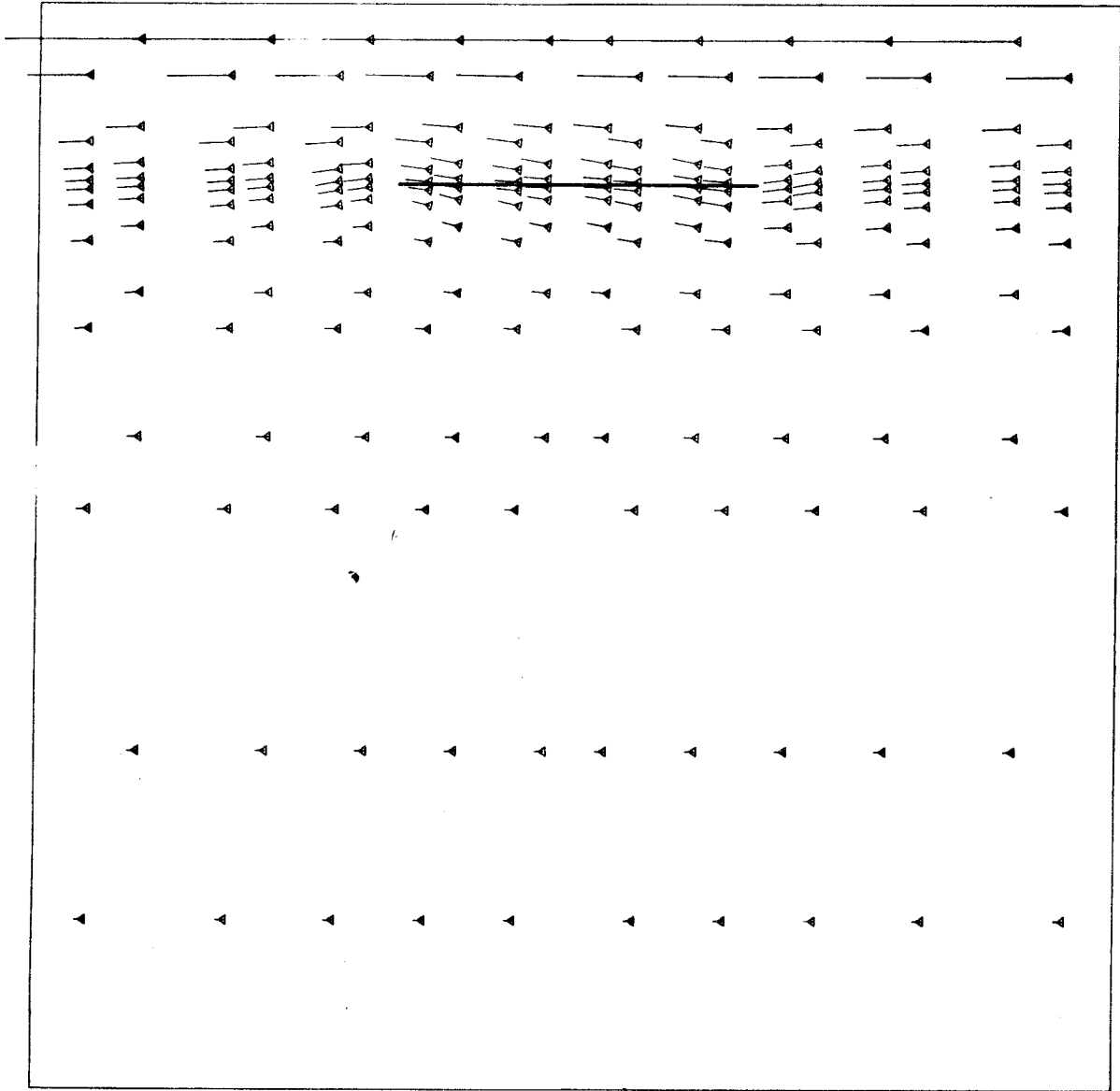
FIGURE 24

KBS HYDROTHERMAL RUN GLOBAL 3 WITH CROSSFLOW
K(Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE $\quad$ 0 $\quad$ 500 M

VELOCITY SCALE $\longmapsto$ = 1.0 * 10 ** -6 M/SEC

200 YEARS
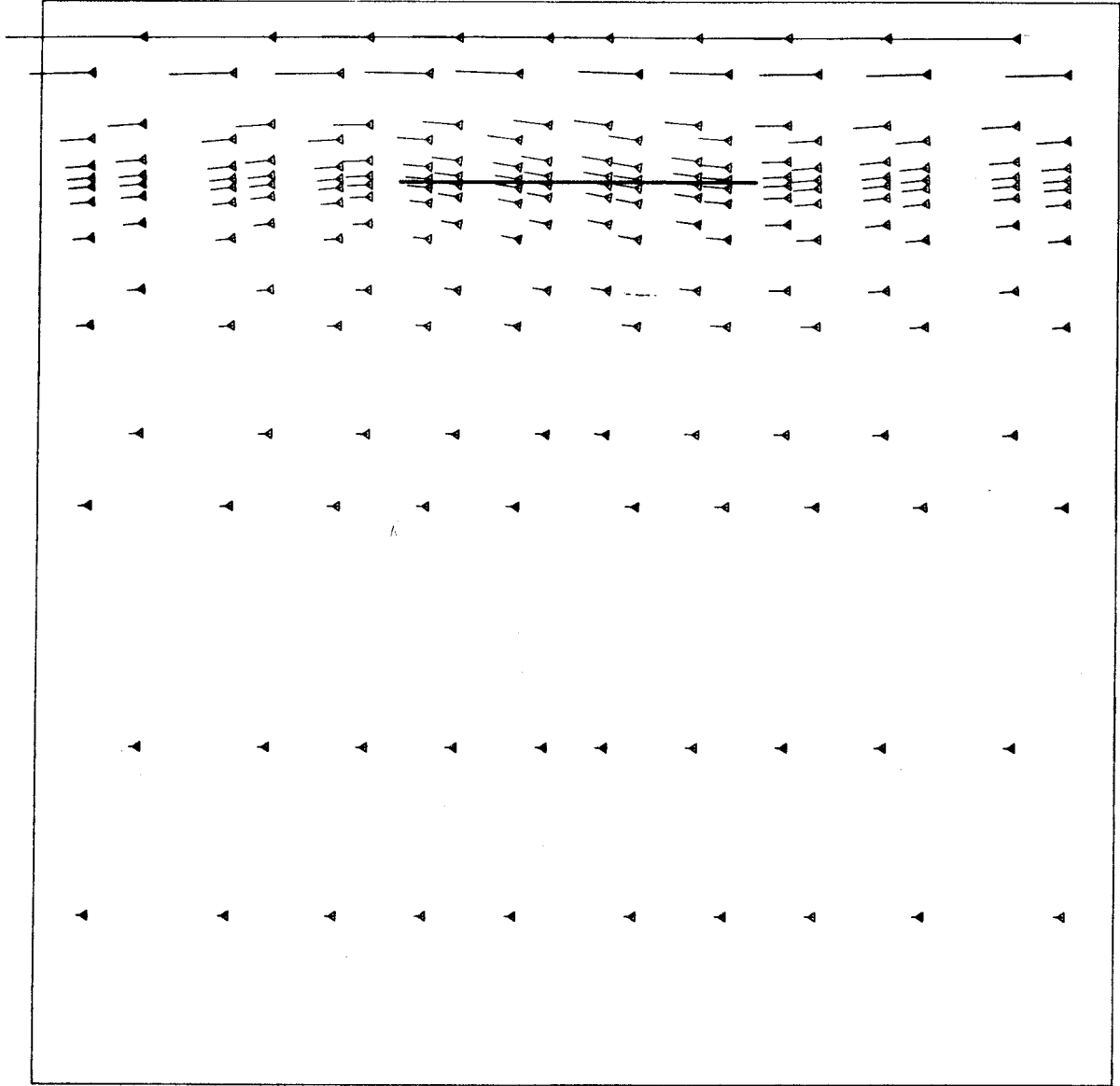
FIGURE 25

KBS HYDROTHERMAL RUN GLOBAL 3 WITH CROSSFLOW
K(Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE   0 ⊢___|___|___| 500 M

VELOCITY SCALE ⊢────┤ = 1.0 $\ast$ 10 $\ast\ast$ -6 M/SEC

1000 YEARS
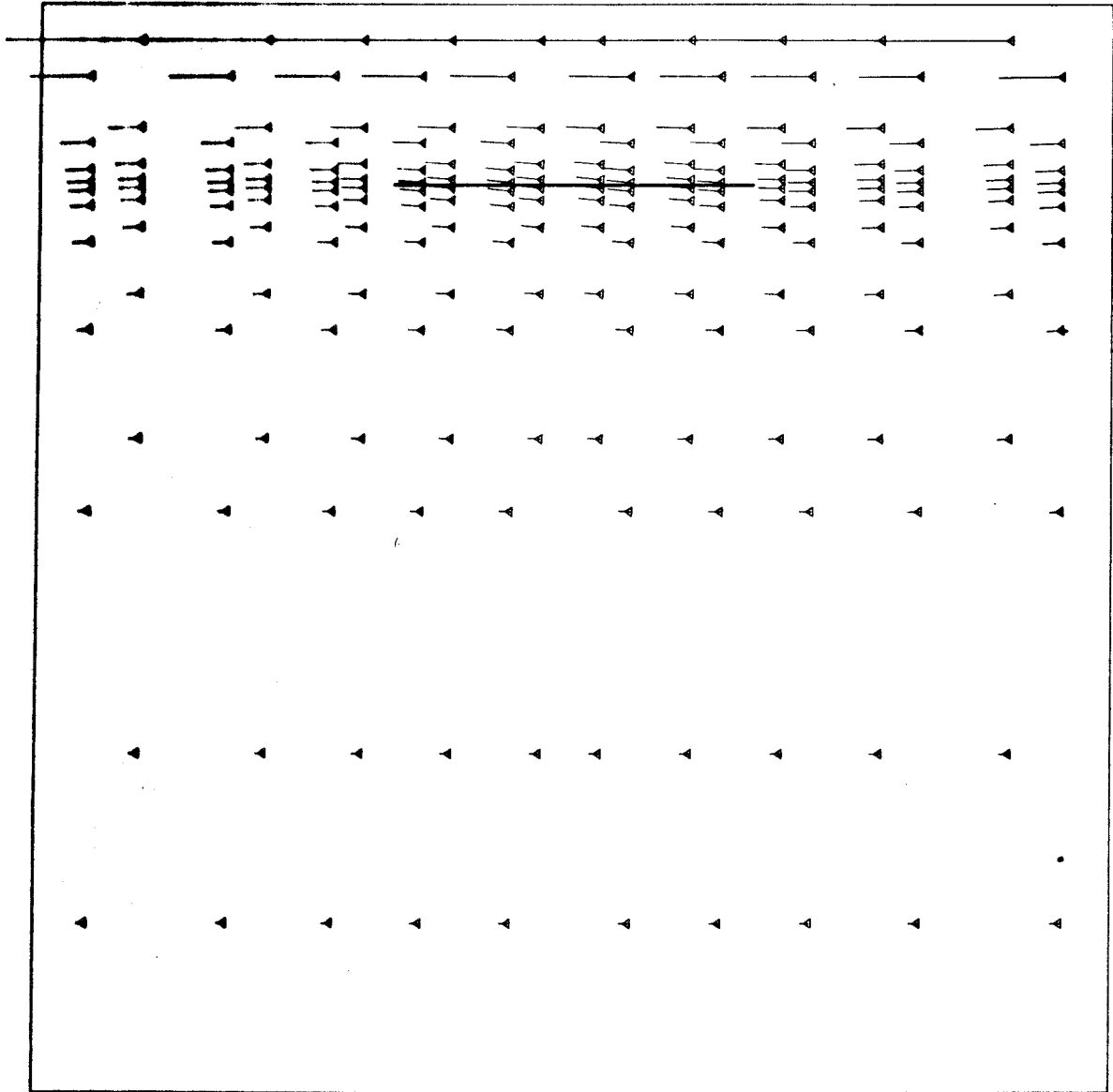
FIGURE 26

KBS HYDROTHERMAL RUN GLOBAL 3 WITH CROSSFLOW
K(Z) ANISOTROPIC

VELOCITY PLOT

GEOMETRY SCALE $\qquad$ 0 ⊢——————⊣ 500 M

VELOCITY SCALE ⊢————⊣ = 1.0 ✳ 10 ✳✳ -6 M/SEC

10 000 YEARS
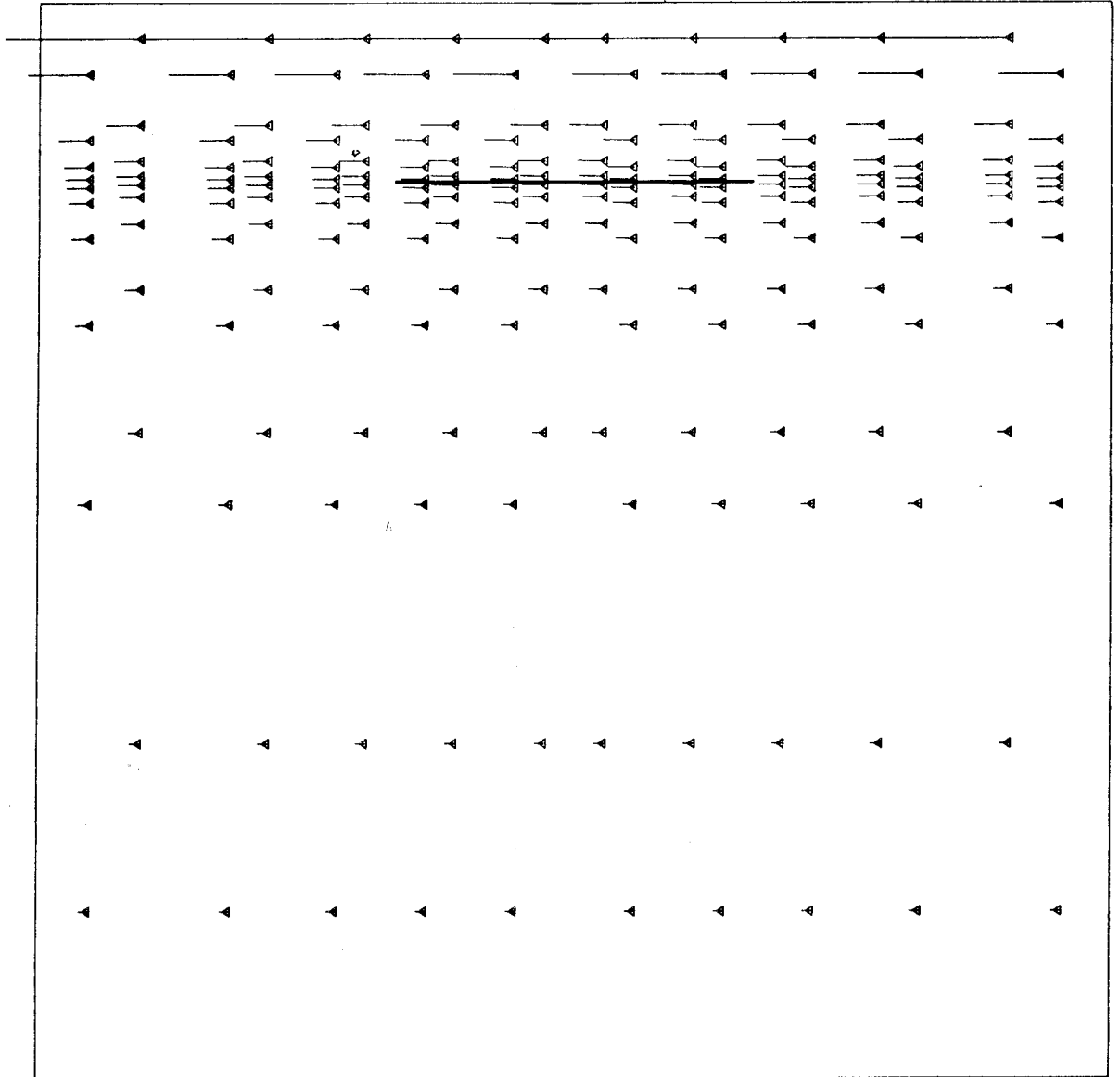
FIGURE 27

KBS HYDROTHERMAL RUN GLOBAL 3 WITH CROSSFLOW
K(Z) ANISOTROPIC

VELOCITY PLOT

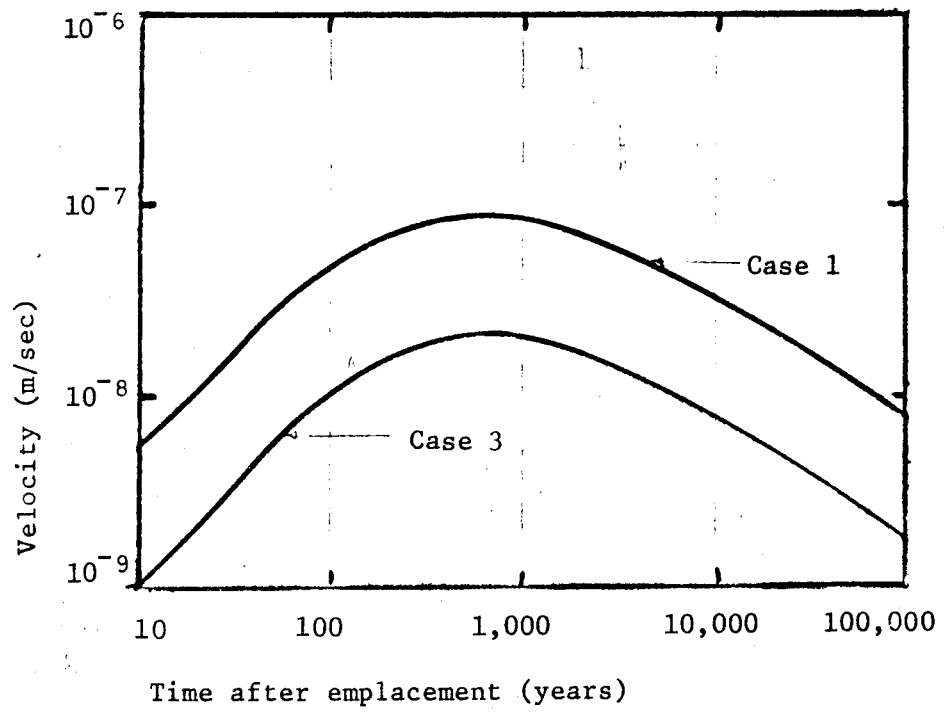GEOMETRY SCALE   0 ⊢——┴——┴——┘ 500

VELOCITY SCALE ⊢———┤ = 1.0 ✳ 10 ✳ ✳ -6 M/SEC

100 000 YEARS

Figure 29. Average hydrothermal vertical velocity above the repository

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      A model study
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      för kapsling av kärnbränsleavfall. Lägesrapport 1977-09-27
      samt kompletterande yttranden.
      Korrosionsinstitutet och dess referensgrupp

32   Long term mineralogical properties of bentonite/quartz
      buffer substance
      Preliminär rapport november 1977
      Slutrapport februari 1978
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      Jan Rennerfelt
      Orrje & Co, Stockholm 1977-11-07

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      Lagar, normer och bedömningsgrunder
      Christina Gyllander
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    kärnbränsle och förglasat högaktivt avfall
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    ett bergförvar
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    AB Atomenergi 1977-10-19

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    Taivo Tarandi
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    Preliminär rapport oktober 1977
    Slutrapport februari 1978
    Lars Y Nilsson
    John Stokes
    Roger Thunvik
    Inst för kulturteknik KTH

48 The mechanical properties of the rocks in Stripa,
    Kråkemåla, Finnsjön and Blekinge
    Graham Swan
    Högskolan i Luleå 1977-09-14

49 Bergspänningsmätningar i Stripa gruva
    Hans Carlsson
    Högskolan i Luleå 1977-08-29

50 Lakningsförsök med högaktivt franskt glas i Studsvik
    Göran Blomqvist
    AB Atomenergi november 1977

51 Seismotechtonic risk modelling for nuclear waste disposal
    in the Swedish bedrock
    F Ringdal
    H Gjöystdal
    E S Hysebye
    Royal Norwegian Council for scientific and industrial
    research

52 Calculations of nuclide migration in rock and porous media,
    penetrated by water
    H Häggblom
    AB Atomenergi 1977-09-14

53   Mätning av diffusionshastighet för silver i lera-sand-bland-
      ning
      Bert Allard
      Heino Kipatsi
      Chalmers tekniska högskola 1977-10-15

54   Groundwater movements around a repository

      54:01   Geological and geotechnical conditions
               Håkan Stille
               Anthony Burgess
               Ulf E Lindblom
               Hagconsult AB september 1977

      54:02   Thermal analyses
               Part 1 Conduction heat transfer
               Part 2 Advective heat transfer
               Joe L Ratigan
               Hagconsult AB september 1977

      54:03   Regional groundwater flow analyses
               Part 1 Initial conditions
               Part 2 Long term residual conditions
               Anthony Burgess
               Hagconsult AB oktober 1977

      54:04   Rock mechanics analyses
               Joe L Ratigan
               Hagconsult AB september 1977

      54:05   Repository domain groundwater flow analyses
               Part 1 Permeability perturbations
               Part 2 Inflow to repository
               Part 3 Thermally induced flow
               Joe L Ratigan
               Anthony S Burgess
               Edward L Skiba
               Robin Charlwood

      54:06   Final report
               Ulf Lindblom et al
               Hagconsult AB oktober 1977

55   Sorption av långlivade radionuklider i lera och berg
      Del 1 Bestämning av fördelningskoefficienter
      Del 2 Litteraturgenomgång
      Bert Allard
      Heino Kipatsi
      Jan Rydberg
      Chalmers tekniska högskola 1977-10-10

56   Radiolys av utfyllnadsmaterial
      Bert Allard
      Heino Kipatsi
      Jan Rydberg
      Chalmers tkniska högskola 1977-10-15