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The Hengill geothermal area, Iceland: Variation of temperature gradients deduced from the maximum depth of seismogenesis

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Abstract

Given a uniform lithology and strain rate and a full seismic data set, the maximum depth of earthquakes may be viewed to a first order as an isotherm. These conditions are approached at the Hengill geothermal area, S. Iceland, a dominantly basaltic area. The likely strain rate calculated from thermal and tectonic considerations is 10^{-15} s⁻¹, and temperature measurements from four drill sites within the area indicate average, near-surface geothermal gradients of up to 150° C km⁻¹ throughout the upper 2 km. The temperature at which seismic failure ceases for the strain rates likely at the Hengill geothermal area is determined by analogy with oceanic crust, and is about $650 \pm 50^{\circ}$ C. The topographies of the top and bottom of the seismogenic layer were mapped using 617 earthquakes located highly accurately by performing a simultaneous inversion for three-dimensional structure and hypocentral parameters.

The thickness of the seismogenic layer is roughly constant and about 3 km. A shallow, aseismic, low-velocity volume within the spreading plate boundary that crosses the area occurs above the top of the seismogenic layer and is interpreted as an isolated body of partial melt. The base of the seismogenic layer has a maximum depth of about 6.5 km beneath the spreading axis and deepens to about 7 km beneath a transform zone in the south of the area. Beneath the high-temperature part of the geothermal area, the maximum depth of earthquakes may be as shallow as 4 km. The geothermal gradient below drilling depths in various parts of the area ranges from $84 \pm 9^{\circ}$ C km⁻¹ within the low-temperature geothermal area of the transform zone to $138 \pm 15^{\circ}$ C km⁻¹ below the centre of the high-temperature geothermal area. Shallow maximum depths of earthquakes and therefore high average geothermal gradients tend to correlate with the intensity of the geothermal area and not with the location of the currently active spreading axis.

1. Introduction

In volcanic and geothermal areas, subsurface heat distribution is related to the distribution of partial melt and reservoir characteristics and is therefore important for volcanic hazard reduction and geothermal exploitation. Unfortunately, few methods can determine temperatures below drilling depth. Some geophysical methods may detect partial melt indirectly, including

An additional method is study of earthquake distribution. At low temperatures and high strain rates rock deforms by brittle fracture and is therefore seismogenic. At high temperatures and low strain rates rock deforms by plastic flow and is aseismogenic. Under intermediate conditions rock is semi-brittle, and

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seismic tomography (e.g., Thurber, 1983), S-wave attenuation mapping (e.g., Einarsson, 1978) and magnetotelluric sounding (e.g., Beblo and Björnsson, 1980), and estimates of the minimum temperature of melt for that lithology may be made.

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Fig. 1. Strength profile of the crust (after Hill, 1992). The maximum strength of rocks in the brittle domain is shown for normal, strikeslip and thrust faulting. In the plastic field, the maximum strength of rock is governed by the power flow law. The laws that determine the maximum strength of rock in the semi-brittle field are not precisely known and the shape of the strength profile there suggested by Scholz (1988) is dashed. Extrapolation of the power law is shown as a dotted line for comparison. T_1 and T_2 are the temperatures at which the least and most refractory minerals enter the plastic domain.

deforms seismically or aseismically depending on the strain rate (Fig. 1). Laboratory experiments and theoretical studies provide insight into the physical conditions under which these processes occur (e.g., Scholz, 1988).

The transition from brittle to plastic deformation, via semi-brittle behaviour, is most strongly dependent on lithology and temperature and weakly dependent on strain rate. In areas where the lithology and strain rate are fairly uniform and the geothermal gradient is positive, the maximum depth of seismicity may be viewed, to a first order, as an isotherm. The topography of the base of the seismogenic layer may then indicate variations in the depth to this isotherm. In active volcanic areas, aseismic volumes embedded in the seismogenic crust may represent isolated volumes of partial melt.

The depth to the base of the seismogenic layer has been studied for several regions and compared with laboratory predictions of the maximum temperature of brittle fracture (e.g., Sibson, 1982; Meissner and Strehlau, 1982). A basic aim of these early studies was to test the theory that the depth to the base of the seis-

mogenic layer could be modelled as a temperaturedependent horizon assuming linear geothermal gradients calculated from surface heat flow or extrapolations of borehole measurements. Sibson (1982) found that, for several non-subduction parts of the Western United States, the depth to the base of the seismogenic layer correlated strongly with the predicted depth of transition from frictional sliding to quasi-plastic flow in rocks containing quartz. Meissner and Strehlau (1982) studied areas in China, Greece, Germany, California and the Basin and Range Province in the Western United States. Assuming linear geothermal gradients, they modelled the depth distribution of earthquakes as being temperature-controlled in a wet upper crust overlying dry lower crust. They concluded that stress in the crust was built up by creep from below and that this, together with the increase in rock strength in the brittle layer with depth, resulted in the largest earthquakes nucleating near to the base of the seismogenic layer.

Hill (1992) studied the maximum depths of earthquakes at Long Valley Caldera, California and the Phlegrean Fields Caldera, Italy. The Long Valley Caldera area is dominantly quartzo-feldspathic, for which lithology a temperature for the base of the seismogenic layer of 250-500°C is predicted for reasonable estimates of water content and local strain rate. The maximum depth of earthquakes varies from less than 4 km beneath the active resurgent dome and Mammoth Mtn. to 9 km beneath distal areas. No measurements of geothermal gradients were available, but the maximum depths of seismicity were consistent with hypothetical thermal models of the area. The Phlegrean Fields Caldera area is dominantly feldspathic, and temperatures of 600-800°C are predicted for the base of the seismogenic layer, which is at a depth of 4-5 km. Data from a 3-km-deep borehole indicated a geothermal gradient of about 100°C km⁻¹ in the shallow crust, and these results therefore suggest an increase in geothermal gradient at depths greater than 3 km.

Kong et al. (1992) and Toomey et al. (1988) studied the maximum depth of seismicity beneath the mid-Atlantic ridge (MAR) at 26°N and 23°N respectively. Large and small maximum hypocentral depths correlated with low and high MAR topography respectively, findings that are consistent with shallow high-temperature material being associated with large surface extrusion rates.



Fig. 2. Map of the Hengill area showing geologic features and place names mentioned in the text. The three volcanic systems are shown schematically with the fissure swarms outlined and the Grensdalur central volcano and Mts. Hengill and Hromundartindur dashed. Inset shows location of the area within Iceland.

The presence of aseismic volumes in the seismogenic crust may provide supporting evidence for partial melt. At Kilauea, Hawaii, a summit magma chamber is known to exist from frequent volcanic tremor, eruptions, earthquake swarms and cyclic ground deformation. Thurber (1984) detected a low-velocity volume at 2–4 km depth beneath the summit using seismic tomography. This volume is aseismic and is embedded in the seismogenic crust, suggesting that it contains partial melt.

The Hengill area in Iceland (Fig. 2) contains extensive geothermal resources, several volcanic units and a currently active spreading center (e.g., Foulger and Toomey, 1989). Temperature gradients from over 30 boreholes up to 2,265 m deep are available (e.g., Hersir et al., 1990) and an extensive database of local earthquakes accurately located using a three-dimensional, tomographically-derived velocity model. Exceptionally good data are thus available to study variations in the depths to the top and bottom of the seismogenic layer there. In this paper, I assess the ambient strain rate, and the predicted temperature at which the crust ceases to be seismogenic. I map the top and bottom of the seismogenic layer using the seismic database and interpret the results in terms of structure and local variations in geothermal gradient of this part of the currently active accretionary plate boundary, the transform zone and the associated geothermal area.

2. Rock rheology

2.1. Theory

Stresses in the Earth cannot exceed the strength of rocks (Brace and Kohlstedt, 1980) and this parameter thus governs stress release by rock failure and earthquake occurrence. Laboratory experiments show that bulk rock strength depends strongly and non-linearly on mineralogy, lithology, temperature, pressure, the presence of preexisting fractures, fluid content and pore pressure, and more weakly on strain rate (e.g., Kirby and Kronenberg, 1987).

At small to moderate depths, where temperatures are low, deformation occurs primarily as seismic frictional sliding on faults, with some aseismic deformation occurring within weak zones e.g., fault gouge. At large depths, where temperatures are high, deformation occurs by aseismic plastic flow. Between these two fields is a transition zone, where deformation is governed by a mixture of these two processes and which represents the base of the seismogenic zone (Scholz, 1988). The relative thinness of the zone separating the surface, brittle layer from the deeper, plastic layer results in plate-like behaviour of the surface layer, with stress buildup at its base caused by creep of the material below. Large earthquakes nucleate at the base of the brittle layer as a result (e.g., Meissner and Strehlau, 1982).

In the brittle layer, failure is pressure-dependent but independent of lithology and temperature (Byerlee, 1978). It approximately follows Byerlee's Law (Byerlee, 1978):

 $\tau = \mu \sigma_n'$

where τ and σ_n' are the shear and effective normal stresses acting on the fault surface, and μ , the coefficient of static friction, is approximately 0.75. The effective normal stress, σ_n' , is governed by the law:

 $\sigma_{n}' = \sigma_{n} - P$

where σ_n is the applied normal stress and *P* is pore fluid pressure. Assuming hydrostatic fluid pressures, and incorporating a geometrical factor to account for different faulting modes (normal, strike-slip or thrust, e.g., Sibson, 1982) the minimum shear stress needed to cause failure is a function of depth for different tectonic regimes (Fig. 1).

At high temperatures, rock deforms by bulk plastic flow and its strength is highly dependent on temperature and lithology and more weakly dependent on strain rate as indicated by the power law:

$$\tau = \left[\left(\frac{\dot{\epsilon}}{A}\right) \exp\left(\frac{H^*}{RT}\right)\right]^{1/n}$$

where $\dot{\epsilon}$ is the strain rate, A and n are constants, H^* is the activation energy, R is the universal gas constant $(8.31 \times 10^{-3} \text{ kJ}^{\circ}\text{K} \text{ mole}^{-1})$ and T is absolute temperature (e.g., Sibson, 1982). This equation predicts that strength falls off very rapidly with increasing temperature (Fig. 1). Independent information is required to relate temperature to depth in order to predict the strength of rocks at varying depths within the Earth.

At intermediate temperatures, rock is neither completely brittle nor plastic and its deformation involves a combination of frictional sliding and plastic flow (Paterson, 1978) and is pressure-sensitive. This is known as the semi-brittle field or the brittle-plastic transition (Scholz, 1988). In contrast to the velocityweakening that characterises the brittle field, velocitystrengthening occurs in the semi-brittle field, rendering it aseismogenic (Tse and Rice, 1986; Scholz, 1988). Where the thermal gradient is positive, this zone lies between the shallow, brittle, and the deep, plastic layers. The peak of rock strength lies within this zone and ruptures associated with large earthquakes nucleating near the base of the brittle layer may extend down through the semi-brittle layer to the level of maximum strength. The upper part of the semi-brittle field therefore deforms by coseismic dynamic slip and interseismic plastic flow (Scholz, 1988).

The upper boundary of the semi-brittle field occurs where the physical conditions are such that the most susceptible mineral enters the plastic domain. The lower boundary occurs at conditions where sufficient minerals have become plastic to enable plastic behaviour of the bulk rock.

2.2. Complicating factors

Flow laws determined by laboratory experiments have large uncertainties as a result of large data scatter. In addition, application of the results to the Earth is problematic because it is unclear which lithology best models rock at great depth. Furthermore, the behaviour of bulk rock in situ may deviate from that of laboratory specimens because of factors which are usually quantitatively unassessable:

(1) The presence of water decreases the temperature of onset of plasticity in rocks (Kirby, 1983; Scholz, 1988), e.g., by 150–200°C for granite. The depth of water circulation in the crust is not well-known, but evidence from mines, boreholes and conductivity measurements suggest that water-filled, interconnected pore space exists down to depths of several km (Brace and Kohlstedt, 1980). The maximum depth of water circulation may be the level at which plastic flow begins, since plastic creep would work to "heal" cracks. The semi-brittle and plastic layers may contain water, but have low permeabilities.

(2) Abnormal fluid pressures and preexisting fractures decrease strength in the brittle layer.

(3) Dislocation creep processes may be important in the semi-brittle and plastic layers as indicated by microstructural and textural evidence from naturally deformed xenoliths and massifs (Kirby and Kronenberg, 1987).

(4) The presence of non-hydrous fluids, e.g., CO_2 would modify behaviour. (5) Creep mechanisms may be sensitive to grain size (Kirby and Kronenberg, 1987).

2.3. Application to the behaviour of the Earth

Despite large uncertainties in extrapolating laboratory measurements to bulk rocks, *in situ*, the results have been found to be consistent with those of other studies. The thickness of the high-velocity, low-attenuation oceanic lithosphere determined from explosion seismology is consistent with plate thicknesses calculated from cooling models (Parsons and Sclater, 1977), which provides evidence that lithospheric thickness is thermally controlled. The maximum depths of intraplate oceanic earthquakes and those resulting from bending of the lithosphere are limited by the 700°C isotherm (Parsons and Sclater, 1977; Bodine et al., 1981; Wiens and Stein, 1983; Kirby, 1983). The model of an elastic plate, overlying material whose viscosity decreases with depth according to a power law, successfully models crustal necking and surface topography in continental extension regimes, e.g., the Basin and Range Province, U.S.A. (Fletcher and Hallet, 1983). The maximum depths of earthquakes in parts of the U.S.A., Europe and Asia are consistent with temperature predictions assuming uniform geothermal gradients (Sibson, 1982; Meissner and Strehlau, 1982; Hill, 1992). These diverse studies suggest that thermal models and the results of laboratory studies may be applied quasi-quantitatively to maximum earthquake depths.

3. Tectonics of Iceland and the Hengill area

Iceland is the largest subaerial exposure of accretionary plate boundary in the world, with over 700 km of spreading plate boundary and two transform zones exposed. The accretionary plate boundary is comprised of *en echelon* volcanic systems which are made up of swarms of dykes, fissures and normal faults and usually a central volcano associated with a high-temperature geothermal area (>200°C in the upper 1 km) (e.g., Sæmundsson, 1979). This suggests the presence of shallow, long-lived, fractionating magma chambers. These systems are thought to be analogous to oceanic spreading segments.

The Hengill area is a triple junction and is the meeting point of the Reykjanes Peninsula Volcanic Zone, the Western Volcanic Zone and the South Iceland Seismic Zone (Fig. 2). Extensive research for geothermal prospecting has elucidated well the structure and history of the area (e.g., Sæmundsson, 1967; Hersir, 1980; Torfason et al., 1983; Stefansson et al., 1983; Hersir et al., 1984, 1990; Steingrimsson et al., 1986; Arnason et al., 1987; Foulger, 1988a, b; Foulger and Toomey, 1989; Walker, 1992). Three parallel volcanic systems occur there, of which the Grensdalur and Hromundartindur systems are almost extinct as a result of recent migration of the locus of spreading to the currentlyactive Hengill system (Fig. 2). This system is dominated topographically by the 800-m-high Mt. Hengill, which is comprised largely of a single hyaloclastite formation, and therefore does not represent a polygenetic eruptive site.



Fig. 3. Map showing the epicentres of earthquakes used in this study, lines of the cross sections shown in Fig. 5 and locations of the geothermal maxima detected by fumarole gas geochemistry. The low-velocity body detected by seismic tomography beneath the NE part of Mt. Hengill is shaded. Schematic tectonic features are shown as for Fig. 2.

Seismic tomography and gravity studies provide no evidence for a substantial, shallow magma body beneath Mt. Hengill, though a small low-velocity body in the depth range 2-4 km beneath the NE flank of Mt. Hengill may represent an isolated body a few km³ in volume of up to 7% partial melt (Fig. 3) (Foulger and Toomey, 1989; Foulger and Arnott, 1993). The most recent magmatic activity is thought to be a dyke injection episode in 1789. No surface volcanism occurred then but contemporary accounts suggest that an intense earthquake swarm affected 30 km of the fissure swarm from Mt. Hengill north to Thingvellir, where at least 60 cm of subsidence occurred (Palsson, 1945). The lack of a mature central volcano is a characteristic that the Hengill system shares only with the systems of the Reykjanes Peninsula Volcanic Zone. These systems are probably transitional in nature between typical oceanfloor spreading segments and typical Icelandic ones.

Geothermal resources are widespread throughout the whole area and surface heat loss induces continuous, small magnitude earthquake activity as a result of thermal contraction and cracking in the heat sources (Fig. 3) (Foulger and Long, 1984; Foulger, 1988b). A hightemperature geothermal area encompasses the three volcanic systems (Fig. 2). Geochemical analysis of fumarole gasses, and the distribution of the thermalcracking earthquakes, indicate that separate heat sources underlie each of the three volcanic systems (Fig. 3). A magnetotelluric measurement made at Nesjavellir, 3 km north of Mt. Hengill, detected a highconductivity layer, that is interpreted as a layer of partial melt, a few km thick at a depth of 7.5 km (Hersir et al., 1990). A low-temperature geothermal area (<150°C in the upper 1 km) is associated with the transform branch in Ölfus, the southern part of the area (Fig. 2) and is thought to result from circulation of water through deep faults.

4. Strain rate and geothermal gradient in the Hengill area

4.1. Strain rate

No measurements of strain are available from the Hengill area, and therefore the strain rate may be estimated only. Processes that cause crustal strain in this area include volumetric contraction of the geothermal heat source and crustal extension spanning the currently-active spreading axis in the Hengill system. Both these processes will cause crustal extension.

The strain rate resulting from thermal contraction of the heat sources may be calculated from the volumetric contraction rate:

$$strainrate = \frac{\text{volume contraction}}{3 \times \text{total volume}} = \frac{H\gamma}{3VC_{p}\rho}$$

where *H* is the heat power output, γ is the coefficient of thermal expansion for basalt ($\approx 16.2 \times 10^{-6} \text{ K}^{-1}$), *V* is the total volume of cooling rock, C_p is the specific heat of basalt ($\approx 1.3 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$) and ρ is the rock density ($\approx 3 \times 10^3 \text{ kg m}^{-3}$).

The heat loss over the 70-km² high-temperature geothermal area is approximately 350 MW (Bodvarsson, 1951). In order for an estimate of the volumetric contraction rate to be made, the thickness of the cooling layer must be estimated. The continuous, small magnitude earthquake activity of the Hengill area is thought to be generated by cooling-contraction cracking and thus to indicate the extent of the cooling volume (Foulger, 1988b). The present study shows that the thickness of the seismogenic layer is fairly constant and about 3 km throughout the area (see below). Assuming an average thickness of 3000 m for the cooling layer, the strain rate may be calculated to be approximately 3×10^{-15} s⁻¹.

The time-averaged crustal extension rate in Iceland is approximately 2 cm a^{-1} (DeMets et al., 1990) and the width of the zone of strain accumulation about 200 km (Heki et al., 1993). Crustal spreading considerations thus suggest a strain rate of the order of 10^{-15} s⁻¹. This estimate is very similar to that made using cooling considerations.

4.2. Geothermal gradients

Substantial drilling for geothermal resources has been conducted in the area. Over 20 relatively deep (over several hundred meters) wells have been drilled at the Nesjavellir site north of Mt. Hengill including the deepest well in any Icelandic high-temperature geothermal area which is 2,265 m deep (Fig. 2). A single well approximately 900 m deep has been drilled at the Sleggjubeinsdalur site, south of Mt. Hengill. Eight wells 300–1,000 m deep have been drilled at the Reykjakot site, in the Grensdalur system, and wells up to approximately 1 km deep have been drilled in Ölfus to the south. The temperature gradients measured are controlled primarily by hydrothermal circulation, are very variable from site to site, and from well to well at a single site.

At the Nesjavellir site, temperatures of approximately 300°C are reached at about 2 km depth beneath the hottest, southernmost part of the well site. This decreases to about 250°C in wells to the north. Geothermal gradients vary from over $300°C \text{ km}^{-1}$ in the upper few hundred metres to about $70°C \text{ km}^{-1}$ at greater depths. The zeolite mineralogy of borehole shards indicates that the reservoir is not in thermal equilibrium but is cooling in some areas and heating up in others (Franzson, 1988).

At Sleggjubeinsdalur the average gradient in the upper 1 km is about 205° C km⁻¹ (Steingrimsson et al., 1990). At the Reykjakot site boreholes reveal very high near surface geothermal gradients with temperatures approaching 200°C in the upper 100–200 m. In all the boreholes deeper than 300 m a temperature inversion occurs at 100–500 m depth, indicating that the high

surface temperatures result from lateral flow of hot fluids, and at a depth of about 0.75 km the temperature varies from about 180–225°C, increasing to the north (Xi-Xiang, 1980). In the low-temperature area in Ölfus, geothermal gradients of about 150°C km⁻¹ have been measured in the upper 1 km.

Three temperature maxima are detected by fumarole gas geochemistry in the high-temperature geothermal area and have temperatures up to 310°C, 300°C and 270°C (Torfason et al., 1983) (Fig. 3). The depths to these temperatures are unknown, so geothermal gradients cannot be calculated. However, these findings indicate that three distinct heat sources fuel the high-temperature geothermal area.

A tomographically-imaged, low-velocity zone occurs at 2-4 km depth beneath the north edge of Mt. Hengill (Toomey and Foulger, 1989; Foulger and Toomey, 1989) (Fig. 3). If this contains partial melt, minimum temperatures of about 1150°C must occur there. Similarly, the high conductivity layer detected by a magnetotelluric measurement at 7.5 km depth beneath Nesjavellir must also reach this temperature if it contains partial melt (Hersir et al., 1990).

5. Temperature of the base of the seismogenic layer in the Hengill area

Previous studies have estimated the temperature of the base of the seismogenic layer by downward extrapolation of geothermal gradients calculated from surface heat flow or measured in shallow boreholes (e.g., Sibson, 1982; Meissner and Strehlau, 1982; Hill, 1992). This approach is based on the assumption of conductive heat flow and i s inapplicable to the Hengill area since:

(1) Boreholes up to about 2 km deep within the area show that geothermal gradients are extremely variable with depth, and controlled by hydrothermal circulation, and

(2) Heat-loss considerations suggest that geothermal gradients may be severely non-linear immediately around high-temperature heat sources, e.g., magma bodies (Björnsson et al., 1980).

For these reasons, we examine laboratory studies of rock rheology and theoretical work to estimate the temperature of the base of the seismogenic layer, and calculate the average geothermal gradient between that horizon and the maximum depth of drilling. This con-

Table 1Flow-law parameters for diabase

A (MPa ^{-1/n} s ⁻¹)	H* (kJ/mol)	n	Reference	
2.2×10^{-04}	260	3.4	Shelton and Tullis (1981)	
6.2×10^{-02}	276	3.0	Caristan (1982)	

trasts with the more usual approach of using the maximum depth of earthquakes to test the hypothesis that this horizon is temperature controlled under the assumption that the geothermal gradient is fairly constant throughout the crust.

The lithology of the Hengill area is dominantly basaltic and likely to behave in a similar way to diabase. The results of only two laboratory studies of the flow law of diabase are available (Table 1) (Shelton and Tullis, 1981; Caristan, 1982). Stress drops associated with earthquakes are generally greater than 10 bar, so seismicity will not occur where rock strength is less than this. Using this assumption, the flow laws given in Table 1 predict a temperature of the base of the seismogenic layer of about 885°C (Shelton and Tullis, 1981) or 775°C (Caristan, 1982) for strains of approximately 10^{-15} s⁻¹.

These estimates are exceptionally high when compared with estimates of the limiting temperature of the mechanically strong lithosphere. Comparisons of the depths of oceanic, intraplate earthquakes (Bodine et al., 1981) with theoretical thermal models (Parsons and Sclater, 1977) suggest that the rheology of the oceanic lithosphere is controlled by that of dry olivine and that earthquakes do not occur at temperatures higher than 700°C and may cut off in the temperature range 600-700°C (Kirby and Kronenberg, 1987; Kirby, 1983; A. Kronenberg, pers. commun., 1994). The discrepancy in these values and those suggested by the flow laws for diabase may be attributed to the several complicating factors listed above. In particular, the presence of water in the rock in situ could account for the difference.

For the current study, a temperature of the base of the seismogenic layer of $650 \pm 50^{\circ}$ C is used, since the weight of evidence suggests that this is realistic for bulk rocks *in situ*. It is apparent from the discussion above, however, that this temperature is poorly known, and may even vary over the area if the amount of water present, the strain rate, or several other parameters vary. The absolute geothermal gradients calculated below on the basis of a temperature of $650 \pm 50^{\circ}$ C should therefore be viewed with caution. More confidence may be placed in the *variations* in gradient over the area, however, which reflect the variations in depth to the base of the seismogenic layer.

6. Seismicity of the Hengill area

The area is highly seismically active on a daily basis at a rate of about 1 magnitude 0 event per day (Icelandic local magnitude). The activity was studied in detail in 1981 when 23 temporary seismic stations were deployed for 3 months. Over 2000 locatable earthquakes were recorded, an average of 20 per day (Foulger, 1988a). The epicentral distribution correlated positively with surface heat loss and this, coupled with the discovery that many events had tensile-crack type focal mechanisms, led to the hypothesis that the earthquakes represented contraction cracking in the heat sources of the geothermal area.

A subset of the best-recorded earthquakes was used to perform a simultaneous inversion for three-dimensional structure and hypocentral parameters (Thurber, 1983; Toomey and Foulger, 1989; Foulger and Toomey, 1989). The improved earthquake locations, including hypocentral depths, are accurate to better than a few 100 m, and these refined locations were used for the present study (Fig. 3). Poorly-recorded earthquakes were eliminated from the data set and the most accurately located only are used here. The simultaneous inversion did not extend into the transform zone and the best earthquake locations there are those obtained using a one-dimensional crustal model (Foulger, 1988a). The hypocentral depths of those events may be in error by up to 1 km.

7. Mapping the topography of the seismogenic layer

The events were sorted according to epicentral location into subsets underlying square areas 0.5 km on a side. The topographies of the top and the base of the seismogenic layer were then mapped using the method of minimum curvature under tension (Smith and Wessel, 1990). This method interpolates the data with a surface having continuous second derivatives and minimum total squared curvature. It approximates the shape of a membrane under tension flexed to pass through the data points, and is a reasonable quantitative way of contouring irregularly-distributed data points that yields a surface that fits the data but does not contain unreasonably sharp gradients. Areas unconstrained by close data points should nevertheless be viewed with skepticism. In addition, it should be borne in mind that an incomplete data set will result in underestimates of the maximum, and overestimates of the minimum depths of seismicity. Areas where large numbers of events are located are the most reliably constrained.

The depth to the 90th percentile event (or calculation of that depth where fewer than 10 earthquakes lay within the subset) was used to avoid bias by small numbers of outliers (Fig. 4). The depth to this horizon varies from about 4 to 6 km within the volcanic complex. Within the Grensdalur volcano, the seismicity is of generally of moderate depth and slightly deeper (≈ 5 km) in the east and west than in the north and south $(\approx 4 \text{ km})$ (Figs. 4 and 5a). Within the Hromundartindur system, the activity is relatively deep beneath the south part of Mt. Hromundartindur but shallows to ≈ 4 km beneath the intensely seismic Klambragil (Figs. 2, 4, 5a and 5b). The largest depths to the base of the seismogenic layer lie within the Hengill fissure swarm south of Mt. Hengill and in a zone extending from the north part of Mt. Hengill into the fissure zone to the north (Figs. 4 and 5c). The activity is shallow in the most northerly and southerly parts of the Hengill fissure zone studied, and is shallowest beneath the southern part of Mt. Hengill.

The depth to the top of the seismogenic layer is calculated as the depth to the 10th percentile event (Fig. 6). The topography of this horizon is highly correlated with that of the base of the seismogenic layer except beneath the north of Mt. Hengill, the fissure swarm north of this, and the western part of the Grensdalur volcano. In these areas, relatively large depths to the base of the seismogenic layer is relatively deep (≈ 4 km) lies in the north part of Mt. Hengill and within the fissure zone to the NE of this. This zone coincides with a low-velocity body imaged in the depth range 2–4 km using local earthquake tomography (Figs. 5c and 6) (Foul-



Fig. 4. Map showing the depth to the base of the seismogenic layer within the area studied using tomography (Toomey and Foulger, 1989; Foulger and Toomey, 1989), calculated as the depth to the 90th percentile event. Schematic tectonic features and geothermal maxima are shown as for Fig. 3.

ger and Toomey, 1989). This low-velocity body is therefore aseismic. A neighboring high-velocity body directly beneath Mt. Hengill is characterised by shallower depths to the top of the seismogenic layer (≈ 3 km) and is therefore seismic (Figs. 5c and 6).

Seismicity extends to about 7 km depth beneath the transform zone (Foulger, 1988a). However, the data are insufficiently accurate to detect variations in maximum depth within this zone.

8. Discussion

8.1. The thermal structure of the crust

Constraints other than the maximum depth of seismicity

The continuous, small-magnitude seismicity of the Hengill area is thought to arise from thermal contraction in cooling heat sources (Foulger and Long, 1984; Foulger, 1988b) and the current study suggests that this process occurs at temperatures up to $650 \pm 50^{\circ}$ C. Seismicity does not in general extend to the surface, perhaps because the rocks are weak and jointed there and thus cannot accumulate strain energy. In addition to the depth to the base of the seismogenic layer, the thermal state of the crust is constrained by:

(1) The presence of partial melt, suggested by a tomographically-detected low-velocity zone,

(2) Measurements of geothermal gradients in boreholes. These have been conducted at four sites within the area, and

(3) Temperature maxima within the geothermal area from fumarole geochemistry.

Partial melt in the crust

The low-velocity body detected by tomography beneath and to the northeast of Mt. Hengill may contain up to 7% partial melt (Foulger and Toomey, 1989; Foulger and Arnott, 1993). It is elongated parallel to the fissure swarm (Fig. 6) and is close to the Nesjavellir drilling site where exceptionally high temperatures have been encountered at shallow depth and where zeolite mineralogy suggests that the geothermal area has been recently heated (Franzson, 1988). The most recent magmatic activity in the area is thought to be a dyke injection event from Mt. Hengill along the fissure swarm to the north for about 30 km in 1789, and this partial melt body may have been activated or created then.

The current study shows that the low-velocity volume is aseismic but that earthquakes occur beneath it (Fig. 6) and in close juxtaposition laterally (Fig. 5c). This strengthens the case for partial melt there. Since basalt melts at approximately 1150°C, and seismicity does not occur above 650 ± 50 °C, large thermal gradients may surround this body.

There is good correlation between this body and a zone of relatively large depth to the top of the seismogenic layer (Fig. 6). This illustrates how a map of the top of the seismogenic layer indicates areas where there is room above the seismic layer for isolated magma bodies. Other areas of this kind are south Mt. Hromundartindur and a large area south of Mt. Hengill at the locality of the Orustuholshraun lava field (Fig. 2). Neither of these two latter areas are well constrained by seismic data, however. For the most part correlation between the depths to the bottom and the top of the seismogenic layer suggest that a roughly constant 3km-thick seismogenic layer underlies the area (Figs. 4 and 6).

Geothermal gradients measured in boreholes

1. *Nesjavellir*. Temperatures of 250–300°C are detected at about 2 km depth beneath Nesjavellir. The depth to the base of the seismogenic layer varies from



Fig. 5. Cross sections showing the hypocentral distribution of events beneath the area. The lines of section are shown in Fig. 3. (a) NW–SE section passing through the north part of Mt. Hengill and the Grensdalur volcano. (b) WSW–ENE section passing through the southern part of the Hengill system and the Klambragil area. (c) SSW–NNE section within the currently active accretionary plate boundary. Velocity contours (km s⁻¹) for this section from the three-dimensional crustal structure calculated using tomography are shown (Foulger and Toomey, 1989).

about 6 km beneath the south part of the well site to 4 km at a distance of 2 km further to the NE, beneath the northernmost part. Assuming a temperature of the base of the seismogenic layer of $650 \pm 50^{\circ}$ C, then this would predict average geothermal gradients in the depth range 2–6 km beneath south Nesjavellir of $87 \pm 12^{\circ}$ C km⁻¹ and in the depth range 2–4 km beneath north Nesjavellir of $200 \pm 25^{\circ}$ C km⁻¹.

The much higher average gradient predicted for north Nesjavellir than for south Nesjavellir is inconsistent with the decrease in shallow geothermal gradient observed from south to north and, therefore, it is likely



Fig. 6. Map showing the depth to the top of the seismogenic layer within the area studied using tomography (Toomey and Foulger, 1989; Foulger and Toomey, 1989), calculated as the depth to the 10th percentile event. Schematic tectonic features are shown as for Fig. 3, and the low-velocity body north of Mt. Hengill is outlined.

that the absence of earthquakes deeper than 4 km beneath north Nesjavellir results from a cause other than high temperature, e.g., low strain or an incomplete earthquake data set.

A temperature of 1150° C at 7.5 km depth is predicted by interpretation of a magnetotelluric sounding at Nesjavellir. This suggests that the geothermal gradient below the depth of the base of the seismogenic layer, in the depth interval 6–7.5 km, is $333 \pm 33^{\circ}$ C km⁻¹. This high value may reflect a fairly localised, very high gradient in the neighborhood of the partial melt layer, rather than a uniform gradient, in agreement with the hypothesis of Björnsson et al. (1980) that large temperature discontinuities occur in the neighborhood of partial melt. The thermal data for Nesjavellir are summarised in Fig. 7 and Table 2. 2. Sleggjubeinsdalur. The average temperature gradient measured in this well is variable with depth but the average gradient of 205°C km⁻¹ for the upper 1.2 km is slightly lower than gradients revealed beneath Nesjavellir. No earthquake data are available in the vicinity (Fig. 4), so no reliable estimate may be made of the depth of the 650 ± 50 °C isotherm.

3. *Reykjakot.* Temperatures reach 180–225°C at 0.75 km depth, increasing to the North. The base of the seismogenic layer lies at about 5 km depth, suggesting an average geothermal gradient of $105 \pm 17^{\circ}$ C km⁻¹ for the depth range 0.75–5 km beneath the well site. This thermal gradient is somewhat higher than that obtained for south Nesjavellir, within the currently active accretionary zone.



Fig. 7. Summary of information on the geothermal gradient beneath Nesjavellir down to 7.5 km depth. The increase in geothermal gradient below 6 km depth suggests that a sharp decrease in conductivity must occur there. The observations could be explained by a relatively thin layer of unfractured, conducting rock separating the partial melt thought to exist at 7.5 km depth from the solid, relatively permeable rock above, according to the hypothesis of Björnsson et al. (1980). In that case the geothermal gradient would not be uniform in the depth range 6-7.5 km.

4. Ölfus. The maximum depth of seismicity here indicates temperatures of about $650 \pm 50^{\circ}$ C at 7 km depth and shallow geothermal gradients are about 150°C km⁻¹ in the upper 1 km. These data imply a gradient of $83 \pm 8^{\circ}$ C km⁻¹ for the depth interval 1–7 km.

Temperature maxima within the geothermal reservoir

The three isolated temperature maxima within the high-temperature geothermal area identified by fumarole gas geochemistry are centered over the Grensdalur central volcano, the Klambragil area within the Hromundartindur system and beneath south Mt. Hengill (Fig. 4). Isolated heat sources and enhanced geothermal gradients at these localities are probable.

The Grensdalur temperature maximum is very close to the Reykjakot well site and not well constrained by earthquake data (Figs. 2 and 3). Extrapolations from nearby seismic areas indicate a depth to the seismogenic base of about 5 km, approximately the same as beneath the Reykjakot well site and thus a similar thermal structure is plausible (Fig. 4).

The Klambragil temperature maximum lies immediately adjacent to the most seismically active part of the area and a large data set constrains the maximum seismic depth to be about 4 km (Figs. 4, 5a and 5b). This is the shallowest, well-constrained depth to the seismogenic base detected in the area. If the shallow geothermal gradient is about the same as that measured at Reykjakot in the topmost 0.75 km, then an average gradient of about $137 \pm 22^{\circ}$ C km⁻¹ is implied for the depth range 0.75–4 km beneath Klambragil. This gradient is considerably larger than predicted for the Nesjavellir and Reykjakot sites and in qualitative agreement with the gas fumarole estimated temperatures of the reservoir fluid source of 20–30°C higher for the Klambragil area than the two drill sites. The seismicity occurs within a high-velocity body in the depth range 2.5–5 km, interpreted as gabbroic intrusions (Foulger and Toomey, 1989).

The temperature maximum beneath the southern part of Mt. Hengill is associated with an exceptionally shallow seismogenic base of less than 2 km. This is constrained by very few earthquakes, however, and would yield unrealistically high estimates of the geothermal gradient. The seismic data set is thus probably incomplete in this area.

8.2. Implications for the structure and tectonics of the Hengill area

As discussed above, large uncertainty is associated with the temperature of the base of the seismogenic layer. Also, the assumption that this layer is an isotherm is an approximation, as factors such as strain rate and

Table 2Geothermal gradients within the Hengill area

Site	Depth range (km)	Gradient (°C km ⁻¹)	Source
Nesjavellir	0-0.8	312	boreholes
	0.8-2	70	boreholes
	2-6	87 + 12	seis. base
	6-7.5	333 ± 33	seis. base + MT
Sleggjubeinsdalur	0-1.2	205	boreholes
Reykjakot	0-0.75	240-300	boreholes
	0.75-5	105 ± 17	seis. base
Grensdalur	0-0.75	240-300	extrapolation
	0.75-5	105 ± 17	seis. base
Ölfus	0-1	150	seis. base
	1–7	83 ± 8	seis. base
Klambragil	0.75-4	137 ± 22	seis. base
Whole area	0-seis. base	93 ± 7 to	seis. base
		162 ± 12	

water content may vary over the area. However, the earthquakes are accurately located, and where substantial numbers of events occur, the depth to the bottom of the seismogenic layer is well constrained. The variations in geothermal gradient determined within the area are thus better constrained than the absolute gradients.

The depth to the base of the seismogenic layer varies from about 4 to 7 km and, therefore, has a relief of about 3 km. We thus anticipate that the average geothermal gradient from the surface to the base of the seismogenic zone varies from $93 \pm 7^{\circ}$ C km⁻¹ to $162 \pm 12^{\circ}$ C km⁻¹ throughout the area.

The seismicity inside the currently active spreading zone, is, surprisingly, not shallower than beneath neighboring areas, but includes the deepest earthquakes in the area outside the transform zone. With the exception of the neighborhood of an isolated volume containing partial melt, high geothermal gradients correlate with the intensity of the geothermal area and not with the currently active plate boundary. This suggests that the spreading zone at this part of the plate boundary is not associated with a narrow, localised zone of thermal upwelling on the scale of the fissure swarm.

8.3. Comparison with the mid-atlantic ridge (MAR)

The depth of microseismicity recorded on ocean bottom seismometers (OBSs) deployed along the MAR is 6–10 km (Toomey et al., 1988; Kong et al., 1992). The base of the seismogenic layer is shallow where the volcanic extrusion rate on the sea-floor is high, suggesting that high temperatures occur at shallow depth at these locations, an intuitive result. No correlation between the topography of the base of the seismogenic layer and the surface topography (i.e. extrusion rate) was noted in the Hengill area. Instead shallowing of seismicity correlated with the intensity of the geothermal area.

The depth determinations of the events from the MAR are much less accurate than those available for the Hengill area, rendering comparison difficult. The oceanic events were recorded on very small numbers of stations only and the crustal structure beneath the rift is known much more poorly than beneath the Hengill area. It is therefore possible that the variation in maximum depth of seismicity reported for the MAR is partly the result of random and unmodelled systematic errors.

Before the thickness of the seismogenic layer beneath the MAR may be compared meaningfully with that beneath Iceland, much better constrained microearthquake data from the sea-floor are required. Dense arrays of OBSs are necessary to acquire such data. Comparison of the maximum depth of microseismicity beneath black smoker areas with that of other parts of the marine rift might reveal significant differences.

9. Conclusions

(1) The continuous, seismic thermal contraction cracking at the Hengill geothermal area occurs at temperatures of less than $650\pm50^{\circ}$ C. Seismic volumes indicate cooling, solidified intrusions at temperatures lower than this.

(2) A shallow, isolated, elongate, aseismic magmatic body with a volume of a few km³, which may have been formed or reactivated 200 years ago, possibly underlies the northeast side of Mt. Hengill.

(3) Depressions in the top of the seismogenic layer indicate areas where shallow, isolated magma bodies may exist.

(4) The thickness of the seismogenic layer is roughly constant and about 3 km throughout the Hengill high-temperature area.

(5) The geothermal gradients below drilling depths and above the depth to the bottom of the seismogenic later vary from about $83 \pm 8^{\circ}$ C km⁻¹ to $137 \pm 22^{\circ}$ C km⁻¹, i.e., by about 54^{\circ}C km⁻¹. The variation across the area is more confidently determined than the absolute values.

(6) Away from isolated, partial melt bodies, the geothermal gradient beneath the currently active spreading axis is not anomalously high compared with immediately neighboring areas. Geothermal gradients correlate with the intensity of the geothermal area rather than with the plate boundary.

(7) Comparison of results from Iceland with those from the MAR are difficult because of the lack of comparable data from oceanic areas. However, the Icelandic data suggest that comparison of the seismicity of black smoker areas with distal parts of the marine rift could yield significant differences.

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