

Studies of Temperature, Viscosity, Density and Some
Types of Materials Produced in the Surtsey Eruption

by

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I. Temperature.

Temperature was studied with the same optical pyrometer, which I used in the Hekla eruption of 1947-48 (1, p. 22). Two runs of laboratory comparisons with a thermocouple were done during the earlier studies and a third was added now in collaboration with Professor Thorbjörn Sígurgeirsson. The results are shown in Fig. 1. The accuracy of a single pyrometer measurement appears to be 20-25°C. In the third run comparison was also made with Sígurgeirsson's thermocouple, which he used in Surtsey on 15th Oct. 1964.

On the second day of the eruption, Nov. 15th, 1963, I observed at close quarters from the Coast Guard vessel Albert. In daylight no glowing material could be perceived in the eruption columns. In darkness sporadic glowing stones were seen, the temperature not surpassing 650-700°C. At this time the eruption was submarine, and the material thrown out had no doubt been cooled by the water. The low maximum temperature is, nevertheless, remarkable. When one considers that a bomb of 20 cm diameter, surrounded by cold water, should retain its central temperature unaffected for about 10 minutes, one might expect to find occasionally the original temperature of the magma when such and larger bombs split in the air. It seems possible, therefore, that the magma temperature was very low.

The character of the eruption remained the same for several months, i.e. the material was non-luminous, only sporadic bombs of a deep red glow were observed. In the beginning

of April 1964 lava began to flow. On April 9th I observed the lava fountains in the crater from an airplane, in broad daylight. The conditions were not favourable for a temperature measurement, but my estimate was close to 900°C . On August 19th I studied the lava fountains at close quarters, obtaining the temperature 1000°C . On October 15th I got $1070\text{--}1100^{\circ}\text{C}$. On the same day I found 1070°C in a lava tongue at the lava front 900 m from the crater along the flow path. By thrusting a thermocouple deep into this tongue, Sigurgeirsson obtained $1110\text{--}1130^{\circ}\text{C}$. The difference may be due to some surface cooling. In later studies Sigurgeirsson found temperatures around 1140°C with a thermocouple (this report).

It would appear from these data (Fig. 2) that the lava temperature increased very markedly from beginning to the end of the eruption. Unfortunately, comparisons with thermocouple studies are not possible for the period where the main rise took place. But the pyrometer observations are supported indirectly by a study of the erupted material, cf. III.

II. Viscosity.

For the estimate of viscosity in the flowing lava three methods were used: 1) Lava-penetrometer, i.e. an iron stick thrust into the lava (cf. 1, p. 15); 2) Velocity of flow; 3) Waves in the crater lake. The first method was difficult to use because of the great fluidity of the lava. After several attempts I finally (Oct. 15th 1964) obtained as a fair measure of the viscosity in a lava tongue at the front that the stick sinks 10 cm in 0.5 seconds under its own weight. With the assistance of Mr. Bragi Arnason of the University Physical Laboratory this was found to correspond to a viscosity of $5 \cdot 10^3$ poises.

For the same lava tongue the formula for the velocity $v = g h^2 \sin \alpha / 2\eta$, gives $\eta = 2 \times 10^4$ poises, with the values $v = 40$ cm/17 sec; $\alpha = 5^\circ$; and estimated depth $h = 25$ cm. The breadth was 35 cm. The viscosity found in this way is probably a little too great, as forming of crust will impede the flow of such a small tongue.

During my visit of August 19th 1964 I observed waves in the lava pond. The pond was 10-12 m in diameter and surrounded by perpendicular walls. At a certain spot arose 10-15 m high fountains in a rapid succession, this being the main exit of gas, or the top of the funnel. Each time a big fountain arose, a broad wave started from the spot and moved towards the walls. Here the lava surface then rose for a short time by 2-3 m, after which the surface became plain again. There was no reflection of the wave, the energy being lost through the great viscosity. Beside the main exit of gas, big bubbles exploded here and therein the entire pond like vapour bubbles in a porridge; they produced 1/2 - 1 m high gushes, but no waves were formed. These observations suggested a method for obtaining the viscosity in the lava pond.

H. Lamb (2, pp. 625-28) has considered the influence of viscosity on surface waves. With $\nu = \eta/\rho =$ kinematic viscosity, $c =$ velocity of the waves, $\lambda =$ wave length, he defines a number $\theta = \frac{2\pi\nu}{c\lambda}$. Then the type of motion can be judged by the roots of the equation $(x^2 + 1)^2 = 16\theta^3(x - \theta)$. Two complex roots cancel out as they violate certain conditions. The remaining roots are admissible and may be real or complex according to the magnitude of θ . For a low value of the viscosity and not too small wavelengths θ is small. The roots are complex and give ordinary waves of slowly decreasing amplitude with time and distance of travel. In the case of a very viscous fluid, such as treacle or pitch, θ may be large even when the wave-length is considerable. The admissible roots are then both real. One root represents a slow creeping of the fluid towards a state

of equilibrium and the other root represents a wave dying out rapidly. Obviously the margin between periodic and aperiodic motion is found where the two real roots merge together. This I found to occur for $\theta = 1.31$. Hence wave motion is found for $\frac{2\pi r}{c\lambda} = 1.31$; with $c = (g\lambda/2\pi)^{1/2}$ we get $r \leq 2.61\lambda^{3/2}$ (cf. Fig. 3) or $\eta = 2.61g\lambda^{3/2}$. The meaning of the formula is: in a fluid of given viscosity, long progressive waves are possible, but below a certain wave-length, given by the above condition, no such waves are possible. Observing the shortest occurring waves then gives the viscosity.

During my first observation it was clear that the occurring waves were of the order of a very few metres. A second time, October 15th, 1964, I was better prepared for what to observe and conditions were very favourable, cf. Fig. 4. The lava rushed up in fountains at A, then flowed with great speed and in great quantity down a low fall F into a broad stream S and then disappeared into the tunnel G. Waves, 1 m high and 1-2 m broad, continually rushed down the main stream S. At the side of the main stream was a quiet pond V, partly sheltered from the fall by the promontory B. But every now and then waves were formed at the promontory and moved into the pond. I observed the waves from C, in shelter of the narrow wall around the crater. The waves were close to 1 m broad and died out after a travel of 1-2 m from B. Smaller waves were not seen; large gas bubbles exploded here and there in the pond, and especially in the main stream, but they never caused wave motion. Taking the breadth of the observed wave to correspond to a half wave-length, we get with $Q = 1.5$ (cf. III), $\eta = 1 \times 10^4$ poises. As to the state of this lava cf. III. This is probably the best value for the viscosity. The low value obtained with the penetrometer is probably caused by the advanced porosity of the lava in the tongue - the lava was foamlike. But all methods agree as to the order of magnitude.

III. Study of some types of material produced in the eruption.

Description of samples from different times of the eruption:

- 1) Fallen on deck of the Coast Guard vessel Albert at 9 o'clock, 15th Nov. 1963 (second day of the eruption) when the ship sailed under the ash cloud. The material consists of small glass fragments less than 5 mm in diameter; in thin section this is translucent brown sideromelane with very small gas bubbles indicating the very beginning of gas release; practically unexpanded glass. All grains sink in water. Bulk density of watersoaked material 1.4. The glass is all of a single type in contrast to 2). There is a fair amount of loose crystals which is not the case in 2).
- 2) Fallen on deck of the Coast Guard vessel Albert Nov. 16th. The material consists mainly of expanded irregular small lumps, 1-2 cm in diameter. About 1/13th by weight floats in water. In each lump one recognizes generally two or three types of glass: a) a dark-grey skin, 1 mm thick; it is much cracked and distorted as a result of expansion of the interior of the lump. But the glass in the skin is devoid of bubbles. Clearly the skin was formed by quenching of each lump before expansion began." b) under the skin is brown glass with fine pores. Sometimes it occupies all the interior, but mostly the central part consists of c) much expanded glass of bluish lustre, due to oxydation of the iron; there is no sharp limit between b) and c).

These small lumps of pumice are in principle similar to the much larger bread-crust bombs, which one could observe in the Hekla eruption of 1947-48 (1, p. 54). Summarizing the genesis of the Surtsey pumice of sample 2) we conclude: Before gas was released, i.e. under considerable pressure, the magma was split into small fragment, mostly below 1 cm across. The fragments were then quenched on the surface and

it seems the most likely theory, considering later observations of the flow of lava into the sea, that splitting and quenching took place at the sea-bottom (depth 130 m, pressure 14 atm.). As the smooth surface of the original skin suggests a fracture surface, we should visualize the process of splitting as shattering due to rapid uneven cooling. Immediately after formation, the grains come under low pressure, i.e. they were formed in an explosion and immediately thrown into the air. Cooling of such small grains must be a matter of a very few minutes at most, but still there was time for expansion of the interior and for oxydation of the iron. The latter fact is of special interest.

In my study of the Hekla bombs quoted above, it was pointed out that the internal oxydation was related to the decrease of pressure, but because of the slow cooling of the very much larger pieces than we are dealing with here the process was not fully clear. Here we have a magma which must have approached the surface relatively slowly and it had been released of most of the original pressure (many thousand atmospheres) when it was shattered. Yet, it was in a very short time after shattering and after the final, but relatively small relief of pressure, that oxydation took place together with release of gas. It is therefore clear that it is the release of gas which is the direct cause of the oxydation. It is probably the loss of hydrogen and the consequent availability of oxygen in the melt that oxydized the iron.

Besides the primary fragments of pumice, which we have discussed, there is in the sample some amount of such unexpanded grains as we found in 1). These are always much worn. Most of the later material was worn, and we may remark at once that the origin of the wearing was obvious: the greater part

of the material which was thrown up in an explosion dropped back into the crater. It was therefore thrown out many times and much worn in the turbulent sea in the crater before it finally settled as lapilli outside the crater.

- 3) Sample taken by Dr. Thorleifur Einarsson close to Surtsey on Nov. 21st, 1963. Very fine-porous translucent glass containing a considerable amount (10%) of crystals. The biggest grains are 3-4 mm, but the most common size is only 0.05 mm. Worn material.
- 4) Sample taken from a crater wall on Dec. 16th, 1963, by Professor Thorbjörn Sigurgeirsson. Bulk density of dry material 1.20; water-soaked 1.62 (container filled with dry material gives the first figure; addition of water until all is soaked and the water stands at the edge of the container, gives the second figure). Average density of individual grains is then 2.06. Washing cleans away 18% of the mass as fine silt. In the remainder are grains mostly in the interval 1-8 mm; grains of 3-8 mm generally worn.

On August 19th 1964 I took the three following samples from a crater wall on the east side of Surtsey. The sea had eroded the southern half of the crater so that a good section was available.

- 5) Base of wall. Density of dry material 1.36; water-soaked 1.72. Average density of grains 2.11. Washing cleans away 48% of finest material. In remainder, diameters of 1-3 mm most common, but a few grains of 5-10 mm occur. Grains all equidimensional and considerably worn.
- 6) Layer at 6-8 m depth on inner side of crater; one of the coarser layers in the section. Grains of 5-20 mm common. Density, dry 1.30; water-soaked 1.65. Average density of

grains 2.02. Washing cleans away 31% of the material; the remainder mostly in interval 1-10 mm; equidimensional worn grains. Higher layers mostly finer than 6); alternation of lapilli and layers with small bombs, irregular pieces of pumice and fragments of sandstone from the sea bottom. Sample 7) is taken from a lapilli layer.

- 7) Density, dry 1.24; water-soaked 1.68; average density of grains 2.20. Washing leaves 86% of material, in the interval 1-10 mm; equidimensional and worn. Crystallation of same degree as in 3). The samples 4), 5), 6) and 7) give very similar values for the density. The average of the dry material is 1.28 and the average for grain density is 2.10. This grain density corresponds to 23.5% bubble volume in the average grain, if the density of the dense glass is taken as 2.75.

In the Surtsey mound of glassy material as a whole, the material is considerably compressed and the bulk density must be higher than 1.28. But it must be lower than 2.10 and the average of these figures, 1.69, may be taken as a reasonably correct value. To convert the volume of the mound into volume of compact glass of density 2.75 we must then multiply by the factor $1.69/2.75$.

As a crude measure of the mound I take here a cone with 30° side inclination; height 130 m below and 30 m above sea-level; area of section at sea-level 2 km^2 . The volume is then 0.395 km^3 , corresponding to dense glass of 0.243 km^3 . This is about 60% of Hekla's production in the eruption of 1947-48. It must be emphasized that the assumed geometry of the mound is an idealization.

- 8) On Oct. 15th 1964 I took some samples of spatters which had recently been thrown over the rim of the crater in fountain activity. The density of two pieces with a weight of

413 g is 1.42 and of two other pieces of weight 372 g is 1.43. In the 2-4 cm thick spatters the porosity is considerably greater in the middle than towards the surface and the expanded interior has the characteristic blue lustre due to higher oxydation of iron. In a thin slide across a 2-3 cm thick spatter the pores are 0.15-1.05 mm in the crust, i.e. in the original spatter, but increase to 3-4 mm in the centre. A probable figure for the density of the original spatter is 1.5 which corresponds to a pore volume of 45%, or about twice the value found for the material of the early explosive phase of the eruption. This suggests an easier expansion of the gas bubbles, due to higher temperature and lower viscosity than earlier in the eruption.

The spatter is, furthermore, crowded with crystals; a measurement by S. Steinthorsson gave 60% glass, 27% feldspar, and 13% olivine. The olivine crystals are usually much thicker than the glass walls between the bubbles and it is also clear for the feldspar that it must have existed in the melt in the crater. There are no such crystallites or needles that might suggest crystallization in the spatter itself. Thus we may safely conclude that the magma is already 40% crystallized when it appears in the crater. This is 3-4 times the amount of crystals found in the glass of the explosive phase, which may be an indication of difference in viscosity. From the point of view of viscosity we see that the fluid we are dealing with is very special: gas bubbles make out one half of the volume, while the other half is nearly evenly divided between a fluid melt and loose floating crystals.

The high degree of crystallization has also a clear relationship to the temperature. The heat of crystallization is about 90 cal/g and the specific heat is about 0.25. Complete crystallization, therefore, releases heat which

would suffice for raising the temperature by 360°C . By the 40% crystallization of the rising magma the temperature might rise 143° if no heat was lost. S. Steinhörsson (this report) found that the olivine was about the same in the spatter as in the early material. The difference lies in the 27% of feldspar found in the spatter. This difference in crystallization corresponds to a difference in temperature of about 100°C . It is most likely that there was no time for loss of the heat released by crystallization and that of the 1140°C measured in the lava, about 100° were acquired through crystallization of feldspar during the rise of the magma. The temperature "proper" of the magma would then have been near 1040°C . We then arrive at this picture: In the beginning of the eruption the magma reaching the surface of the earth's crust was probably rather cool due to loss of heat to the cold walls of the fracture. The temperature may have been considerably below 900°C , possibly even as low as 700° ; the temperature actually found in exploding bombs. At this time the melt was too viscous for any crystallization to take place in the rising magma. Slowly the erupted magma became hotter as the loss of heat to the walls decreased but during the early, explosive phase the change was small.

The eruption changed over to effusion as a result of blocking of sea-water and the temperature appears to have been still as low as 900°C at that junction. But now gradually a certain chain reaction begins to work: increase of temperature lowers the viscosity, which increases crystallization; this in turn raises the temperature and so on.

There are interesting aspects of general interest in this process in a rising undercooled magma. Suppose the original magma temperature is 1200° . Small amount of crystallization raises the temperature to 1250° and this temperature will

block further crystallization, so the process cannot go further. In another example the original temperature be 900° . Then it appears that the process can run to such a length that the magma crystallizes completely during rise, and the process stops the eruption. In a melt of intermediate temperature the process seems to be of importance.

IV. Flow of lava into the sea.

Several cases exist in Iceland where it is either known or can be safely inferred that a postglacial lava has flowed into water and no very marked effect of the water is discernable. One finds, it is true, one peculiarity: distortions and great cracks that suggest some havoc played by steam. In my first study of the lava in Surtsey I found just such peculiarities where sea erosion had opened access to the interior of a lava flow. On March 20th, 1965, I observed lava flowing into the sea and saw how it was completely split up into small fragments on contact with the water. A 1 m broad tongue flowed pretty rapidly down to the beach during retreat of the sea. The next wave overran the frontal part of the tongue and changed it completely into small fragments that ran down the beach with the backwash. In this way the lava was constantly cut off at the beach, being changed into glass fragments which the waves carried along the shore. Part of this material formed an 8-12 $^{\circ}$ steep "sand" beach along cliffs of the island, and much of it was transported farther to gather as a broad flat on the western part of the island.

In front of the "sand" beach, onto which the lava was flowing, there was a 40-45 $^{\circ}$ submarine slope, probably mainly "sand". It is then clear that at least in some cases the theory

presented first by Fuller (3,4)* is verified: the lava entering water first builds a "delta" of foreset layers of fragmental material on which it then gradually proceeds as if on dry land. Foreset beds of breccia covered directly by a cap of lavas are quite common in the volcanic breccias of Iceland (5,6). Although the Surtsey material is not very typical for these breccias, some similarity of origin may be assumed.

V. Some general remarks on the release of magmatic gas and resulting changes in its composition.

The multi-component gas given off by a magma, and sampled for analysis, must generally be expected to have a composition rather different from that of the gas originally contained in the magma, due to the complications of the process of gas release. To analyze this process one must consider, on one hand, how the gas in general is given off, on the other hand, study the differential release due to different solubilities of the gas components. Of some of the relevant factors in this process we have a fair idea, others are quite unknown.

* "A fluid lava on encountering a local body of water would tend to granulate like molten slag and would thus form a fine breccia, which would accumulate to a depth approximately equal to that of the water. The fine breccia would settle until its surface attained an angle of repose which, owing to the roughness of the fragments, would be relatively steep. If the molten cascade continued to pour out into the water, the accumulation of granulated glass would gradually advance like the foreset bedding of a delta. The inclined bedding would be preserved by the thin sheets and the ropy or ellipsoidal masses, which failed to granulate. Except for the possible effects of rising steam, the flow would gradually advance on top of these foreset beds as if on dry land".

We consider first the release of gas in general. The observation mentioned in III, 3, concerning the unexpanded crust of pumice, indicates that when the magma met the seawater, probably at a depth of 130 m, i.e. under a pressure of 14 atmospheres, gas release by bubble-formation had not begun. This pressure corresponds to the weight of a 50-65 m high column of magma that is more or less porous. We may then expect that later, when the lava had a free subaerial surface in the crater, gas release by bubble formation took place only in the uppermost 50 m or so of the lava column. The extreme fineness of the bubbles (less than 1 mm) in the spatter studied (III, 8) might even suggest that the bubbles began to form at a much shallower depth; otherwise the bubbles would have expanded due to decrease of pressure.

Such small bubbles are incapable of rising individually in a melt of viscosity 10^4 poises: they move passively with the rising magma (a bubble of 1 mm diameter needs about 2 hours to rise 1 cm through the melt; Stokes law). It is only or mainly where differential or shearing movement takes place in the magma that bubbles may coalesce and form larger bubbles that are capable of marked relative rise; such bubbles may be seen to burst at the surface of the lava (cf. discussion of gas release in Einarsson 1949, pp. 9-15). Fountain activity must be considered as due to the bursting of such bubbles. These large bubbles may be said to be of a second generation, while the study of volcanic materials seems clearly to show that the first and primary generation consists of a dense network of very small bubbles, changing the magma into a sponge. This after all is the normal process of gas release from an oversaturated fluid. If the gas release could be impeded until the magma was very much supersaturated, i.e. until the pressure had fallen, say to atmospheric pressure, the process of release would be rather explosive and the release would take place into the free atmosphere and not into the closed spaces of the small bubbles.

Analogy to geyser activity, where the water boils explosively after superheating, would then be very close. But such a process seems not to take place, considering the generally fine spongy state of the material thrown out in fountains; explosive release of gas throughout a mass would produce a different type of material.

The main release of gases from the rising magma then takes place into closed spaces of bubbles, first at the highest pressure, later at a gradually lower pressure, and it is this gas that becomes available for analysis.

If the solubilities of the gas components were all equal then the free phase in the bubbles and the dissolved phase would have the same composition. This is at any rate true if the gas volume is a small part of the melt. But for a large gas volume it is not true; very sparse components will then have a less percentage in the free phase than they had originally in the dissolved phase.

If on the other hand the solubilities are different for the various components, and this must be assumed to be the case, then the free phase will be significantly different from the original dissolved phase, whether the gas volume is small or large. We do not know the solubilities and an exact calculation of this difference cannot, therefore, be made. But for sake of demonstration of the effect we shall choose some arbitrary figures.

We shall first consider a stage where the volume of the bubbles is 5% of the magma and the pressure is assumed to be 10 atmospheres which seems realistic. The stage is assumed to last long enough for equilibrium between the two phases of gas to be reached. In the free phase the gas is assumed to have the following composition in mole percentages: H₂O 80%; SO₂ 5%; CO₂ 8.8%; H₂ 6%; A 0.2%. Argon is arbitrarily included because of its interest in connection with the K/A dating method.

We next assume that the water content of the magma is 0.4% by weight, in equilibrium with the partial pressure 8 atm. of the vapour in the bubbles. 1 cm³ of the melt then contains 54 cm³ of vapour, reduced to 1 atm. and a temperature of 1150°C, and the corresponding solubility for 1 atm. partial pressure is 6.75 cm³.

For the other gases we have no data and we must select arbitrary figures. The solubility in water at 20°C and 1 atm. is SO₂ 39.4 cm³/cm³; CO₂ 0.878 cm³; H₂ 0.0184; A 0.037 cm³. We shall tentatively assume that for the magma, and reduced to 1 atm. and 1150°C, the solubilities are in all cases 1/5 of the above values. As the partial pressures in the gas phase are given, we can find the gas dissolved in the melt: H₂O 54 cm³/cm³; SO₂ 3.94 cm³; CO₂ 0.155; H₂ 0.0022; A 0.00015.

To these amounts we now have to add the free phase. The quantity per cm³ is multiplied by 10 to reduce to 1 atm. pressure, and by the factor 5/95 as the gases are 5% of the magma. We then get the additions: H₂O 0.42 cm³ giving a total of 54.42 cm³; SO₂ 0.0263, total 3.97; CO₂ 0.0463, total 0.201; H₂ 0.0316, total 0.0338; A 0.0011, total 0.00125. The original percentages are thus in the same order: 92.7; 6.76; 0.34; 0.058; 0.0021.

The difference between this and the released gas is very marked. In the process very little of the water and SO₂ has been freed, but 23% of the CO₂, 93.5% of H₂ and 88% of A.

We now assume that the magma part considered rises close to the surface. The bubble volume will increase by a factor of 10, due to change of pressure, and become 34.5% of the magma volume. At the same time further gas will be released and we assume a stage with 50% gas volume under a pressure of 1 atm. and equilibrium between the dissolved and free phases. The free gas is considered to have the same composition as in the first case. The gas per cm³ of melt is now in each case 1/10th of what it was in the first case, assuming validity of Henry's law. We then find the following percentages for the original gas in the

melt: 91; 6.52; 1.52; 0.88; and 0.03. The following amount of the original gas was retained in the melt: H₂O 87%; SO₂ 89%; CO₂ 15%; H₂ 0.37%; and A 0.75%.

The gases that finally leave the melt, and eventually are sampled for analysis, will be a mixture of gases released at various depths, and would then correspond to intermediate conditions between the two calculated cases.

As a general rule it is clear that small components of low solubility will very largely disappear from the melt.

If we consider a free lava surface then gases will diffuse out of it, no partial pressure will be built up as the gases are carried away in the atmosphere, and the gas content of the magma should decrease exponentially with time. But this process is certainly much slower than the expulsion by means of bubble formation and cannot contribute markedly to sampled gases. But insofar it does, the diffusivity of the various gas components must influence the percentages.

In summary, the gases begin to be released as densely set very small bubbles at a depth of 50 m or less in the rising magma. The surface of the bubbles being relatively large, equilibrium may be established between the free and dissolved phases. As the hydrostatic pressure decreases, the bubbles expand, partly without and partly due to further intake of gas. In small bubbles equilibrium between the gas phases may perhaps be established near the surface but for larger bubbles, that have been formed by coalescence at a greater depth, this will hardly be the case. Their gas will therefore be representative of equilibrium at a greater depth. In sampled gas the components of low solubility will be greatly enriched in relation to the other components.

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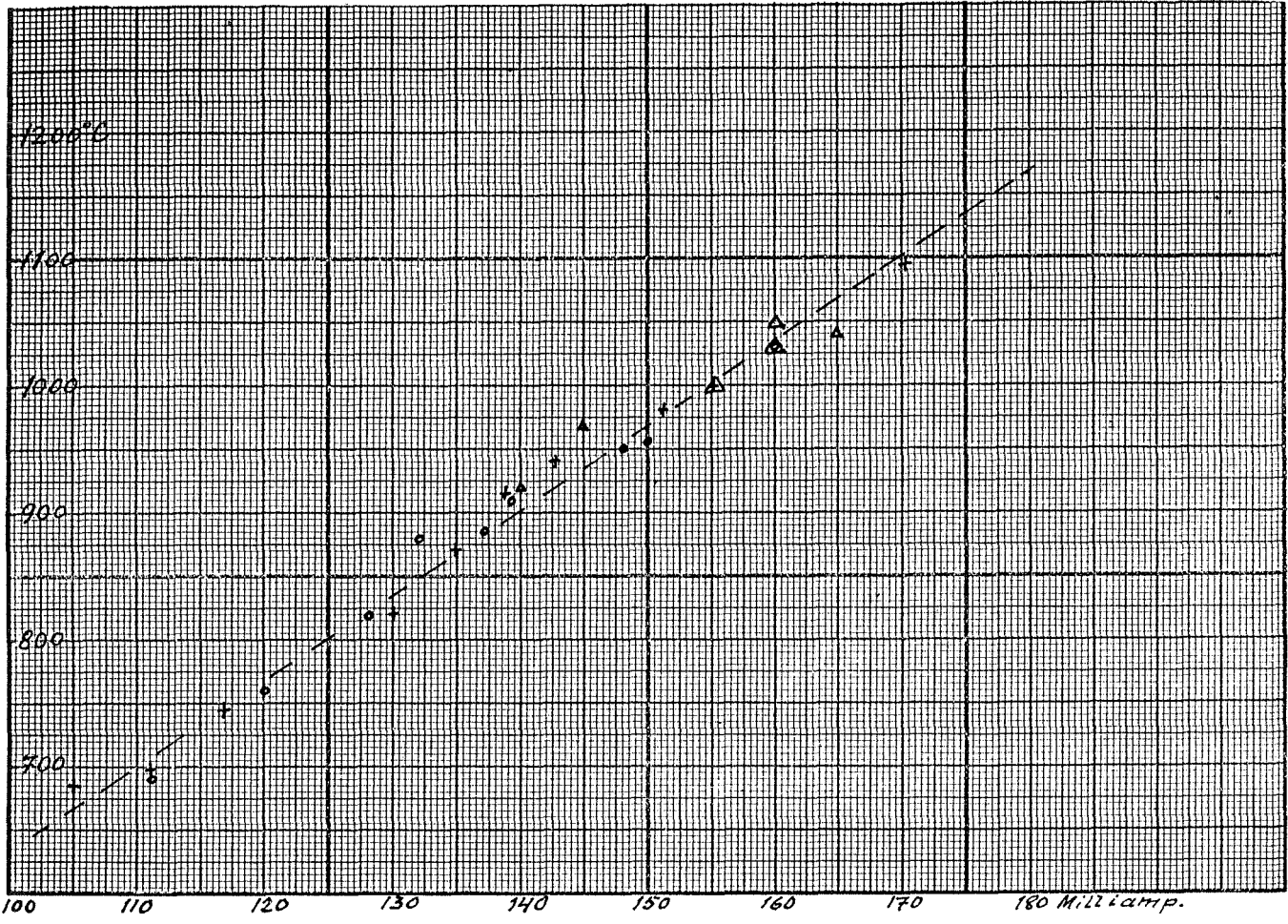


Fig. 1. Three separate comparisons (+, o, A) of pyrometer readings (Milliampere) with thermocouple temperature.

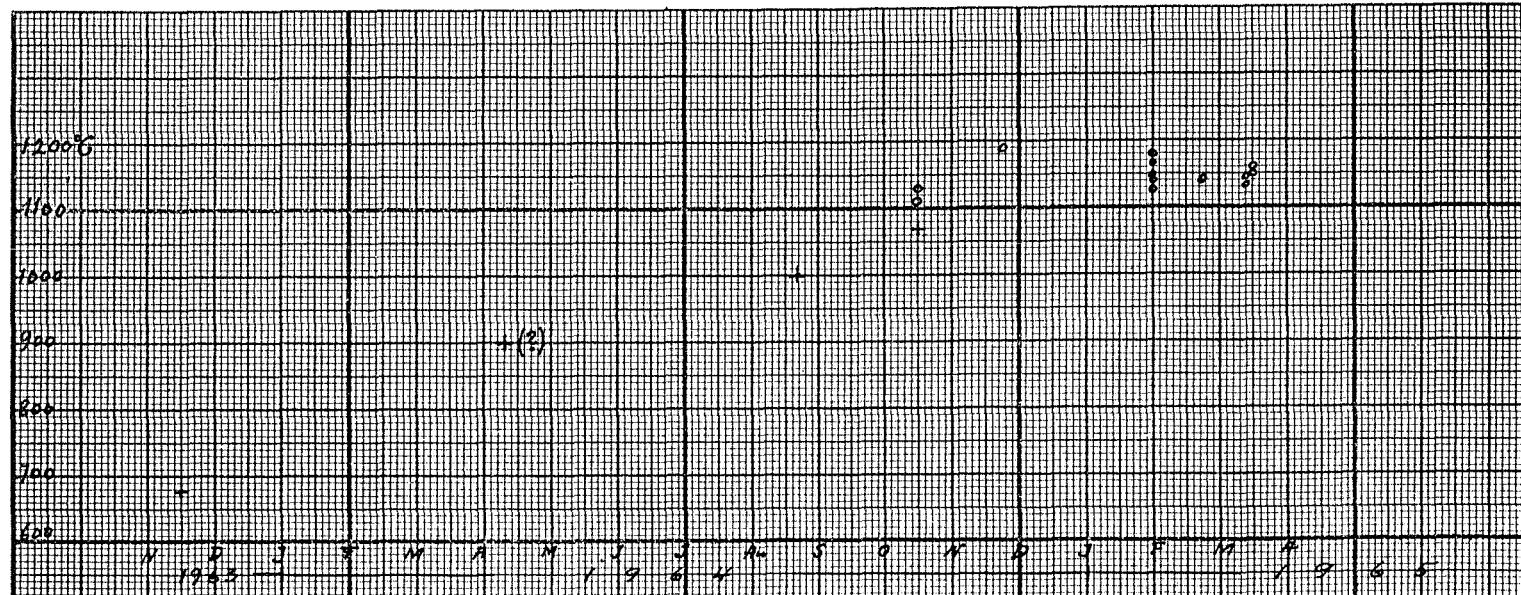


Fig. 2
 Temperature measurements in Surtsey.
 + Optical pyrometer (Dr. Einarsson)
 o Thermocouple (Th. Sigurgeirsson)

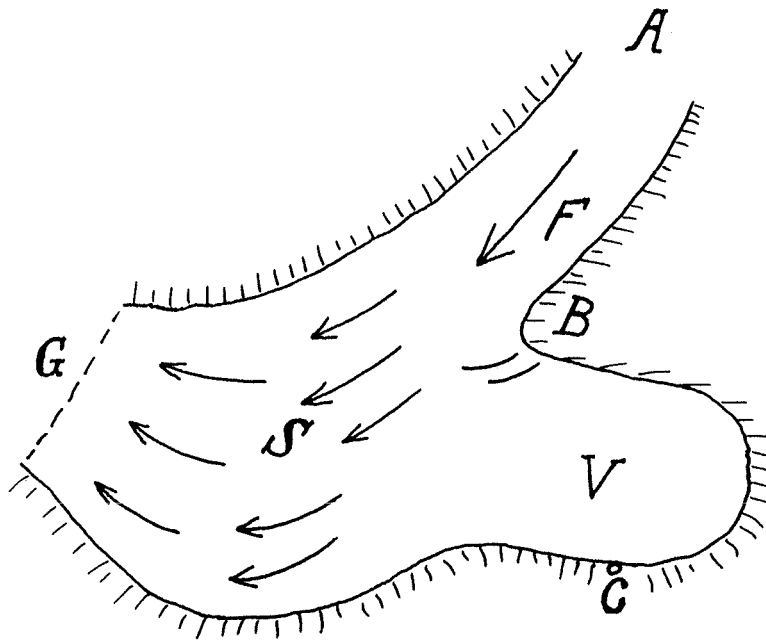
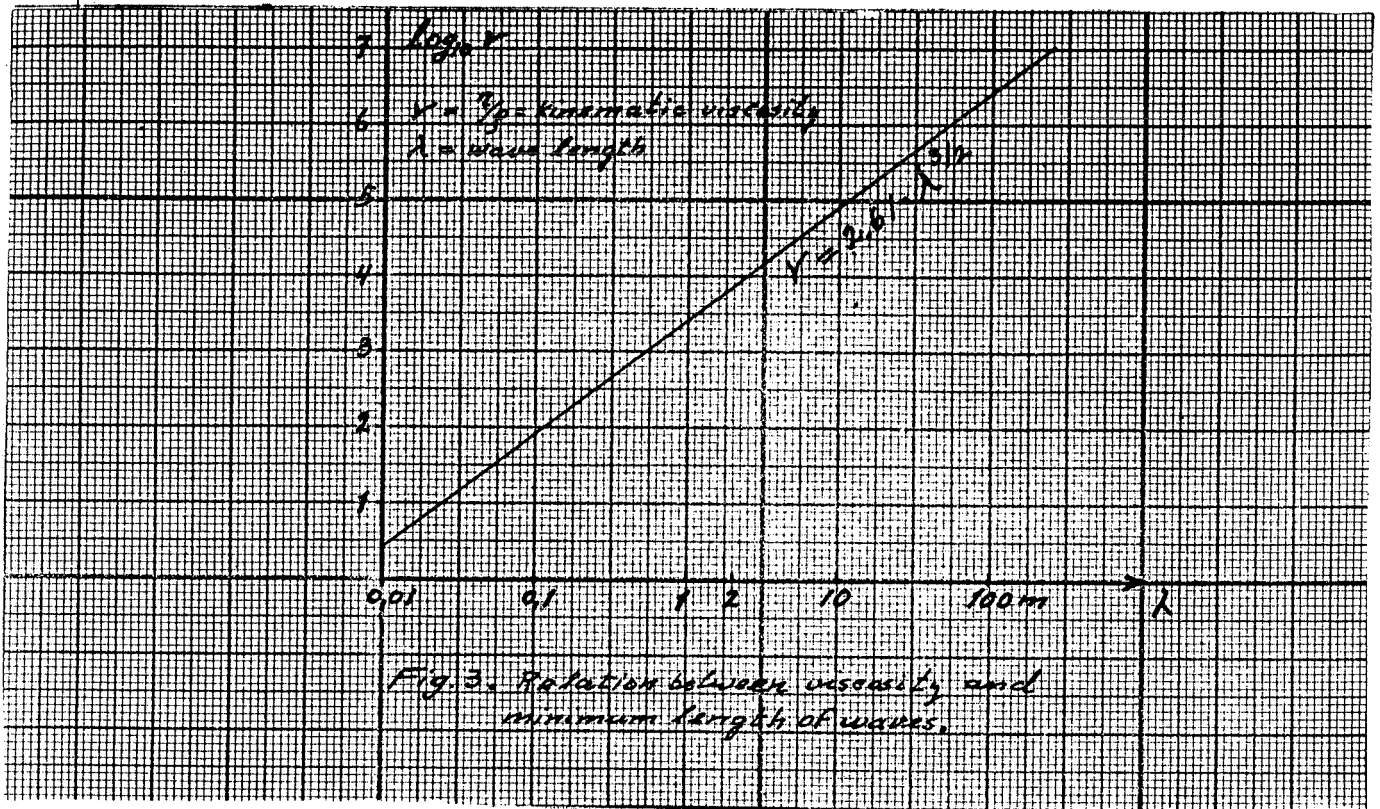


Fig. 4. Main stream of lava, F-S-G, and formation of waves at B.