GEOTHERMOMETRY OF THE KEBAN MINE AREA, EASTERN TURKEY

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ABSTRACT. — The Keban mine is one of the most important lead and zinc producers in Turkey. It is located 54 km NW of Elazığ County, Eastern Turkey.

The Keban metamorphic massif consists of calc-schist, dolomite marble, phyllite and marble. It forms part of the eastern Taurid belt, a prolongation of the Alpine orogenic belt.

The principal structural feature of the area is the northern extension of the Malatya-Keban anticline, a major recumbent fold with a NE-SW axis. Later movements acting in different directions gave rise to N-S, E-W and NE-SW directed folding and faulting over the previous anticline. Small bodies of quartz-syenite porphyry of Paleocene age intrude the metasediments.

Skarn zones have developed in association with the intrusion of quartz-syenite porphyry. These are located mainly within the metasediments. The magnetite deposit of Zereyandere, the scheelite deposit of Kebandere and the main sulphide deposit were formed as part of the process of skarn formation. Minor amounts of some manganese minerals, and the minerals vanadinite and descloizite derived from the main sulphide deposit, are also found in the district.

Using various methods, a temperature range of formation between 620° and 78° C is estimated for the ore minerals of the main sulphide deposit. For the ore minerals of the scheelite and magnetite deposits the range is from 743°C down to 225°C.

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INTRODUCTION

The temperatures of formation of minerals and ore deposits have long been paid attention, as this data can be helpful in the solution of abstract geological problems in economic geology. A series of temperature determinations in an ore deposit could establish the original tem-

Fig. - Index map of Turkey showing the Keban area.

perature gradient from which the direction of motion of the ore carrying fluids could be inferred and findings of this kind with the aid of other local factors could lead to new prospects.
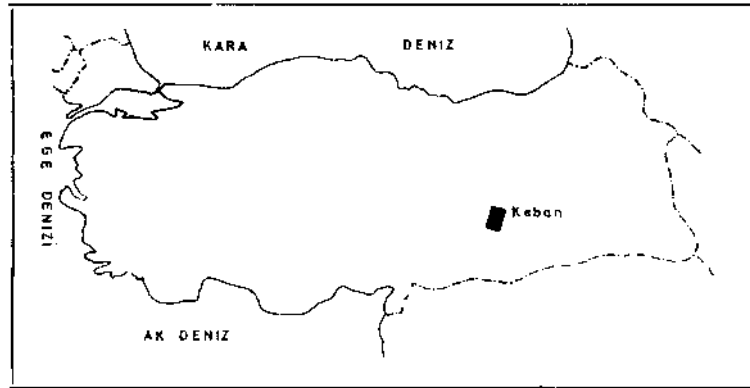
In the Keban area there have been found several materials suitable for different kinds of method suggested by others, but their application and reliability have long been in argument. Since the majority of them are of limited reliability, the combined results for the area will give only an approximate temperature range within which most of the minerals are believed to have formed. Among the methods applied are the iron content of sphalerite, the iron deficiency in pyrrhotite and d_{102} spacing of arsenopyrite. The indirect results obtained from microscopic studies will also be given in addition to the direct results. The contact metasomatic deposits at the Zereyandere and Kebandere section unlike the main sulphide deposit, have no minerals from which temperatures may easily be estimated. Although approximate, a heat flow estimate provides a general idea for the maximum temperatures which were present within the contact zone. Some microscopic evidence appears also to be helpful in connection with the heat flow estimation.

TEMPERATURE OF THE MAIN SULPHIDE DEPOSIT

Sphalerite as a geothermometer

Although controversial, this method has been used widely. Kullerud (1953) has shown experimentally that the amount of iron present in solid solution in sphalerite is a function of the temperature of formation, provided that excess iron is available at the time of formation. In the presence of excess sulphur, iron will form pyrite rather than enter into sphalerite. As it is known now, the sphalerites are characterized by high content of iron which indicates sulphur deficiency; excess sulphur would have formed pyrite instead of iron in sphalerite. In many places, however, sphalerite is found in what appears to have been in equilibrium with pyrrhotite and chalcopyrite at the time of formation. Since neither pyrrhotite nor chalcopyrite are sulphur-saturated minerals their presence indicates that sulphur was not in excess during the deposition of associated minerals.

Sphalerite samples purified by means of flotation techniques were used for the cell-size measurements from which the temperature estimations have been done.

The cell-size values were determined using a wide-angle focusing XRD and a Phillips powder camera (114.59mm in diafaeter) loaded with Industrial-G fast X-ray film. The estimated temperatures of crystallization of these sphalerites uncorrected for confining pressure are summarized in Table 1.

Table - 1

Sample no.	Cell-size A°	Mol. % FeS	Temp. °C
109	5.4185	20.00	620
120/1	5.4169	16.25	540
229	5.4172	16.90	550
B-N/1	5.4184	20.00	620
165	5.4168	16.25	540

The trace element analysis for sphalerite shows a maximum individual concentration of 4425 ppm Mn, 1092 ppm Cd and a maximum combined concentration of 5215 ppm of trace elements in sphalerite. Studies of Kullerud (1953) and Skinner (1959) on the effect of Mn, of Kullerud (1953) on the effect of Cd, and of Toulmin (1960) on the effect of Cu on the sphalerite-pyrrhotite equilibrium relations prove that no correction is required to the temperature estimates owing to the low concentration of these elements.

Iron deficiency in pyrrhotite

Arnold (1962) has recently determined experimentally the solvus relationship between hexagonal pyrrhotite and pyrite. According to him the percentage of iron in pyrrhotite was temperature dependent, but insensitive to pressure. The temperature values obtained from the pyrrhotite geothermometer from the Highland-Surprise Mine, where sphalerite occurs along with the pyrrhotite geothermometer, were found to agree well with values obtained from the sphalerite. Subsequently Buseck (1962) and Kullerud *et al.* (1963) have attempted to apply the method to monoclinic pyrrhotite-pyrite assemblages. Following the suggestions some arguments, mainly regarding the minor elements in pyrrhotite, have been raised by some others. Among them Sawkins *et al.* (1964), with the aid of fluid inclusions study of fluorite and quartz, and sphalerite as a geological thermometer, have proved that the pyrrhotite is not safe enough to use as a thermometer. He found a maximum temperature far below 250°C, below which no pyrrhotite is known to form. On the other hand pyrrhotite itself indicates a temperature between 450°C and 530°C.

Despite the above arguments, an attempt was done on two natural pyrrhotites, associating with pyrite from the main sulphide deposit. The d_{102} values after correction for shrinkage give d=2.090 A° and d = 2.081 A° corresponding to 49.62 atomic percent iron and 48.68 atomic percent iron respectively. But on the diagram of the equilibrium relations in the system FeS-S prepared by Arnold, there is a gap for iron content over 47.50 atomic percent. Therefore the pyrrhotite from the main sulphide deposit of the Keban area was found not applicable as a geological thermometer.

Arsenopyrite-pyrite-loellingite relation

Clark (1960) has experimentally shown that d_{131} spacing of arsenopyrite is a function of variation of the arsenopyrite composition, namely d_{131} spacing is related to the temperature at which the arsenopyrite with known composition forms. Therefore measurement of d_{131} spacing makes possible an estimation of its temperature of formation. Arsenopyrite, pyrite and loellingite coexist in many polished specimens taken from the different localities through the underground workings, and four of them were determined by the method described by Clark on a Phillips XRD. The temperature results are tabulated in Table 2.

Table - 2

Sample no.	d_{131} A^o	Temp. °C
106	1.6330	450
122	1.6334	472
146	1.6332	465
154	1.6332	465

Four analysed samples give an arithmetic mean of 461°C for a temperature of formation. This value is lower than the critical value of 491°C above which pyrrhotite forms instead of pyrite. These results are all obtained from arsenopyrite shown by microscopic study to be of later formation than associated sphalerite. Arsenopyrite of earlier formation than sphalerite has also been found. However, such arsenopyrite exists alone and so could have formed at any temperature between 300°C and 702°C.

Microscopic evidence

The ore microscopy study provides some evidence for formation temperatures of certain mineral assemblages. Of them unmixing temperatures given by exsolution patterns are of primary importance. Temperatures deduced from this type of evidence range from 78°C to 550°C at most.

The presence of covellite lamellae as an exsolution within neodigenite indicates a temperature above 78°C (Palache et al., 1944). Coexistence of blaubleibender covellin with covellite marks a temperature below 157°C (Moh, 1964). Valeriite exsolution in the chalcopyrite is attributed to the temperature of 225°C by Borchert (1934) who also establishes a temperature of formation of 500°C for bornite exsolution in chalcopyrite. Despite lack of agreement on the figures for sphalerite-chalcopyrite assemblage provided by Buerger (1934), Borchert (1934), and Schwartz (1931), 350°-400°C, 550°C and 650°C, respectively, the average temperature of sphalerite found by its iron content is in good agreement with what is suggested by Borchert. In short, the microscopic evi3ence suggests a progressive decrease in temperature at which the metallic mineral assemblage of the main sulphide deposit formed.

TEMPERATURE OF THE CONTACT METASOMATIC DEPOSIT

Heat flow estimation

Although approximate, a heat flow estimate gives a general idea of the upper temperature that existed within the ore-bearing contact zones around the igneous rock. According to Jones (1934) and Lovering (1935), the temperature gradient away from the contact into the country rock is sharp at the beginning but it rapidly becomes less as the country rocks are heated. The temperature at the contact, stays fairly invariable for a long period in course of cooling of the igneous rock and the country rock is warmed. It is explainable by a «heat wave» running outward from the corresponding igneous rock (Jones, 1934). This heat wave moves slowly, and warms the country rock rapidly; after the rocks are warmed up, the cooling is slow. From this in a relative distance within the contact zone, temperatures are similar to those in the contact between the host and igneous rock. Moreover the temperature in contact is relatively higher than any spots in the country rock. If a temperature in the contact is known, this provides an upper temperature at which the minerals formed.

By using the table developed by Jaeger (1957) the contact temperature for an igneous rock with a melting range between $1200° - 700°C$ is $691°C$. This value considers the latent heat of solidification and heat loss by volatiles. Although the figure appears to be much higher for an igneous rock having a short melting range, this may be accepted as being the maximum temperature for hottest part of the rock. This maximum temperature is in broad agreement with the $743°C$ suggested by Kullerud for the association pyrite/garnet.

Lovering (1955) suggests that a temperature of approximately between $850°C$ to $900°C$ is about suitable for quartz monzonite or granite magmas in the hotter parts of the intrusive body; this suggestion appears to cover syenite magmas as it lies between the two magmas in the classification. According to Lovering, the leading edge of a sill, dyke or other intrusions, may be as low as $700° C$.

Microscopic evidence

The contact metasomatic deposits provide comparatively limited evidence for estimating the temperature of formation. Although it is very rare chalcopyrite grains include valeriite exsolution indicating a minimum temperature of $225° C$. The widespread presence of pyrite and of zoned anisotropic grossularite correspond to a temperature of formation below $743° C$ (Kullerud, 1953). The formation of wollastonite indicates a temperature range from $660° C$ to $800° C$.

In short, in the light of different suggestions, a maximum approximate temperature of $700° C$ for the contact metasomatic deposits is suggested as the most suitable temperature below which the metallic minerals form in order of decreasing temperature.

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