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Volcanologic concerns of the siliceous metasedimentary xenoliths included in historic lava-flows of Lanzarote (Canary Islands)

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Abstract

A large number of siliceous metasedimentary xenoliths appears in some historic lava flows of Lanzarote (Canary Islands). In most cases those xenoliths are constitute of α-cristobalite, α-tridymite, calc-silicates (wollastonite 2M and diopside) or paragenesis of calc-silicates and α-cristobalite. The presence of microfossils ghosts, detrital grains and sedimentary structures reveal that all these xenoliths are formed by thermal metamorphism of sedimentary rocks, many of them rich in radiolarians. The variable amount of radiolarians and the other components (calcareous microfossils, micrite, mudstones) determine the formation of silica minerals or calc-silicates during the thermal transformations. The thermal transformation would have taken place in the contact of the sedimentary rock with an intrusive body or shallow magmatic chamber corresponding to an endogenous stage of the 1730-1736 eruption. The high temperatures reached in this process, together with the mineralogical composition of the siliceous sedimentary rock, rich in opal-CT (porcellanite), determine the formation of tridymite and cristobalite instead of quartz. The information obtained from the study of the transformed sedimentary xenoliths also allows us the quantification of depth and temperature of the magmatic chamber, which coincide with those obtained by other methods.

Riassunto

Un grande numero di xenoliti silicei metasedimentari si ritrovano in alcune colate di lava storiche di Lanzarote (Isole Canarie). In molti casi questi xenoliti sono formati da α-cristobalite, α-tridimite, calc-silicati (wollastonite 2M e diopside) o paragenesi di calc-silicati e di α-cristobalite. La presenza di fantasmi di microfossili, granuli detritici e tessiture sedimentarie indicano che gli xenoliti si sono formati per termometamorfismo di rocce sedimentarie, molte delle quali ricche in radiolari. La quantità variabile di radiolari e di altri componenti quali microfossili calcarei, frammenti di micrite o di pelite, determina la formazione di minerali silicei e calc-silicatici durante le trasformazioni termiche. Queste sarebbero avvenute per contatto con un corpo intrusivo o una camera magmatica poco profonda durante la fase endogena dell'eruzione del 1730-1736. Le alte temperature insieme con la composizione mineralogica delle rocce sedimentarie ricche in CT-opale (porcellanite), determina la formazione di tridimite e cristobalite invece che di quarzo. Le informazioni ottenute hanno permesso di quantificare la profondità e la temperatura della camera magmatica, che coincidono con quelle ottenute con altri metodi.

1. Introduction

An important feature of the volcanic eruptions which occurred in Lanzarote (Canary Islands) from 1730 to 1736, was the abundace of xenoliths included in the lava flows. The majority of these xenoliths corresponds to ultramafic rocks – peridotites, dunites, wherlites – mostly of deep origin (Sagredo, 1969). Another group of xenoliths corresponds to sedimentary rocks – limestones, sandstones, mudstones and radiolarites – from the NW Africa marginal basin (Araña & Ortiz, 1991).

Among the xenoliths of sedimentary origin, the siliceous rocks (which constitute 80%) stand out. Some mineralogical and textural characteristics of these siliceous xenoliths studied here have not been previously described, or they are rare in nature. This fact permits us to say that such features are directly related to the particular volcanic scenario of the eruptions of 1730-1736 in Lanzarote (Ortiz et al., 1986a).

2. Geological Setting

Lanzarote, the most eastern of the Canary Islands, is mainly constituted of alkaline basalts. In this island the accumulations of subhorizontal lava flows from ancient formations (from 16 to 6 m.y.) and the several cones from more recent formations are outstanding (Fuster et al., 1968). There have been two historical eruptions on the island (1730-1736 and 1824), being the eruption of the XVIII century one of the most important basaltic eruptions suffered humanity. The eruption that took place between the years 1730 and 1736 threw 1 km^3 of lava covering about 200 km^2, almost a third of the island. The effusive vents were aligned on a 14 km fissure, following an ENE-WSW axis which coincides with one of the main volcano-tectonic trends of the archipelago. In the area of this eruption, there are strong surface thermal anomalies that have been correlated with a shallow residual magma chamber (Araña et al., 1984; Ortiz et al., 1986b).

The initial lavas of the eruption, emitted in the central part of the fissure, are far richer in ultramafic xenoliths, whereas sedimentary ones are almost always restricted to later emissions at the edges of the fissure. These sedimentary xenoliths are mainly of a clear colour and their size very rarely exceeds 5 cm. There is no reaction of any kind observed between the xenolith and the present host rock except for a slight increase in the porosity of some of them. The xenolith distribution is not even, but on the contrary higher punctual concentrations can be appreciated as irregularly distributed, which makes it impossible to calculate their real proportion.

3. Petrology

After a mineralogic study by petrographic microscope and X-ray diffraction, three kinds of xenoliths are defined in relation to the main minerals.

1) Xenoliths with cristobalite
2) Xenoliths with tridymite
3) Xenoliths with calc-silicates

Xenoliths with cristobalite

The main mineral is low-cristobalite (up to 95%) but in some xenoliths, quartz (normally less than 20%, and exceptionally up to 50%) or calc-silicates (diopside or wollastonite 2M-parawollastonite up to 20%) may appear. By X-ray diffraction this cristobalite has a strong peak at 4.05 Å. The height/width ratio of this peak at 50% intensity varies between 64 and 200. These values are higher than those found in low cristobalite of obsidian vugs and hydrothermal springs (Henderson et al. 1971). The opal-CT, that usually is present in sedimentary siliceous rocks, does not appear in these xenoliths, as its indicative weak peak at 4.32 Å (Jones & Segnit, 1971) has not been identified. Thin section observations show these xenoliths are composed of a mosaic of criptocrystalline cristobalite that differs from the typical textures of the opal – CT (Bustillo, 1982). Radiolarians, silica spicules, microforaminifers and in some cases, echinoderms and ostracodes may appear between the cristobalitic groundmass. Other times only ghosts of those microfossils appear, and in some cases it is not possible to see any traces of them. Detrital silt quartz, may rise up to 20%. In the xenoliths with a higher quartz content, this mineral is also a component of the previously mentioned microfossils.

Xenoliths with tridymite

These are composed of low tridymite (more than 90%) and quartz (up to 10%). By X-ray diffraction this low tridymite differs from standard low tridymite (PDF 18-1170, Smith, 1974) because the strongest peak is at 4.32 Å.

Under optical microscopy the tridymite crystals present various sizes, forming from microcrystalline up to macrocrystalline mosaics (>50μm). The tridymite crystals are rectangular or wedge-shaped and are sometimes arranged in aggregates (rosettes). These textures were found by Schneider & Florke (1985) in silica brick after a firing process at 1.200°C and may correspond with the disordered tridymites defined by these authors.

Scattered patches of black radiolarites or isolated ghosts of radiolarians and occasionally microforaminifers may appear in these xenoliths. In these cases the radiolarians are constituted of microquartz (less than 30 μm).

Xenoliths with calc-silicate

Their composition is variable and two kinds of calc-silicates are found: wollastonite 2M (parawollastonite) and diopside.

Some xenoliths are constituted of wollastonite 2M-parawollastonite (100-30%), quartz (50-5%), α-cristobalite (40-5%), dolomite (up to 15%) or calcite (up to 10%). Others are formed by diopside (85-70%), quartz (up to 30%) and α-cristobalite (up to 10%).

Under optical microscopy these xenoliths exhibit criptocrystalline textures with ghosts of radiolarians, microforaminifers and other calcareous microfossils of doubtful identification because of the transformations. The radiolarians are constituted of microquartz, calcite or microcristalline parawollastonite.

Silt and fine sand grains (up to 30%), mainly of quartz, appear isolated or forming millimetric layers.

4. Discussion and conclusions

Sedimentary primary rocks of the xenoliths

The presence of microfossil-ghosts, detrital grains and sedimentary textures (bioturbation, erosive contacts, fine stratification) reveals that all these xenoliths come from sedimentary rocks, many of them with radiolarians. Sediments bearing radiolarians exist in the oceanic crust, according to the samples cored in the DSDP in continental margin sediments of Northwest Africa. The age of these siliceous rocks range from Oxfordian to Early Miocene, but they are more frequently found in the Paleocene-Eocene (Von Rad et al., 1977). Rocks of such nature crop out in a section of about 30 m of the Basal Complex (from Albian to Upper Cretaceous) of the nearby island of Fuerteventura (Robertson & Bernoulli, 1982). Also, near the bottom (2700 m.) of a geothermal well, drilled in Lanzarote, sicilificated limestones of Paleocene age have been found (Sanchez & Abad, 1986).

Foraminifers from some xenoliths, suggest a range in age from Cenomanian to Lower Miocene. On the other hand the age of the radiolarians determined in some xenoliths is Paleocene-Eocene. Because of this, we think that the xenoliths may come from the series with siliceous sedimentary rocks found in the Fuerteventura Basal Complex and the geothermal well drilled in the same area of the Lanzarote historical eruptions. Due to the great sedimentary gaps, these age ranges can be translated in a few hundreds or even ten s of meters of sedimentary column.

The sedimentary rocks described in the Fuerteventura Basal Complex and in the near drills of the DSDP, present a normal silicification, typically diagenetic (quartz and opal-CT); on the contrary, the silicification of the xenoliths can only be explained by a thermal transformation able to generate different silica and calc-silicate minerals (α-cristobalite, α-tridymite, wollastonite 2M and diopside). In fact, in low temperature systems the deposited siliceous tests (opal-A) are partially dissolved and silica overgrowths may precipitate, (refered to as opal A', by Hein et al., 1978). As these amorphous silica polymorphs are buried deeper in the sedimentary column, a less soluble disordered cristobalite-tridymite phase, (opal-CT), forms. Later the opal-CT recrystallizes to quartz. The transformation opal A-A' → opal-CT → quartz is a response to age, temperature, burial depth, and lithofacies (see reviews in Bustillo, 1980; Williams et al., 1985 and Williams & Crerar, 1985).

It is well known that the mineralogical composition of the sedimentary siliceous rocks depends on the general diagenetic evolution of the siliceous sediments. The diagenetic path of the continental margin sediments of Northwest Africa (Von Rad et al, 1977) shows that from the Upper Cretaceous to the Eocene, the amount of opal-CT in siliceous rocks is relatively high. This means that according its diagenetic history, due to age, the sedimentary siliceous rocks, before their thermal transformation, were porcelanites (formed mainly by opal-CT) and cherts (formed mainly by quartz). In our case, as the silica xenoliths constituted only of quartz are very scarce (less than 2%), we think the majority of its primary silica rocks were mostly porcelanites, because in thermally transformed cherts the quartz should have undergone only crystallographic changes (Joesten, 1983; Gurky & Gurky, 1989; Keller et al., 1985).

Thermal transformations

The study of the mineral paragenesis of the xenoliths reveals the existence of four mineral assemblages (Fig. 1), without taking into account the amount of quartz, because it is considered that quartz is always a remnant of the primary sedimentary rock (detrital grains or crystals included in the microfossils). These mineral assemblages are the consequence of the thermal gradient induced in sedimentary layers, mainly siliceous, by the intrusion of a magmatic body.

The variable amounts of radiolarians and the other components in the primary sedimentary rocks determine the composition of metamorphic

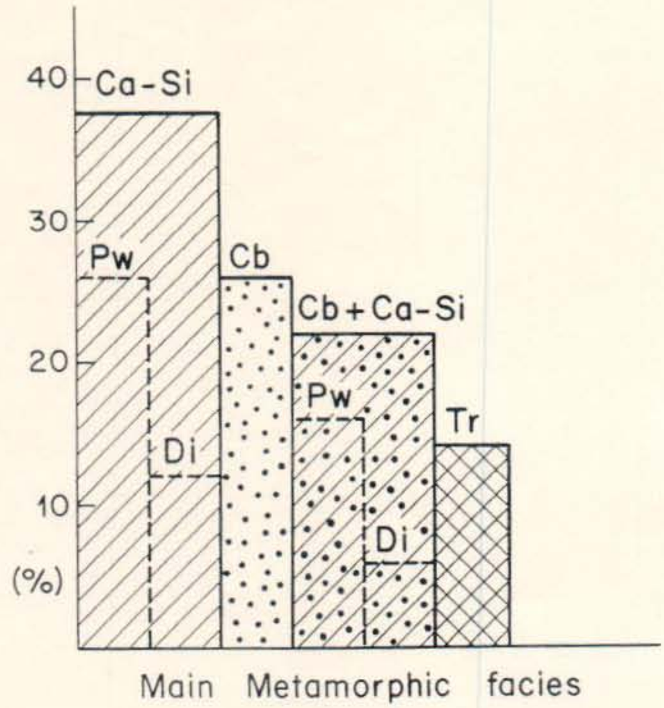

Fig. 1 – Relative percentage of the main mineral assemblages presented by the xenoliths.
Pw = Wollastonite 2M (Parawollastonite), Di = Diopside, Ca-Si = Calc-silicates, Cb = Cristobalite, Tr = Tridymite, Cb+Ca-Si = Cristobalite and Calc-silicates.

xenoliths. Cristobalitic and tridymitic xenoliths come from radiolarites. When the radiolarites contain carbonates (microfossils or micrite) parawollastonite is formed. The diopside that often appears in xenoliths with detrital quartz, could come from calcareous radiolarites interbedded with magnesium rich mudstones and sandstones.

It is difficult to know the stability fields of the different minerals produced in this process as there is a lack of experimental diagrams. According to Deer et al. (1978), the ability of cristobalite and tridymite to occur in unstable forms, outside their equilibrium field, makes it difficult to draw definitive conclusions as to conditions of formation. Related to parawollastonite there is very little data, but if we consider that Phillips & Griffen (1981) point out that it is found together with wollastonite, its stability field could be used, and this means a lower limit of 500-600°C. (Tanner et al. 1985).

The data known about opal siliceous rocks affected by thermal metamorphism (Murata et al, 1979, Kastner & Siever, 1983), showed that the rocks made up of opal-CT suffer diagenetic or metamorphic changes which are expressed in a higher order of the opal-CT, or in a change to quartz, always depending on the temperature. This process has not come out in the studied xenoliths because quartz is neither an important nor frequent mineral. Opal-CT does not exist and morover other infrequent siliceous phases such as tridimite appear.

In relation to the above explanation, in the sedimentary rocks near the contact with the magmatic intrusion, temperatures must be over 875°C. In fact, under such thermal conditions and with a pressure of 2 kb, tridymite seems to be the only stable siliceous phase (Black, 1954) and the opal-CT rocks are transformed into tridymite. Terrigenous quartz and quartz microfossils, meaning less the 10% of the rock, do not alter and just suffer crystallographic changes as it happens in cherts forced to a high degree metamorphism. In these temperature ranges, between 900° and 800°C, a partial fusion might be possible, and in this sense Lacroix (1946) describe tridymite formed after a quarzite fusion. Later on, during cooling, this tridymite will invert into a α-tridymite. The scarce xenoliths presently made up of α-tridymite, are those that contain less relics from the sedimentary rock. This agrees with the fact that tridymite xenoliths are those which suffered the most intense heating.

A few meters farther from the contact host rock-magmatic body the temperature decreases (i.e. to 700-500°C), corresponding to this zone xenoliths composed soley of cristobalite or those formed by cristobalite calc-silicate paragenesis. As result of the effect of these temperatures the initial opal-CT was arranged producing a cristobalite even more ordered than the one generated in hydrothermal processes at temperatures between 200° and 600°C (Henderson et al., 1971).

The α-cristobalite, in which the initial opal-CT has been transformed would be stable in every zone with temperatures lower than 800°C for that reason this is the most frequent mineral in xenoliths coming from sedimentary silica rocks.

5. Volcanic scenario

It is difficult to accept that the thermal transformation here described, could have been produced during the transport of the xenoliths in the ascending lavas of the 1730-1736 eruptions. In fact, according the high velocity of this ascent (Ortiz et al., 1986a; Armienti et al., 1991) there was no time – only some hours or a few days – to generate the mineral paragenesis presented by the xenoliths. Most probably, the existence of a thermal transformation of the sedimentary rocks took place in its contact with an intrusive body or shallow magmatic chamber corresponding to an endogenous stage of the 1730-1736 eruptions. In fact, as it was already said, the meta-sediments sampled by these eruptions correpond to a very short section of the materials through which the 1730-1736 conduits passed. If the thermal transformations had been induced by an old chamber, it is unlikely that a specific part in the rim of such supposed old chamber, were the only crustal material selectively sampled along the 1730-1736 conduits.

The metamorphic gradient induced by the basaltic intrusion in the host rock, mainly depends on the characteristics of these rocks which have already been described in previous sections. This gradient will also depend on depth, shape and volume of the intrusion and on the time in which the temperatures of the intrusive body remained or evolved. The information obtained from the study of transformed sedimentary xenoliths also allows us to quantify some of these parameters of special volcanologic interest.

The depth of the intrusion can be indirectly calculated by knowing the age and thickness of the sediments affected by the thermo-metamorphic process. The lower limit of these affected sediments cannot be very far under the Mesozoic-Cenozoic horizon which is located at about 3000 m. b.s.l. under Lanzarote due to the uplifting of crustal blocks (Araña & Ortiz, 1991). As for the upper limit, one of the xenoliths shows interlayers of volcanic tuffs and radiolarites with foraminifers which could be correlated to cores extracted at 2600 m. deep in the near geothermal drilling (Sanchez & Abad, 1986). In agreement with these data, the intrusion could have been located at depths of 3000±500 m. which is on the limits determined through other methods and models for shallow magmatic chambers of the 1730-1736 eruptions (i.e. see Ortiz et al., 1986b and Pavia et al., 1977) whose residual temperature justifies the current strong superficial geothermal anomalies (Araña et al., 1984).

The shape of the intrusion cannot be precised either, although taking into account the characteristics of this volcanism, it can be imagined as a tube-shaped body, locally widened with respect to the upper channels of the eruption and to those used by the magma to ascend from its sources. The horizontal section of the magmatic intrusion could be elliptical with an axis of about 14 km, equivalent to the length of the eruptive fissure, and a minor axis of 1 km., also equivalent to the width of the 1730-1736 eruptive fringe. In order to calculate the volume of the intrusive body or magmatic chamber, it is necessary to know the height of this chamber whose limits are given by the depth before determined (3000±500 m.). Nevertheless, this parameter admits a wide evaluation margin as the intrusion is not necessarily static and could move from the lower limit to the upper one in a period of time which would have occured before the eruption. In any case, it would be enough if the chamber had a height of 200 m to stock a quantity of magma double than the volume of lava issued in the eruption (1 km^3).

Regarding the time of activity of this chamber, it seems normal to consider it under a steady state, for thermal purposes, most of the time the eruption lasted (6 years). Nevertheless it is very possible that the chamber was established long before. As a matter of fact, in a latter and much less important eruption (year 1824) in the same area, seisms were felt ten years before, which indicates activity in a shallow chamber. For that reason it seems possible that the chamber had at least an endogenous activity of some years. By the contrary, considering the less favourable case, the influence of this magmatic chamber should be restricted to the long non-eruptive periods (many months) which occurred between 1730 to 1736.

There are several works that come from the classical formulations of Carslaw & Jaeger (1959) which permit us to evaluate these thermal-metamorphic processes induced by an intrusive body. In an analogical model elaborated by Rubia et al. (1970) the intrusion of a basaltic dike of 100 m thick and a temperature of 1100°C was considered in a zone of similar thermal diffusivity located at 500 m depth, with an initial temperature in accordance with the terrestrial gradient. It was also established that the dike started cooling down in the very moment of its intrusion, which implies a more unfavorable case than Lanzarote's intrusion that can be considered with a stable temperature for a long period of time due to its continuous feeding from the magmatic source. Following this model, whose parameters and conditions can be considered valuable in the case we are studying, 6 years after the intrusion, temperatures in the host rock would be 530°C, 247°C and, 267°C at a distance from the intrusion of 5, 15 and 25 m. As the change of the state in the basaltic magma occurs below 800°C, the temperature in the contact zone must be considerably higher, during the period (months or years) that the chamber was being fed from deep sources. Temperatures reached in the zone of the host rocks nearest to the contact with the intrusion, should have been enough to produce the total fusion, while in any of the farther areas the mineral transformations previously described could be produced not only depending on the induced temperature, but on the original composition of the deposit.

Finally, it seems interesting to point out that a total or partial fusion of the radiolarian layers (almost 100% of SiO_2) should provoke a contamination in the magma. Such a possibility should be taken into account in future studies of the anomalous silica enrichment (Ibarrola, 1969; Brandle & Fernandez, 1979) observed in the lavas emitted during the final phases of the eruptions of 1730-1736 in Lanzarote.

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BIBLIOGRAPHY

Araña, V., Díez, J.L., Ortiz, R. & Yuguero, J. (1984): Convection of geothermal fluids in Timanfaya volcanic area (Lanzarote, Canary Is.). Bull. Volcanol, 43-3, 667-677.

Araña, V. & Ortiz, R. (1991): The Canary islands: Tectonics, Magmatism and Geodinamic framework. In "Extensive Magmatism and Structural Setting" A. Kampunzu & P. Lubala (Eds). Springer Verlag. Heidelberg, 209-249.

Armienti, P., Innocenti, F., Pareschi, M.T., Pompilio, M. & Rocchi, S. (1991): Crystal population density in not stationary volcanic systems: estimate of olivines growth rate in basalts of Lanzarote (Canary Is.). Mineralogy and Petrology, 44, 181-196.

Black, G.P. (1954): The significance of tridymite in igneous and metamorphic petrogenesis. Miner. Mag., 227, 512-518.

Brandle, J.L. & Fernández, S. (1979): On the non-existence of tholeiitic series in the Canary Islands. Chem. Geol., 26, 91-103.

Bustillo, M.A. (1980): Diagenesis de silex. Rev. Inst. Invest. Geol., Barcelona, 34, 237-248.

Bustillo, M.A. (1982): Ageing features in inorganic continental opals. Est. Geol., 38, 335-344.

Carlslaw, H. & Jaeger, J. (1959): Conduction of heat in solids. Oxford Univ. Press, London, 510 pp.

Deer, WA, Howie, R.A. & Zussman, J. (1978): Rock-forming Minerals vol 4 – Framework silicates. Longmans, London, 435 pp.

Fuster, J.M., Fernández, S. & Sagredo, J. (1968): Geology and Volcanology of the Canary Islands. Lanzarote. I. Lucas Mallada, CSIC, Madrid, 177 pp.

Gursky, J.J. & Gursky, M.M. (1989): Thermal alteration of chert in the Ophiolite Basement of Southern Central America. In: "Siliceous Deposits of the Tethys and Pacific Regions", J.R. Hein & J. Obradovic (Eds.), 217-233.

Hein, R.J., Scholl, D.W., Barron, J.A., Jones, M.G. & Miller, J. (1978): Diagenesis of late Cenozoic diatomaceous deposits, and formation of the botton stimulating reflector in the Southern Bering sea. Sedimentology, 25, 155-181.

Henderson, J.H., Jackson, M.L., Syers, J.K., Clayton, R.N. & Rex, R.W. (1971): Crisobalite authigenic origin in relation to montmorillanite and quartz origin in bentonites. Clays and Clay Min., 19, 229-238.

Ibarrola, E. (1969): Variation trends in basaltic rocks of the Canary Islands. Bull. Volcanol. 33, 729-777.

Joesten, R. (1983): Grain growth and grain boundary diffusion in quartz from the Christmas Mountains (Texas) Contact Aureola. Am. J. Sci., 283-A, 233-254.

Jones, J.B. & Segnit, E.R. (1971): The nature of opal I. Nomenclature and constituent phases. J. Geol. Soc. Australia, 18, 57-68.

Kastner, M. & Siever, R. (1983). Siliceous sediments of the Guayanas basin: the effect of high thermal gradients on diagenesis. J. Geol. 91, 629-641.

Keller, W., Stone, C.H.G. & Hoersch, A.L. (1985): Textures of Paleozoic chert and novaculite in the Ovachita Mountains of Arkansas and Oklahoma and their geological significance. Geol. Soc. Am. Bull., 96, 1353-1363.

Lacroix, A. (1946): Sur un nouveau cas de formation de tridymite par fusion d'une enclave quartzique d'un volcan andesitique. C. R. Acad. Sci Paris, 223, 409.

Murata, K.J., Dibblee, J.R. & Drinkwater, J.L. (1979): Thermal effects of large bodies of intrusive Serpentinite on Overlyin Monterey Shales, Southern Diablo Range, Area, California. U.S. Geol. Survey Prof. Paper, 1082, 8 pp.

Ortiz, R., Araña, V. & Valverde, C. (1986a): Aproximación al conocimiento del mecanismo de la erupción de 1730-1736 en Lanzarote. An. Fisica, B 82, 127-142.

Ortiz, R., Araña, V., Astiz, M. & García, A. (1986b): Magnetotelluric study of the Teide (Tenerife) and Timanfaya (Lanzarote) volcanic areas. J. Volcanol. Geotherm. Res. 30, 357-377.

Pavía, J., Ortiz, R., Banda, E., Rodríguez, E. & Araña, V. (1977): Perfil sísmico del NE de Lanzarote. In: "Perfiles sísmicos profundos en España", A. Udías (Ed), Madrid, 135-144.

Phillips, W.R. & Griffen, D.T. (1981): Optical Mineralogy. The monopaque minerals. W.H. Freeman and Company. San Francisco, 667 pp.

Robertson, A.H.F. & Bernoulli, D. (1982): Stratigraphy, Facies and Significance of Late Mesozoic and Early Tertiary Sedimentary Rocks of Fuerteventura (Canary Islands) and Maio (Cape Verde Islands). In: "Geology of NW. African Continental Margin" V. Von Rad, K. Hinz, M. Sernthein & E. Seibold (Eds.) Springer-Verlag. Berlin, 498-528.

Rubia, J., Doria, J., Osorio, A., Calleja, J. & Araña, V. (1970): Distribución transitoria de temperaturas en las proximidades de una capa magmática intrusiva en un proceso de enfriamiento. Estudios Geol, 26, 273-280.

Sagredo, J. (1969): Origen de las inclusiones de dunitas y otras rocas ultramáficas en las rocas volcánicas de Lanzarote y Fuerteventura. Estudios Geol, 25; 189-223.

Sánchez, J. & Abad, J. (1986): Sondeo geotérmico Lanzarote-1. Significado geológico y geotérmico. An. Fisica, 82: 102-109.

Schneider, H. & Flörke, O.W. (1982): Microstructure, chemical composition and structural state of tridymite. N. Jb. Mineral. Abh. 145, 280-290.

Smith, D.K. (1974): Joint Committee on Powder Difraction Standards Selected powder diffraction data for minerals. Data Book, first edition.

Tanner, S.B., Kerrich, DM, & Lasaga, A.C. (1985): Experimental kinetic study of the reaction: calcite + quartz wollastonite + carbon dioxide from 1 to 3 kilobars and 500° to 850°C. Am. J. Sci. 285, 577-621.

Von Rad, V., Riech, V. & Rösch, H. (1977): Silica Diagenesis in continental margin sediments of Northwest Africa. Initial Reports of the DSDP, 41, 879-905.

Williams, L.A., Parks, G.A. & Crerar, D.A. (1985): Silica Diagenesis I. Solubility controls. J. Sedim. Petrol. 55, 301-311.

Williams, L.A. & Crerar, D.A. (1985): Silica Diagenesis II. General Mechanisms. J. Sedim. Petrol. 55, 312-321.