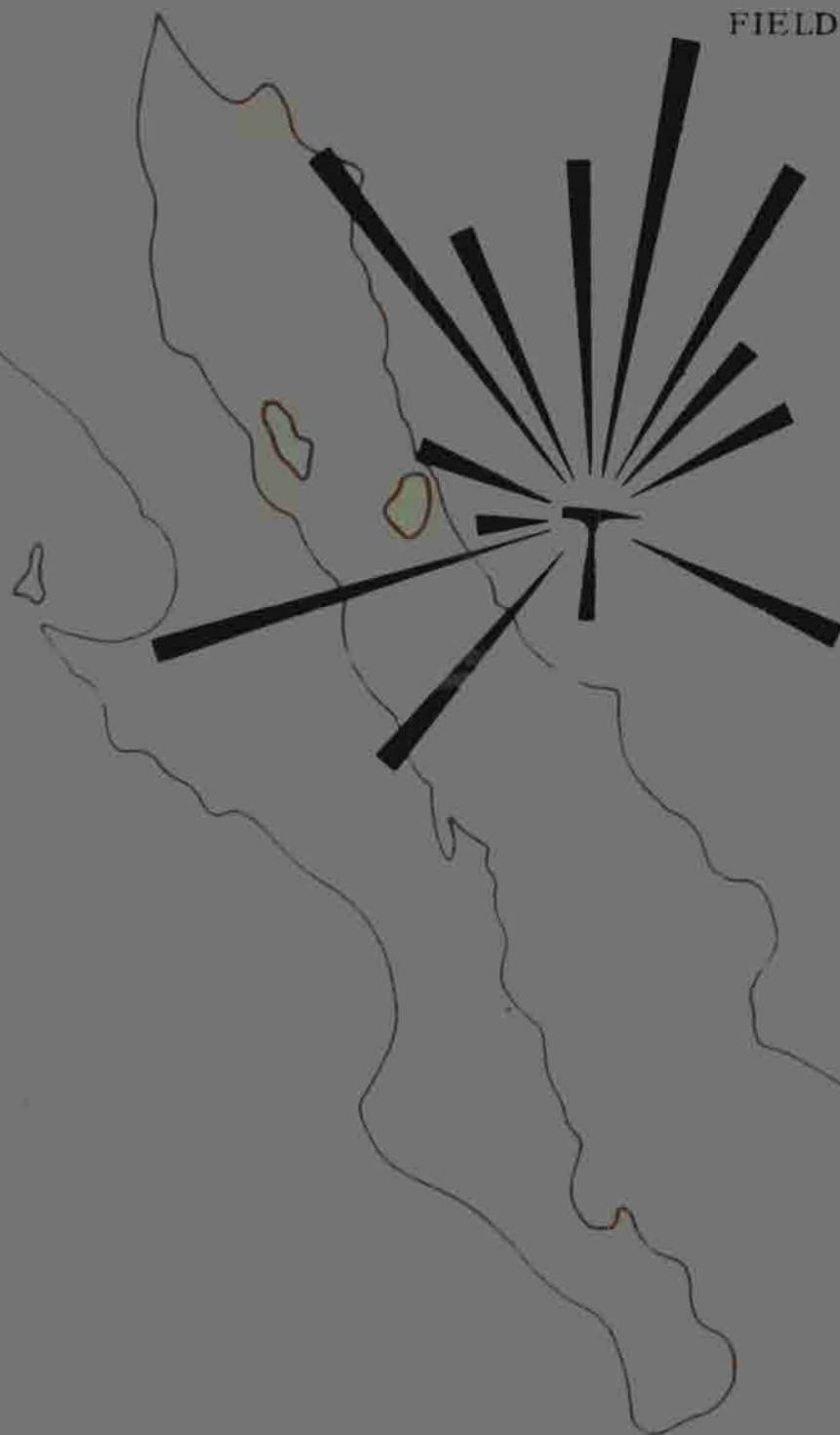


Geology of Northwestern Mexico and Southern Arizona

FIELD GUIDES AND PAPERS



Edited by Luc Ortlieb and Jaime Roldán Q.
1981

GEOLOGY OF NORTHWESTERN MEXICO
AND SOUTHERN ARIZONA

DESCRIPTION OF FIELD TRIPS PREPARED FOR
THE GEOLOGICAL SOCIETY OF AMERICA,
CORDILLERAN SECTION 1981 ANNUAL MEETING,
HELD IN HERMOSILLO, SONORA, MEXICO, MARCH 21-30,
AND SPONSORED BY THE INSTITUTO DE GEOLOGIA U.N.A.M.
IN COOPERATION WITH THE UNIVERSIDAD DE SONORA.

Geología del noroeste de México y del sur de Arizona

Descripción de excursiones preparadas para la reunión anual de 1981
de la Geological Society of America, Cordilleran Section, efectuada
en Hermosillo, Sonora, México, del 21 al 30 de marzo, y organizada
por el Instituto de Geología, U.N.A.M., en cooperación con la
Universidad de Sonora.

EDITED BY

LUC ORTLIEB AND JAIME ROLDAN Q.

ESTACION REGIONAL DEL NOROESTE
INSTITUTO DE GEOLOGIA, U.N.A.M.
HERMOSILLO, SONORA, MEXICO.

1981

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Our first meeting of the Cordilleran Section, G.S.A., in Mexico seemed to be an excellent opportunity to organize a series of field trips covering various aspects of the geology of Sonora, Baja California and southern Arizona. The Editors would like to thank particularly all the field trip leaders and contributors who submitted camera-ready manuscripts by the due deadline (mid-January 1981); they are sorry not to have been able to include all the field trip guides.

The volume would not have been made possible without the generous help of several persons of the Estación Regional del Noroeste, Instituto de Geología, U.N.A.M. : César Jacques, Bernard Colletta, José Luis Rodríguez, Roberto Montaña, Rosa María González, Gloria Saitz and Maricela de Roldán.

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Esta es la primera reunión de la Sección Cordillerana de la G.S.A. que se celebra en México, lo cual ha dado una excelente oportunidad para que los geólogos de ambos países intercambien sus experiencias. Se organizaron una serie de excursiones, que cubren varios aspectos de la geología de Sonora, Baja California y Arizona. Los Editores expresan su agradecimiento a todos los guías de las excursiones y autores por haber enviado sus artículos preparados para su reproducción en offset en la fecha límite (mediados de enero 1981) y lamentan no haber podido incluir todos los libretos-guías de las excursiones programadas.

Este volumen fue posible hacerlo gracias a la valiosa ayuda de colaboradores de la Estación Regional del Noroeste del Instituto de Geología, U.N.A.M. : César Jacques, Bernard Colletta, José Luis Rodríguez, Roberto Montaña, Rosa María González, Gloria Saitz y Maricela de Roldán.

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GEOCHEMISTRY AND ECOLOGY OF SALT PANS AT GUERRERO NEGRO,
BAJA CALIFORNIA

FIELD TRIP No. 1 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
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1981 ANNUAL MEETING

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ABSTRACT

Just south of Guerrero Negro, the 50 km length of Laguna Ojo de Liebre concentrates sea water to about 1.4X normal. Beyond the distal end of the lagoon, these waters flooded an extensive area of sabkha. An intertidal zone of dark algal mats includes precipitation of aragonite, high-magnesian calcite, and dolomite, and minor gypsum. The higher and more extensive part of the sabkha was reached only by occasional spring tides aided by storm winds which then precipitated gypsum and halite that was accumulated up to 2 m thick. Some brine remained interstitially to crystallize additional gypsum and halite, as well as polyhalite and magnesite (no anhydrite has been found).

In recent years most of the gypsum/salt flats have been covered by concentrating ponds for the production of salt. Mineralogy and geochemistry (major and minor elements and stable isotopes), of evaporite minerals and interstitial brines, are described for one section sampled prior to construction of the ponds, and for three taken from the periphery of the present ponds.

INTRODUCTION (W.T.H.)

Guerrero Negro is halfway down the Pacific coast of Baja California, 720 km south of the USA border on Highway 1 (Fig. 1). South of Guerrero Negro (GN), around the head of the Laguna Ojo de Liebre (LOL), was once the largest sabkha complex in North America (Fig. 2). Several smaller sabkhas occur elsewhere on both coasts of Baja California (Fig. 1). Beginning in 1957 the major part of the natural sabkha at LOL has been covered with ponds for solar concentration and crystallization, in turn making this the world's largest producer of salt. Both the remnants of the sabkha and the artificial ponds are unique laboratories for the study of the geochemistry and ecology of the hypersaline environment.

A few words concerning background of this trip and its leaders. In May, 1960, F. J. Phleger of Scripps Institution of Oceanography, who had studied the lagoons of the Mexican coast for many years (Phleger & Ewing, 1962), invited Holser to examine the unique evaporite deposits that were then still extensively exposed at LOL. Following this preliminary inspection, Holser returned to LOL in August, 1961, with J. R. Bradshaw of Scripps, for extensive sampling, including four shallow core holes. Holser returned again in May, 1965, with D. J. J. Kinsman for further surface sampling; subsequently they studied the 1961 cores. One result of this field work was a description of polyhalite replacing gypsum (Holser, 1966), largely from 1960 samples. Stable isotope data were published by Holser and Kaplan (1966), and Holser (1979b). Subsequently, Phleger (1969) and Kinsman (1969) published general descriptions of the LOL evaporites. Additional, previously unpublished data from the 1961 and 1965 samplings (which were in areas now covered by evaporating ponds) are given below. Holser's work at LOL was jointly supported by the Chevron Oil Field Research Company, and their contribution is gratefully acknowledged. In 1975 Holser, then in the Department of Geology at the University of Oregon, suggested to B. J. Javor of the Department of Biology that she might extend her previous work on algal ecology to higher salinities by studying those in the GN area. Javor spent several seasons in the field, and completed a Ph.D. dissertation (Javor, 1977, 1979). Subsequently she was employed as a biologist at GN by Exportadora de Sal, S.A., and is still continuing her biological studies at LOL.

In 1979 Ortlieb and Pierre began a geochemical and mineralogical study of the evaporites of Baja California, including three cross-sections at LOL outside the newer evaporating pans. These studies, including extensive chemical and stable isotope analyses of both sediments and brines, are continuing. One paper is in press (Pierre & Ortlieb, 1980), and that material as well as previously unpublished data is included in the description below, as well as in presentations at the meeting (Pierre, 1981; Pierre & Person, 1981). Ortlieb's investigations are part of a general study of the geodynamics of the Quaternary in the region of the Gulf of California, sponsored by Misión O. R. S. T. O. M. and by the Instituto de Geología, Universidad Nacional de Mexico. Pierre's investigations of stable isotopes are sponsored by the Comité national de recherche scientifique of France, and carried out in the Laboratoire d'hydrologie et de géochimie isotopique, Université Paris-Sud, France.

In the following descriptions it was not always possible to integrate the earlier results of Holser with the more recent and intensive studies by

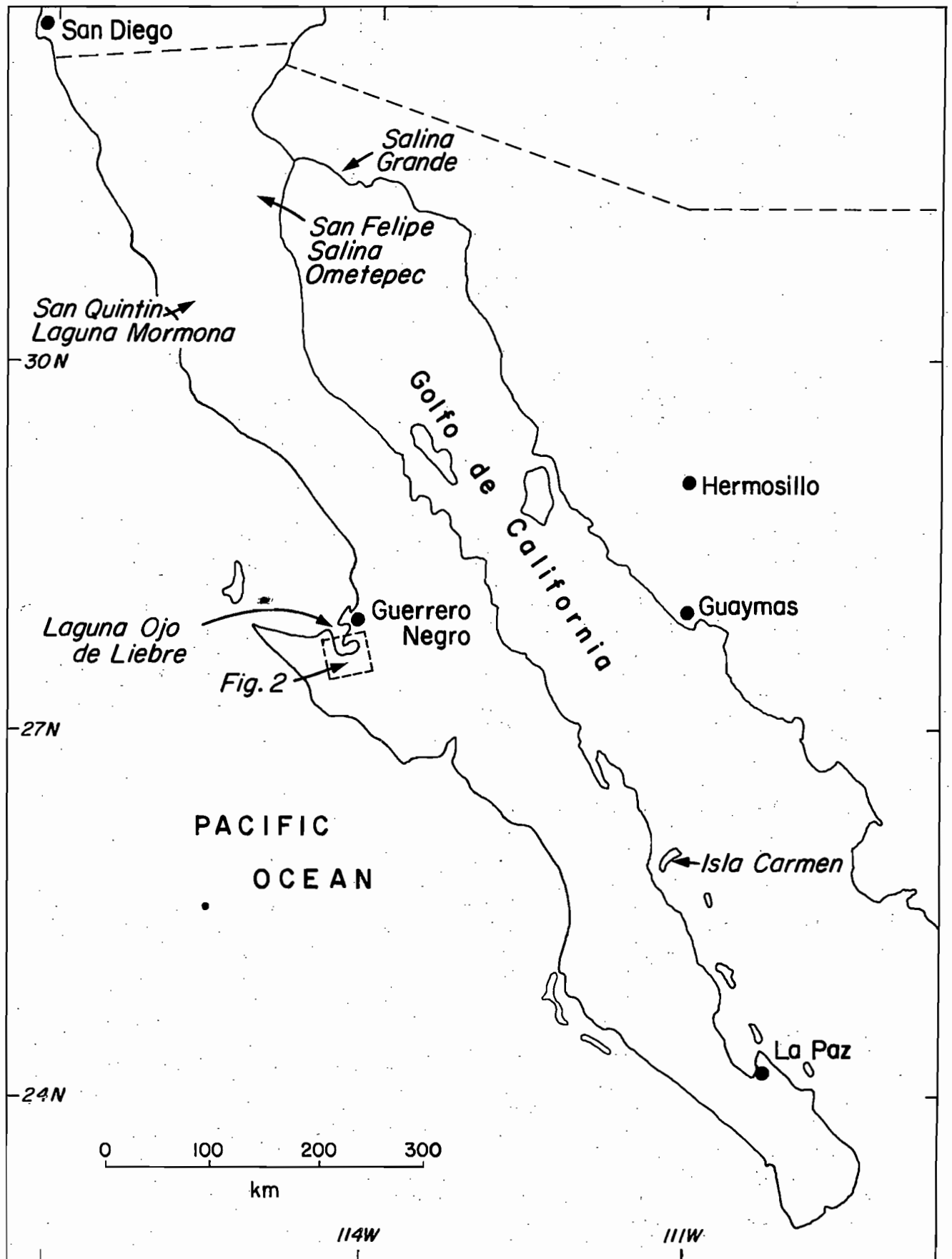


Figure 1. Index map of Baja California, locating Guerrero Negro and Laguna Ojo de Liebre, and other sabkha areas of Baja California.

Pierre and Ortlieb at LOL. Recent and continuing research at other sabkhas in Baja California (Fig. 1) is also relevant: Laguna Mormona (Vonder Haar, 1975; Horodyski & Vonder Haar, 1975; Vonder Haar & Gorsline, 1977; Pierre & Ortlieb, 1980), Laguna Ometepe (Kinsman, 1969; Butler, 1970; Shearman, 1970; Smith, 1973; Vonder Haar & Gorsline, 1977; Pierre & Ortlieb, 1980), and Salina Grande (Vonder Haar & Gorsline, 1977; Pierre & Ortlieb, unpublished).

HISTORY (W.T.H.)

The most dramatic era of this region's history began in November, 1858, when the whaler brig Boston, under Capt. C. M. Scammon, found a way through a narrow channel into the winter calving ground of the California Gray Whale (Scammon, 1868, 1969, 1970). Capt. Scammon named it Boston Lagoon, but it soon became famous as Scammon's Lagoon, a name that still competes with the older Spanish Laguna Ojo de Liebre. "Lagoon whaling" was extremely easy and profitable, although not without its dangers in smashed whaleboats, and the following winters saw a fleet of 10 or 20 whalers in the lagoon. By 1867, however, Scammon noted that " . . . so few whales are found there that it has been abandoned as a whaling ground; the decaying carcasses and bleaching bones strewn along the shores give evidence of the havoc made by the most enterprising and energetic class of seamen that sail under our national flag." Scammon went on to state that "The salt fields of Ojo Liebre are capable of supplying an almost unlimited quantity of excellent salt. . . . A year or two after the whaling commenced, vessels were dispatched from San Francisco, Upper California, for cargoes of salt; the first two, after cruising a length of time off the desired port, returned with the account that no such lagoon existed, or if it did, no channel could be found to get into it. A third vessel [brig Advance] was sent with a master determined to either find the place or 'break something'; he lost his vessel between Black Warrior [Guerrero Negro] and Upper Lagoon [Laguna Manuela]. Subsequently, the late Captain Collins, of San Francisco, a gentleman of much experience and a skillful seaman, obtaining the most reliable information at hand, sailed for the place that seemed to baffle the efforts of his predecessors to find. In due time, he arrived at the desired haven, without difficulty procured a cargo of salt, and returned to San Francisco. These voyages were followed up for a length of time, but the low price of the article compelled the proprietors to abandon the trade" (Scammon, 1868).

The salt of Laguna Ojo de Liebre was known to the Spanish missionaries before 1757; it was on the lands of Misión San Ignacio (Barco, 1973, p. 159; Clavigero, 1937, p. 31). However, the mission needs were apparently filled since 1717 by their salt works on Isla Carmen (Fig. 1) which lay only a few kilometers offshore in the Gulf of California. Clavigero's (1937, p. 30) description of that operation, published in 1789, is of both historical and geochemical interest.

"It begins at a distance of a mile and a half from the sea, and it extends for so many miles that the end is not seen with the eyes; it appears to the observer as an immense plain covered with snow. Its salt is very white, crystallized, and pure, without a mixture of earth, or of other foreign bodies. Although it is not so hard as rock salt, it is broken with picks, and in this manner it is cut into square cakes of a size such that each laborer can carry one of them on his back. This work is carried on in the first and the last hours of the day, because at other times the reflection

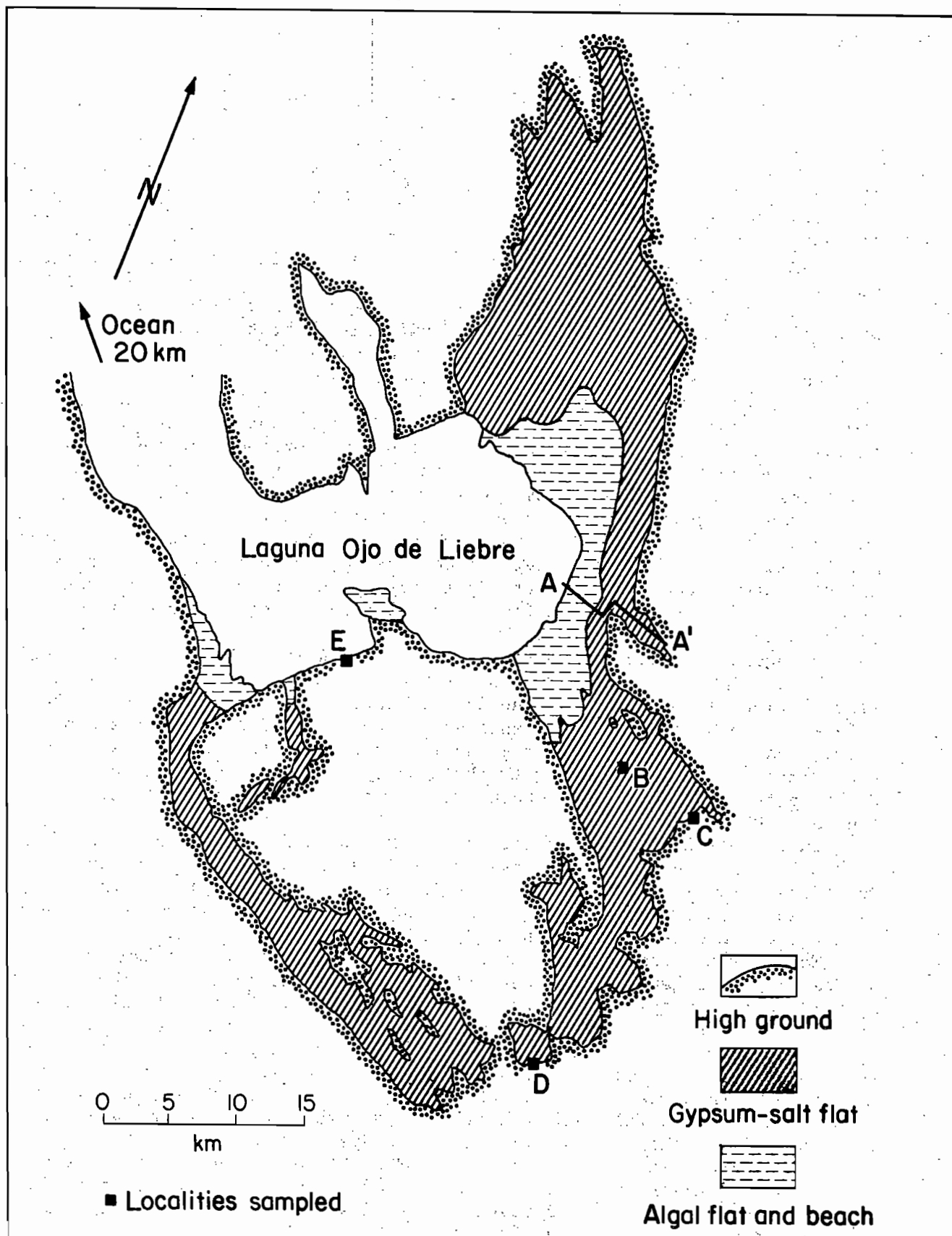


Figure 2. Map of the sabkha at Laguna Ojo de Liebre as it appeared prior to the construction of concentrating ponds, showing localities sampled both before (A, B) and after (C, D, E) pond construction, and reported in this guide.

of the rays of the sun on it is so bright that it dazzles the workmen. Although all the merchantmen of Europe might go there to load salt from the salt bed, they never could exhaust it, not only because of its great size but principally because all that salt which is taken from it is reproduced at once. When not more than seven or eight days have passed after the quantity necessary for loading a boat has been dug out, the excavation is rapidly refilled with new salt."

The Carmen Island salt deposit is a large pond nearly completely isolated by a Pleistocene coral reef, quite different from the situation at LOL: a modern description is given by Kirkland et al. (1966).

During the eighty years following the whaling period the salt of LOL was intermittently exploited. Due to the extreme isolation of this particular area, and the perennial disorganization of Federal control in this poor and distant territory of Mexico, the development could be carried on by clandestine operators. The Jefe Político of the Territorio de Baja California complained to the Agente de Fomento in Mexico City on 31 March, 1863, that Isla Carmen, Laguna Ojo de Liebre, and San Quintín were all being exploited, but were not under concession (Calderon, 1924). An expedition sent to Baja California in 1887 by the Secretaria de Fomento, found the remnants of a major industry, that had been precipitously abandoned by its operators. After being dug from the salt flats, the salt was moved by handcarts on a track to the pier, where it was loaded onto 80-ton launches and transhipped in the lagoon to ships of 400 to 500 tons (Böse & Wittich, 1912-13, p. 408). The Chinese laborers were barely sheltered in small huts; a contemporary illustration vividly portrays the scene. Throughout this period occasional reference can be found to salt shipment into the port of San Francisco (e.g., Scammon, 1970), where it probably had to compete with salt won from the brines at Redondo and from the British operation at the Liverpool Salt Works on the playa at Salton, both much closer in southern California (Guinn, 1909).

In 1911-1912, a governmental expedition to explore the resources of Baja California thoroughly surveyed the salt deposits at LOL (and San Quintín). From logs of 45 test holes in the three salinas at LOL they estimated a resource of 14×10^6 T NaCl and described the deposits and their origin (Böse & Wittich, 1913, p. 343-409; Wittich, 1916).

Thus matters rested, more or less, until April, 1954, when Exportadora de Sal, S. A., began a modern operation of solar salt recovery. The present operations are described below. Under Daniel K. Ludwig of National Bulk Carriers they made their first shipment in 1957, with a production of 7×10^4 T, mainly to the west coast of the USA and Canada, and to Japan. In 1973, the company was sold to Mitsubishi Corporation of Japan, with the Mexican government retaining 25% of the stock. In 1976, the Mexican government, through the Commission of Mineral Development, acquired the majority of stock. Production is now 5.5×10^6 T per year.

In the meantime the California Gray Whale is back, the waters of the lagoon have been set aside as a nature preserve, and the area continues to fascinate both biologists and tourists with its unique display of the family life of the whale (Gardner, 1963).

PHYSIOGRAPHY AND GEOLOGY OF THE SABKHA AT LAGUNA OJO DE LIEBRE (W.T.H.)

The Laguna Ojo de Liebre extends 50 km into the Vizcaino Desert from the shore of Bahia Sebastian Viscaïno and the Pacific Ocean (Fig. 1,2). The desert rises imperceptibly inland for another 50 km to the foothills of the Sierra de San Borjas. Sandy stream courses in this desert rarely carry runoff. Rainfall is sparse, over half of it falling in the winter months, and is highly variable from year to year (Hastings & Turner, 1965). The mean monthly maximum air temperature is 20-29°C, and the mean monthly minimum temperature is 8-20°C. West and west-northwest winds at speeds of 5 m/sec dominate the lagoon area over half the time. Although these are moist air masses off the Pacific, and dew is common at night or under an early morning overcast, evaporation is greater than precipitation by about 1.8 m/yr⁻¹ (Phleger & Ewing, 1962).

Access of ocean waters to LOL is relatively limited: the single inlet between long sand bars is 4 km wide, and 9-12 m deep, with a vertical sectional area of about 0.04 km². An old (Pleistocene: Phleger, 1965) main channel up to 12 m deep extends to within 1-2 km of the sabkha, but much of the lagoon is shallow and some is exposed at low tide. The total area of evaporation in the lagoon complex is about 500 km², although all of this is not effective in concentrating the water at the distal end of the lagoon where it feeds the sabkha. The tidal range varies from about 1 to 3 m with turbulent velocities of 1.4 m/s at the entrance channel, slowing to 0.2 to 0.8 m/s in the distal part. In spite of this intensive tidal flushing, both the temperature and salinity of the surface water are strongly increased toward the distal area. As examples, a maximum temperature of nearly 25°C at the inner lagoon in June was 4°C warmer than surface water at the lagoon entrance. In February, the temperature of 19.5°C at the inner lagoon was 1.4°C warmer than at the entrance (Phleger & Ewing, 1962). The corresponding seasonal variation of surface water in the inner lagoon is 42 to 51°; an increase of up to 45% over the normal oceanic salinity at the lagoon entrance (Phleger & Ewing, 1962; Holser, unpublished).^{*} Thus warm, dense brines are being formed at the inner lagoon surface, which feed the sabkha and also apparently intermittently reflux to the ocean. "At most times the water (in vertical section in the lagoon) is isohaline with depth at all stations. At a few stations the bottom salinity was significantly higher than that at the surface. This was most pronounced on the morning of July 19, 1956, when salinity in the inner lagoon at the surface was 46.70/oo and at the bottom 47.80/oo, in the channel (halfway to the entrance) at the surface 28.30/oo and at the bottom 38.30/oo, and in the lower lagoon main channel at the surface 35.80/oo and at the bottom 36.80/oo. These data suggest that somewhat more saline and therefore more dense bottom water was being formed in the inner lagoon on [that date] and was flowing seaward along the bottom of the channel. . . . It is conjectured that, in a lagoon, the phenomenon operates on a diurnal rather than a seasonal cycle as in the more open seas [such as the Mediterranean and the inner Gulf of California]. Data are insufficient to allow generalization

^{*}No data are given by Phleger to support his later estimate (Phleger, 1969, p. 825) of ". . . a salinity almost twice that of the adjacent open ocean."

on the frequency or magnitude of this mechanism in Laguna Ojo de Liebre. Limited observations suggest that it is infrequent, but observations were not made at night when such a process is expected to be most active" (Phleger & Ewing, 1962, p. 827). The tidal conditions for the above salinity observations were not stated. In another publication, Phleger (1969, p. 827) states, in contradiction to the above specific data "No evidence was discovered of a possible return flow of saline water out of the lagoon . . ."

The border between the upper lagoon and the sabkha (Fig. 2) is marked by a low beach ridge of sand, composed mainly of quartz, plagioclase, and hornblende, with varying amounts of carbonate shell fragments and oolites. Sands elsewhere in the lagoon contain pelletal apatite (Phleger & Ewing, 1962), but none was found at this beach. The beach ridge is breached by a number of tidal creeks, which are bordered by a carpet of Salicornia and Spartina that makes them show as black winding patches in Figure 3.

Many daily tides and especially semi-monthly spring tides flood an area behind the beach ridge, dominated by black, rubbery mats of blue-green algae which show as gray areas in the aerial photograph of Figure 3. The algal mats originally covered an area of about 100 km² (Fig. 2). The mats include interstitial aragonite, high-magnesian calcite (HMC), dolomite, and gypsum, as will be described in detail below.

Beyond the algal mat area the flats were covered with a hard crust of gypsum and halite, showing as brilliant white in Figure 3. These salt flats originally extended about 500 km². The lagoon side of this area was flooded by many tides, with evaporation sufficient to produce a visible floating film of gypsum and halite crystals. The shoreward parts of the salt flats were beyond the reach of normal tides, but apparently storm surges and persistent onshore winds were able to drive a layer of lagoonal water to the farthest reaches of the flats (Phleger, 1969). This intermittent inflow, probably one or more times a year, was documented by swash marks composed of shells, seaweed and other flotsam that wound across the flats. This kind of storm flooding was documented for Laguna Mormona, Laguna Ometepe, and Salina Grande (Fig. 1) by satellite photographs (Vonder Haar & Gorsline, 1975). The resulting deposit of salt and gypsum varies from a few millimeters thick at its lagoonward edge, to a maximum of 2.5 m. The wide-ranging survey of Böse and Wittich (1916, p. 388-398) showed salt/gypsum thicknesses generally in the range of 10 to 20 cm.* The mineralogy of the evaporite deposit includes gypsum of various morphologies; halite, polyhalite, celestite, magnesite and bassanite, as described below in detail for sections A, C, and D. The evaporite displays a succession of irregular layers 0.2-2 cm thick of light evaporite minerals alternating with darker layers of organic debris and windblown sand.

The porosity of the salts and underlying sands is permeated by a bittern brine. According to Phleger (1962) the brine is residual to evaporation of the surface flooding described above, and "The brine and

*Although Böse and Wittich (1913, p. 386) state that, "La localización de estos pozos está marcada en el bosquejo de carte que acompaña a este informe . . .," no such map of their sampling locations seems to have been published.

evaporite areas seem to be essentially a closed system retaining all the salts present in seawater. No return of brines to the lagoon could be detected, although a minor amount of seepage may occur." It is not clear what tests Phleger could have made would have precluded reflux of the brines to the lagoon. On the land side of the sabkha, although no surface runoff is apparent in the present regime, ground water in wells is fresh, both at the town of Guerrero Negro and at Ojo de Liebre ("Rabbit Well," from which the lagoon derives its name). The relation of such continental subsurface waters to the sabkha brines is unknown.

During all the early work it was clearly recognized that the hydrologic regime could only be demonstrated by a precise levelling of the water table, coupled with a tide gauge at the beach, but such measurements were never made and it is now too late to make them.

The evaporite section is underlain by sands and silts, with isolated crystals of gypsum, halite, and occasional small aggregates of polyhalite, probably crystallized in place. The "high ground" (Fig. 2) surrounding the salt flats and inaccessible to the highest storm tides, is similar in nature to that underlying the salt. In various places, old semi-consolidated alluvium, beach rock and both new and old gypsum dunes may be recognized.

The morphological origin of the depressions occupied by the sabkha salt and gypsum is obscure. According to Phleger (1969) the oldest parts of the sabkha deposit are more than 30,000 years old including one or more Pleistocene interglacials (Phleger, 1969). On the other hand, Phleger (1965) dates the offshore bars that originally enclosed these lagoons (and which must therefore have preceeded the sabkha of the inner lagoon) as accompanying the post-Pleistocene rise of sea level about 6,000-7,000 years B.P. Mina (1957) lists 48 m of Pleistocene quartz sands and 150 m of Pliocene-Miocene sands and clays at the top of the section in the well San Francisquito no. 1 nearby at Lagunitas. According to Phleger (1969) "Abundant evaporites have been found to a depth of 200 m in the cuttings from a water well near the northern edge of the Ojo de Liebre halite flats. These subsurface evaporites may be deposited by interstitial circulation of brine and precipitation of salt within the sediment. They also may represent older deposits that were precipitated when relative sea level was lower than it is today." It is clear that much remains to be learned concerning the geological and paleogeographical setting of these sabkha deposits.

One aspect of sabkha history seems clear: if the concentration of inflowing brines is sufficient to precipitate halite, as it is at LOL, the level of the sabkha always remains uniformly at a little above any relative rise of sea level. This is true not only on a geological but even on an historical scale. The old descriptions cited above of the Carmen Island deposit, and experience elsewhere, indicates that substantial excavations will be "healed" by new salt within a year. Of the undetermined but undoubtedly large tonnage of salt that was dug out of the LOL salt flats in the century from 1860 to 1960, no sign of excavation remains, even on aerial photographs. This fact also leaves some ambiguity in the interpretation of these deposits as natural geological features, including the results of drilling described below.

The mineralogy and geochemistry of the minerals and brines will be described for four separate sections of the sabkha. It is not yet possible to present an integrated picture of the geochemistry, and research is continuing by Pierre and Ortlieb.

BIOLOGY AND BIOCHEMISTRY OF THE SABKHAS (B.J.J.)

General Discussion

Temperate salt marshes around the world tend to have similar biological community structures (Larsen, 1980). Among the higher plants, the lower intertidal is dominated by the grass *Spartina* while the upper intertidal is dominated by a *Salicornia* (pickleweed) assemblage. Blue-green algae are a nearly ubiquitous part of the flora of salt marshes and intertidal shores around the world. In some places, their presence may be detected only through enrichment cultures. In other environments, color-banded beach sand ("Farbstreifensandwatt") composed of vertically zoned communities of eukaryotic algae, blue-green algae, colorless sulfur bacteria, and purple sulfur bacteria develop in loose, intertidal sediments under protected conditions. In still other environments, blue-green algae form thin to thick peats which may or may not calcify. In spite of their cosmopolitan distribution, blue-green algal communities fail to develop into stromatolites in most modern intertidal environments.

The upper intertidal limit of blue-green algal distribution is apparently related to limits of dessication. The lower limits to algal mat distribution have been attributed to current velocities, grazing pressure, and competition from higher algae and plants. Little is known about why eukaryotic algae generally do not inhabit stromatolite environments. In some mats, eukaryotic and prokaryotic algae coexist.

Intramats and submat reducing conditions, and the presence of H_2S in marine mats are common. Often a rich photosynthetic bacterial flora develops as a one-to-several-mm-thick pink lamination directly below the top layer of blue-green algae.

Preservation of the organic component of laminated algal sediments is promoted by anaerobic conditions beneath the mat surfaces. It has also been suggested that low oxidation potentials in the water may inhibit decomposition. When anaerobic conditions remain undisturbed in some Persian Gulf sediments the cellular structures have been preserved as long as 8000 years.

Most environments where $CaCO_3$ precipitation is occurring within modern marine stromatolites are moderately to highly hypersaline. The carbonate system in moderately hypersaline seawater is virtually unstudied. Predictions of solution and precipitation processes of $CaCO_3$ must be based on knowledge of activity coefficients, solubility products, and ionic interactions that are empirically derived. In these particulars the geochemistry of environments inhabited by stromatolites, and the geochemistry of their morphogenesis, are very poorly understood.

Algal Communities at Lagunas Guerrero Negro and Ojo de Liebre

Intertidal algal mats develop in some of the slightly hypersaline marshes and in many of the moderately hypersaline flats. General geological and climatological aspects, as well as the general biology of the blue-green algal communities of these lagoons, are very similar to those environments described on the Abu Dhabi coast of the Persian Gulf (Cardoso et al., 1978), the Laguna Madre area of the Texas Gulf coast, and the Laguna Mormona area of Baja California (Fig. 1; Horodyski & Vonder Haar, 1975).

The major mat types each correspond to the lower, middle and upper intertidal. In the lower intertidal mat, the top photosynthetic layer is actually a bilayer composed of a surface blue-green algal horizon dominated by Microcoleus chthonoplastes (about 3 mm thick), and an underlying purple layer of photosynthetic bacteria predominantly Chromatium sp. (about 2 mm thick). The mat may accrete to a thickness of 10 cm or greater, and is characterized by fine laminations of alternating blue-green algal and photosynthetic bacterial layers. The middle intertidal is characterized by a relatively thin (up to several cm) mat dominated by Lyngbya aestuarii, and only a thin, lower horizon of purple photosynthetic bacteria. One-cm-tall pinnacles that resemble the Precambrian stromatolite, "Conophyton," occur locally in the Lyngbya mats. The upper intertidal is characterized by a crusty, wrinkled mat dominated by Calothrix crustacea. Total mat thickness is about 0.5 cm. None of the mats is lithified.

Experiments on the photosynthetic potential of algal mats in the field gave the following results: temperature optima between approximately 35 and 40°C; all mats were inhibited photosynthetically under full sunlight and they demonstrated maximal photosynthetic rates at about 300-600 W/m² (full mid-day sunlight is about 1000-1200 W/m²); salinity optima relative to normal seawater were Microcoleus mat = 0.4-1.5X, Lyngbya mat = 1.2X, and Calothrix mat = 1.5X; and none of the mats was apparently inhibited by high concentrations of O₂.

Two grazing invertebrates, a gastropod (Cerithidea californica) and a waterboatman (insect: Trichocorixa sp.), are dominant biological factors that limit mat distribution. The mats in effect find refuge from the very destructive cerithids by their ability to grow in environments that exclude the gastropods. Cerithid populations are promoted by: proximity to a fresh source of seawater, and soft and moist subaerial sediments. Waterboatmen are nearly ubiquitous in these environments but they never completely decimate mats. They are restricted by elevated salinities and short lifetimes as adults. Because the Microcoleus mat is the fastest growing, and because many of the sediments are bare in the lower intertidal range of the lagoon, invertebrate grazers apparently restrict Microcoleus mat development on all lower intertidal sediments except those in isolated pools.

Aragonite precipitation occurs in the mats. In Laguna Guerrero Negro, non-mat sediments are silicoclastic sands, so the precipitation of CaCO₃ in the mats could be measured. Absolute CaCO₃ content increased with total mat thickness and organic carbon content. CaCO₃ content decreased with depth in the mats, demonstrating that in a predominantly silicoclastic, slightly hypersaline environment, the CaCO₃ produced in the mats is not stable. Organic geochemistry of the mats has been described recently by Philp et al. (1978).

Bacteria in the Concentrating Ponds

It is well known among solar salt company managers that the pink coloration of brines increases the solar absorption of the brines (heat) and thus enhances evaporation and precipitation. Temperatures of colored brines are almost invariably higher than those of clear brines in the same saltern at any particular time of day. The cause of the coloration is the great proliferation of brine bacteria (halobacteria).

Under conditions of increasing brine concentration of marine seawater, such as is found in commercial salterns, is there an upper brine concentration that limits halobacterial distribution? In the E.S.S.A. saltern, pigmented bacteria are not found in the bitters. Brines are very pink until a density of about 1.26 and they are brownish grey at a density of 1.27. At higher densities, brines are green, either clear or opaque (due to suspended salt crystals), which apparently is the natural color of the brines without pigmented bacteria. Brines up through the bitters stage were recently plated and are now being identified. It is unknown at this time whether viable bacteria that might be found in the bitters are reproducing in that environment.

SALT PRODUCTION AT GUERRERO NEGRO (E.S.S.A.)

Common salt, or sodium chloride, is a very abundant mineral, and is widely used in the chemical and food industries as well as for de-icing roads. Exportadora de Sal, S.A., operates the largest solar salina in the world, and exports all of its production of about 5.5 million tons per year to Japan, the United States and Canada.

The concession authorized to the company by the Mexican government is 39,995 hectares. This consists of two main areas: 21,500 hectares for concentration subdivided into 13 areas in order to control the evaporation process; and 3700 hectares for recrystallization subdivided into 49 areas, named vasos. These ponds are divided by means of 175 km of dikes, of which 111 km in the concentration area is protected to prevent erosion by wind and wave action. Within the system there are another 30 km of canals and 30 km of major transport roads.

There is also a salt washing plant and port facilities for loading barges in an area called Chaparrito. In Guerrero Negro there are shop facilities for maintenance, administrative offices, houses for company personnel, a hospital, church, primary and secondary schools, and recreational facilities. The population of Guerrero Negro is 5400, of which 580 work for E.S.S.A.

Drinking water is taken from six wells, approximately 20 meters deep, and pumped to town and company facilities by means of 10 km of pipelines. Energy is provided by a generator of 1500 kW, two of 730 kW, and two of 800 kW. There are 20 km of transmission lines for the distribution of electricity.

The seawater of Laguna Ojo de Liebre, with a density of 1.0313 and NaCl content of 3.5%, is pumped into the first concentration area by 10 pumps of 1.06 meters in diameter, with a total capacity of 1190 m³/min. This water passes through the concentrating area, and by means of solar evaporation becomes saturated with NaCl. It is transferred to the crystallization area at a density of 1.221 and 22.5% NaCl.

In the crystallization area, the brine deposits 80% of its contained salt. Upon reaching a density of 1.255, the brine is pumped out. This is why the company is developing studies for the recovery of by-products from these bitters.

Every crystallizing vaso is harvested twice a year. The thickness of the layer of harvested salt is approximately 12 cm. The vasos are first completely drained by pumps, scraped and piled in rows by snowplow-like machines, then collected by harvesters. The salt is lifted in the

harvester by means of an elevator belt consisting of salt scoops and a rubber belt. The harvesters were specially designed for the company. Each one has a capacity of 2000 tons per hour.

The salt is transported about 12 km to the washing plant by means of five Kenworth Dart trucks, each one with three trailers. The capacity of each trailer is 120 tons.

The Kenworth Dart trucks unload the salt in the hopper at the washing plant. The hopper feeds eight stainless steel belts where the salt is first rinsed with brine of density 1.174, and then with seawater. This removes and dissolves the superficial impurities from the crystals (insoluble material and salts of calcium and magnesium), increasing the purity from 98.5 to 99.5% NaCl.

After washing, the salt is sent by conveyor belt to a stockpile with capacity of 100,000 tons. Alternatively, it is directly loaded onto barges by means of a system of conveyor belts with a capacity of 1800 tons per hour.

The salt is transported 100 km (moved by five tug boats with a power of 2500 to 3500 H.P.) from Guerrero Negro to Cedros Island. Cedros Island has three unloading docks with a capacity of 1800 tons per hour. The barges are self-unloading, having a conveyor belt below. The capacity of the stockpile of salt on Cedros Island is 900,000 tons.

The dock in Cedros Island has a length of 221 meters and a depth at low tide of 18 meters. It can load ships up to 156,000 tons capacity. The salt is taken from the stockpile at Cedros by means of a machine made by Barber Green, which has a capacity of up to 3000 tons per hour. The system for loading ships consists of conveyor belts and a telescopic hopper.

MINERALOGY AND GEOCHEMISTRY OF SABKHA SEDIMENTS AND BRINES: SECTION A (W.T.H.)

Sedimentary Deposits

Section A (Figs 2, 3, 4) extends from the distal shore of LOL at the abandoned salt-loading tramway known as the "English Bridge" in a nearly easterly direction across the algal flats, then jogs to the north before continuing up the center of a prominent embayment in the salt flat ("Salina Chica" of Böse and Wittich, 1913; currently called "Huizache"). Before construction of the concentrating ponds this section was the most accessible, and was sampled several times. In the central part of this section four holes were cored and sampled down to bedrock (Fig. 5).

Beyond a low beach ridge of detrital sand, black mats of algae cover the first 2 km of the section, and control the carbonate mineralogy that is mixed with the clastic sands. As evaporative concentration increases eastward, evaporite minerals begin to accumulate, although with a continuing component of clastic material as well as layers of decomposing salt grass and other organic flotsam (Figs. 6, 7). In the embayment evaporite minerals clearly dominate the upper layers. The drilled sections (Fig. 5) show several meters of sands and silts underlying the whole sabkha with variable amounts of dispersed diagenetic evaporite minerals.

Although lines in Figure 5 suggest some correlation of sedimentary beds, these are probably not very significant--the changes of sediment type are rather discontinuous, as are the individual mm-to cm-layers that are evident in much of the sediment (Figs. 6, 7).



Figure 3. Aerial view of a cross section (A in Fig. 2) of the sabkha before construction of concentrating ponds. The distal end of the lagoon, and the beach, are in the foreground, followed by the gray intertidal algal flats cut by black tidal creeks. Gypsum-salt flats are stark white, reaching their greatest thickness and degree of concentration in the embayments beyond. The sampled section A follows the causeway of the "English Bridge," which is seen as a straight line leading from the lagoon in the lower right. Reproduced from Holser (1966), with permission of The American Mineralogist.

Mineralogy

Qualitative X-ray diffraction, microscopy, and field observations (by D. J. J. Kinsman and W. T. Holser) show an interesting pattern of mineralogy, as laid out in Table 1.

Sediments under the algal mats include much carbonate mud. High-magnesian calcite (HMC) of 2-3 mm size dominates; X-ray determinations (± 1 -2 mole percent MgCO_3 , using an Si standard) show from 10 to 14 mole percent MgCO_3 . The HMC is consistently associated with debris of Zostera encrusted with foraminifera and red algae. Dolomite is also common, presumably replacing HMC: it contains 44-47 percent MgCO_3 and all of it shows some degree of ordering.

During the 1965 field work D. J. J. Kinsman discovered magnesite at LOL, which he briefly reported previously (Kinsman, 1969), and the following description is supplied by him (personal communication, 1966). It was found (sample 24822-1) in the salt flats (Fig. 4) in a porous layer of quartz-feldspar sand, at 7 cm depth, as 3-mm, hard, dead-white blebs, associated with 3-mm discoidal gypsum crystals and rare celestite needles. The magnesite aggregates were composed of 1.5- μm equant grains. X-ray diffraction confirmed magnesite, with $a = 4.658$, $c = 15.134\text{\AA}$, compared to a standard (XRPDF starred pattern 8-479) $a = 4.6332$, $c = 15.015\text{\AA}$. The expanded lattice could be accounted for by a substitution of 7 mole percent CaCO_3 , and this is confirmed by atomic absorption analysis of a carefully cleaned sample: Ca^{2+} 36,150 $\pm$ 250 ppm, Sr^{2+} 2.8 $\pm$ 1.1 ppm. An expanded lattice is characteristic of Recent magnesites including those of the Trucial Coast on the Persian Gulf (Kinsman), South Australia (Alderman & Von der Borch, 1961), and Great Salt Lake (Graf et al., 1961), although in the latter case the total foreign cations failed to account to the expanded lattice. The HMC and dolomite found in the algal mat area are absent in the salt flats--and even magnesite is only sporadically present there. Huntite, $\text{Mg}_3\text{Ca}(\text{CO}_3)_4$, previously found in the Trucial Coast sabkha (Kinsman, 1967), was looked for but not found at LOL.

Gypsum is ubiquitous at LOL, most evident as layers of coarse crystal mush (Fig. 6, 7), but also common as aggregates and as larger poikilitic crystals in the underlying sands. A discoidal morphology is common, but in fact, a wide variety of growth forms and features are found at LOL and these have not been systematically studied.

Six gypsum samples from core holes on the salt flat (Figs. 4, 5) were analyzed by optical emission spectrography by R. E. Wolven of Chevron Oil Field Research Company. Sr contents were 18136-2 3600, 18136-5 5800, 18136-9 1600, 18136-14 250, 18136-18 380, and 18137-12 1700 ppm. The first two samples are aggregates, in which the presence of celestite cannot be precluded. The purest single gypsum crystals are 18136-9 and 18137-12 (as verified by SiO_2 levels of 0.07 and <0.04 percent), and their strontium levels are consistent with those reported by Butler (1973) in his detailed study of strontium in the Trucial Coast. Other trace elements in these two pure samples are B <3 , <3 ; Ba 3.5, 5; Fe <30 , <30 ; Mg 400, <100 ; and Mn <3 , <3 ppm. It may be significant that the two other samples from hole 18136, which occur as 1- to 2-cm crystals poikilitically enclosed sand grains in sands several meters below the evaporite layers, are an order of magnitude depleted in Sr.

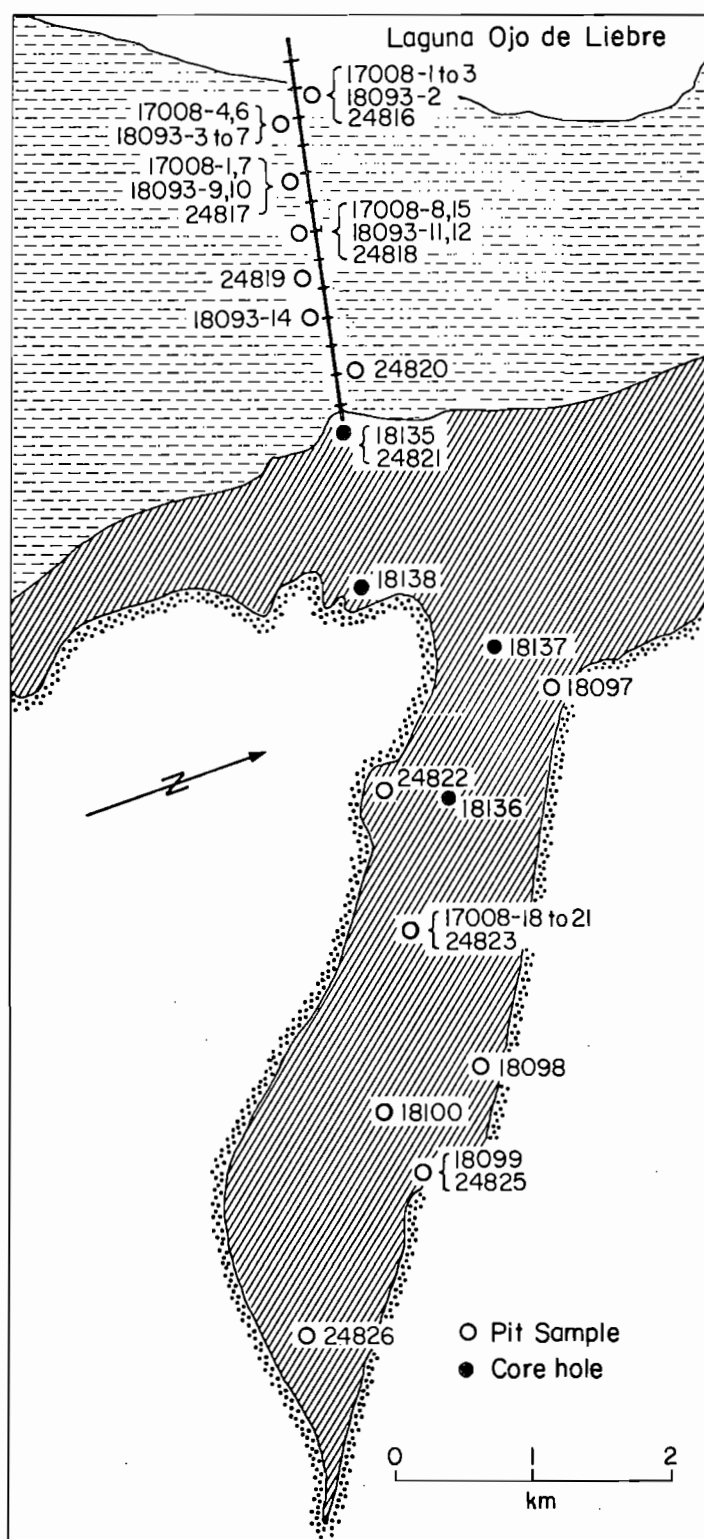


Figure 4. Map of Section A of Figure 2, showing sampling of May, 1960 (17008), August, 1961 (18093-18138), and May, 1965 (24815-24827), all prior to pond construction in this area.

Table 1. Qualitative Mineralogy of Near-Surface Sediments Along Section A, west (lagoon beach) to east (high ground) (Figs. 2-5)

	24816	18093-3	24817	24818	18093-14	24820	See Fig. 5				18097	24822	24823	18098	18100	24825	24826
Quartz	D	D	M	T	T	T	M	-	-	T	T	D	-	-	-	D	D
Plagioclase	M	T	-	T	M	-	-	-	M	T	-	D	-	-	-	D	D
Mg-calcite	D	D	D	D	M	M	T	-	T	T	-	-	-	-	-	-	-
Dolomite	M	-	-	M	-	-	M	-	T	-	-	-	-	-	-	-	-
Gypsum	T	M	M	M	M	M	D	D	D	D	D	D	-	-	M	M	-
Halite	-	-	T	T	T	T	T	T	D	D	D	D	D	D	D	D	D
Polyhalite	-	-	-	-	-	-	-	-	-	M	-	-	M	M	M	M	D
Magnesite	-	-	-	-	-	-	-	-	-	-	-	D	-	-	-	-	T
Celestite	-	-	-	-	-	-	-	-	-	-	-	M	-	-	-	M	-

X-ray diffraction analyses by D. J. J. Kinsman and W. T. Holser, supplemented by field observations.

D = dominant; M = moderate; T = trace; - = not observed (absent, or in some cases not looked for).

At one locality (24825, Fig. 4) a euhedral crystal of gypsum enclosed a core of white bassanite, $\text{CaSO}_4 \cdot 1/2\text{H}_2\text{O}$. Surprisingly, a thorough search both in the field, and in laboratory examination of samples, failed to detect any trace of anhydrite at LOL. Celestite was detected in two samples on the salt flat (Table 1) as 10- μm -long needles; probably it is more widely distributed but unrecognized.

Halite is ubiquitous in the salt flat area. It can be observed crystallizing with gypsum in sheets of tidal water one to a few centimeters deep, both as a floating film of crystals and as a crust on the bottom. Larger crystals, up to 2 cm, occur both in the layered evaporite body and sporadically in the underlying sands. Cube morphology is common, but the incidence of hopper crystals, chevron growths, and other morphological variations has not been systematically studied at LOL. Deposition of gypsum and halite grossly overlap, with no clear horizontal or vertical zonation.

Bromide is an important trace element in halite (for a survey of bromide geochemistry of salt rocks see, for example, Holser, 1979b). Table 2 lists some bromide analyses of salts from Section A--the multiple values are separate analyses of subsamples, mostly distinct crystals of halite from the same general part of the core. Values in the expected range of primary initial crystallization from a marine brine are found both at the western end of the salt flat, and far into the salt flat. But in the thickest salt (hole 18136) the values are about twice, mean 156 ppm Br/NaCl at all levels in the core, with a moderate variation of $2\sigma = \pm 46$ ppm. This indicates that the brines from which these salts crystallized were concentrated in the range 20-40X sea water. This concentration agrees with the analyses of the brine presently permeating the salt flat, as shown below (Table 3).

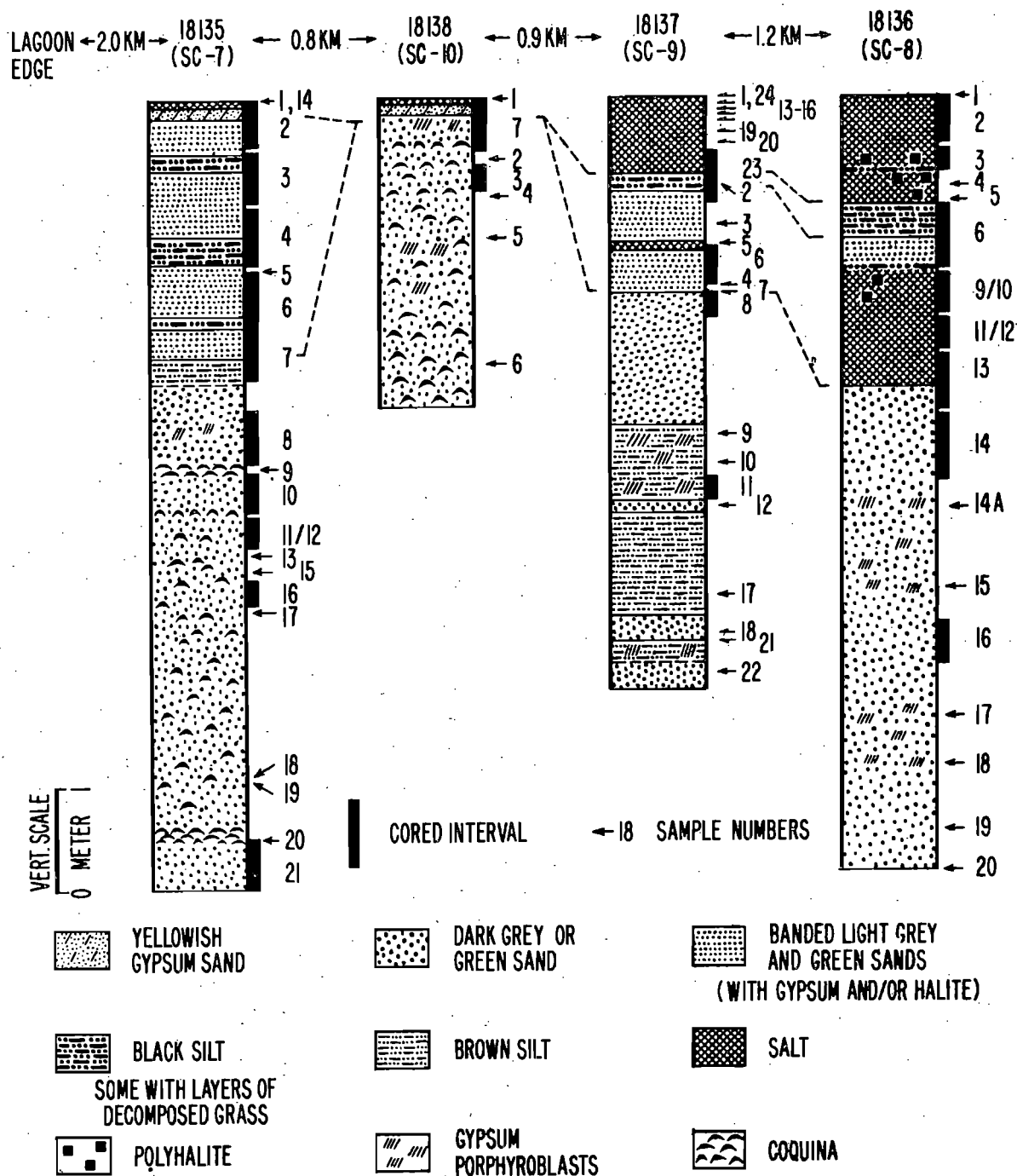


Figure 5. Field logs of the four core holes in the mid-part of Section A (Fig. 4).

Table 2. Bromide Analyses of Halite Along Section A

<u>Sample no.^a</u>	<u>Br/NaCl, ppm^b</u>
17008-9	67
18137-13	72
-16	94
18136-2	121, 153, 122, 179, 171
-3	195, 137, 186 123, 100
-4	129
-5	120
-9	168
-11	106, 149, 269
-13	208
-16	82, 123, 157, 68
17008-18	92
-19	51

^bAnalyses by Gale Baker, Chevron Oil Field Research Co., chloramine-T method.
^aFor positions of samples see Figures 3, 5.

But it is still not clear, as indicated earlier (Holser, 1966) on the basis of very limited bromide data, why halites on the more distal salt flat (17008-18, 19), where even polyhalite is precipitated, have again the lower levels of bromide.

Polyhalite ($K_2MgCa_2(SO_4)_4 \cdot 2H_2O$) at LOL was the first discovery of the mineral in Recent sediments, and its occurrence was detailed in a previous publication (Holser, 1966). Curiously, since that description, it does not seem to have been confirmed from other Recent sabkhas. It was found at many places in the central and distal parts of the salt flats (Table 1), in sections A (Fig. 3), B (Fig. 2) and C (Fig. 2; see below). It occurs as a dead-white caliche-like mass just under the water table (Fig. 6), as well as dispersed aggregates in the underlying sands (Fig. 5). Much of the polyhalite replaces previous gypsum euhedra as a mass of 5- μ m fibers, in many cases roughly oriented perpendicular to the crystal face of the gypsum (Fig. 8). Some polyhalite also appears to be in an interstitial precipitate, but the presence of a fine-grained gypsum precursor cannot be excluded. In most, but not all, polyhalite samples some remaining gypsum is still apparent. Polyhalite from other sections of the sabkha is described by Pierre (1981), and in this guide.

Brines Chemistry

Chemical analyses of a sequence of brine samples are shown in Table 3, arranged geographically from the lagoon on the left, across the algal flats, and through the salt flats. This chemical data should be considered as



(a)



(b)

Figure 6. Two sampling pits on the sabkha of Section A. (a) Near the border between algal flat and gypsum-salt flat (approximately 24820 in Fig. 4). The white layers are gypsum and the black layers are accumulation of organic material, including both salt grass and algal debris. Note the irregularity and discontinuity of the layering. Scale is 15 cm long. Reproduced from Holser (1979a), with permission of the Mineralogical Society of America. (b) On the central salt flat (approximately 24823 in Fig. 4). White layers are mainly halite and gypsum, gray layers still contain some organic material (grass, etc.), and the dead white layer near the bottom of the pit is polyhalite. Scale in inches. Reproduced from Holser (1966), with permission of the American Mineralogist.

Table 3. Chemical and Isotopic Analyses of Brines, mg/l^a

	Standard Seawater	Beach 18093-2	Algal flat 18093-9	Algal flat 18093-11	Edge salt 18135-1	Gypsum-salt flat			
						18138-1	18137-1	18136-1	18096-1
Na ⁺	11,100.	15,100.	42,200.	86,200.	70,500.	86,200.	68,000.	52,000.	29,400.
K ⁺	390.	607.	1,670.	4,500.	7,500.	5,700.	10,600.	11,500.	9,000.
Mg ²⁺	1,330.	1,960.	5,380.	30,000.	33,500.	21,400.	32,800.	52,500.	66,000.
Ca ²⁺	420.	943.	2,110.	370.	218.	126.	248.	130.	95.
Sr ²⁺	8.0	9.4	21.	8.0	8.0	6.4	4.	3.	<2.
Cl ⁻	19,800.	26,600.	82,000.	187,000.	189,000.	190,000.	191,000.	190,000.	200,000.
Br ⁻	66.	100.	270.	1,000.	1,400.	1,110.	1,650.	2,050.	3,200.
SO ₄ ²⁻	2,760.	5,600.	11,000.	31,500.	48,400.	30,700.	46,700.	59,500.	71,900.
δD, ‰	0.	+1.5	+1.6	+0.3	-2.0	-2.1	-2.1	-2.6	-3.0
δ ³⁴ S, ‰	+20.	+18.6			+18.8				+18.6
Density	1.025		1.095		1.242	1.212		1.262	1.265
Mean concentration ^b	1.0	1.47 ±0.05	4.24 ±0.04	19±4	24±2	16±1	26±1	36±5	50±1
Water table, cm	-	0	?	+3.	-1.	-30.	-6.	-1.	0.

^aAll from Section A of Figures 2, 3, 4, 5, except 18096-1 which is from Section B of Figure 2.

Analyses by conventional or atomic absorption methods by Gale Baker and Vina Spiehler, except for Sr by standard addition to optical emission spectrography by R. K. Wolven; δD by S. R. Silverman, all Chevron Oilfield Research Co. δ³⁴S by I. R. Kaplan (Holser & Kaplan, 1966).

^bMean concentration relative to seawater calculated by ratios of probably conservative elements.

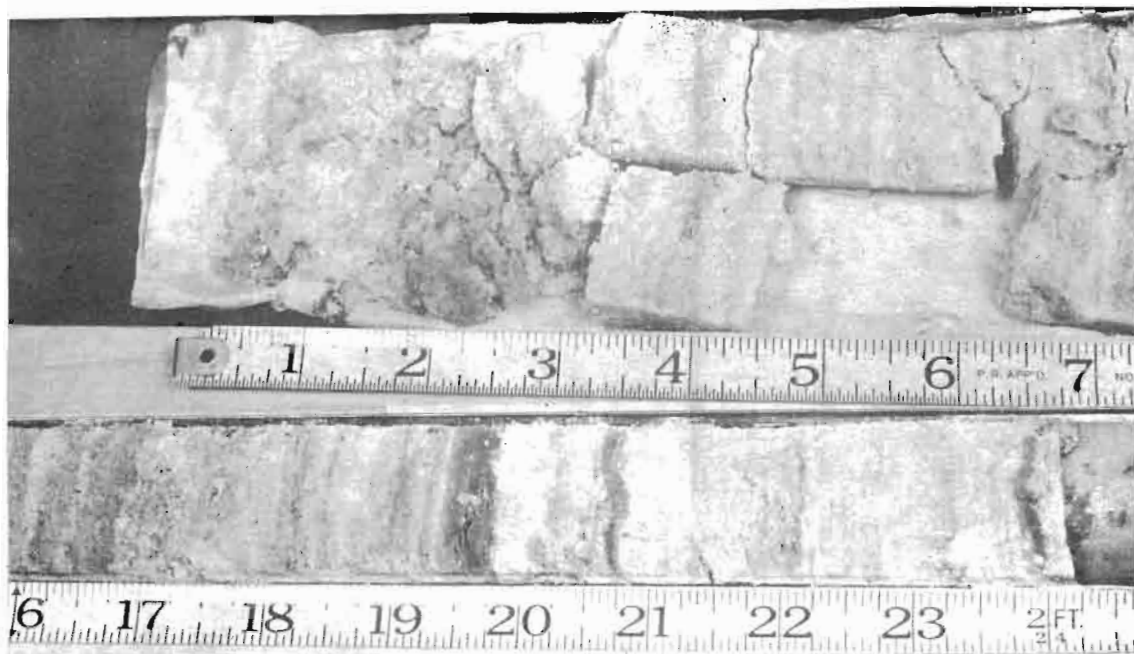


Figure 7. Two parts of the core from hole 18135 of Section A (Figs. 4, 5). Mixed halite and gypsum at top, with decreasing salt and increasing layering of organic debris downward. Scale in inches below surface. Reproduced from Holser (1979a), with permission of the Mineralogical Society of America.

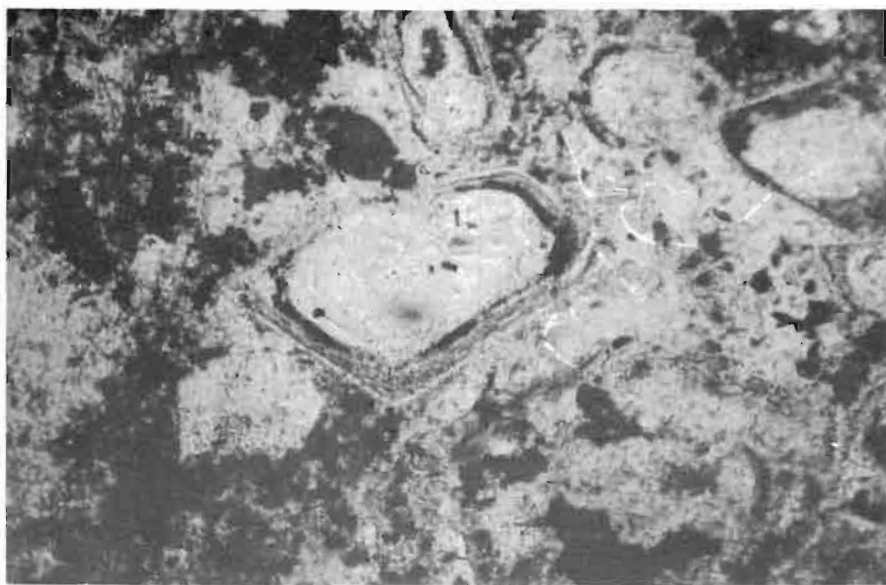


Figure 8. Photomicrograph (X40, plane-polarized light) of a rim of polyhalite replacing a euhedral gypsum crystal; most of the interstitial material is also polyhalite. Reproduced from Holser (1976), with permission of The American Mineralogist.

indicative rather than definitive, in view of the following recognized deficiencies: (1) Analyses of bicarbonate, alkalinity, pH, and Eh were not made; (2) Even allowing for a nominal amount of bicarbonate, the ionic balances are still off by $\pm 2\%$; (3) The sampling is not isochronous, and the relevant data on tide, brine flow, temperature, and humidity are not recorded; and (4) Samples are mainly from brine in the more permeable of the layers just below the water table, but with an unknown contribution from deeper layers. The analyses given here are in some cases corrections of ones published previously (Holser, 1966; Holser & Kaplan, 1966).

In general, the brine analyses represent a series of increasing concentration, as modified by the crystallization and alteration of carbonates, gypsum, and halite; although the sequence does not exactly correspond to the geographical arrangement of Table 3. The analyses show the same general development as in the synchronous sequence from an extensive set of artificial brine pans analyzed in beautiful detail by Herrmann et al. (1973). It is significant that in both LOL and Herrmann's work, sulfate ion is concentrated and remains high through the highest evaporation stage. Taken with the sulfur isotope data (see below) this suggests that sulfate reduction is not quantitatively significant in either locality, despite the obvious reducing conditions under the algal mats at LOL and the negative Eh and zero O_2 content measured in the most concentrated pans by Herrmann et al. (1973). Consequently, the calcium and strontium levels in the brines (Sr was not measured by Herrmann et al.) are forced down to very low levels by precipitation of gypsum and celestite. At the same time, the rise of K and Mg forces the replacement of earlier gypsum by polyhalite. Chloride levels off when halite starts to precipitate, but bromide continues to rise (although at a slower rate) by virtue of its rejection from halite. But the most concentrated brine is not at a level sufficient to expect minerals of the potash-magnesia facies (e.g., Holser, 1979a).

Isotope Geochemistry

Sulfur isotope geochemistry of brines and sulfate minerals along section A was reported in detail by Holser and Kaplan (1966), including specific sample descriptions. In general, the brines are remarkably uniform in composition at $\delta^{34}S = +18.7 \pm 0.1\%$. Any quantitatively important bacterial reduction of sulfate to sulfide, in the brines, would have pushed the $\delta^{34}S$ upward from the normal sea water value of $+20\%$. The levels determined, including water from the lagoon itself, are about 1% lighter than normal for present sea water. Although specific data on $\delta^{34}S$ in sulfate of continental ground water of this area is not available, continental waters are usually lighter in $\delta^{34}S$ than sea water. The low $\delta^{34}S$ could possibly indicate a minor admixture of such continental sulfate in the brines and lagoon. The sulfate minerals from a variety of occurrences in the evaporite deposit (10 analyses) have $\delta^{34}S = +21.0 \pm 0.7\%$. The levels are $1-2\%$ heavier than sea water or the lagoon water, as expected from a slight fractionation during crystallization (for a critical review see Claypool et al., 1980). Polyhalite is not significantly different from gypsum.

Deuterium levels were measured in some brine samples--unfortunately, it was not possible to combine these analyses in the usual way with measurements of oxygen isotopes. The results (previously summarized in Holser, 1979b) are given in Table 3. Evaporation of the brine initially causes a

rise in δD , but at higher salinities a sharp reversal towards lighter δD occurs, in accord with the experimental and theoretical results of Gonfiantini (1965) and Gat (1980). The decrease in δD landward could either be due to this type of evaporative reversal, or alternatively it could be due to dilution by customarily light continental ground waters. It is possibly significant that the sequence of δD is ordered more closely with the geographic sequence of Table 3 (see Fig. 4) than it is with the various determinations of the degree of concentration of salts expressed in the data of Table 3. Perhaps this indicates that the waters of the brines have been subjected to a uniform increase of evaporative stage from west to east (or uniform mixing with continental waters), whereas the chemistry of their dissolved constituents may have been modified by re-solution or alteration of previously deposited minerals.

Discussion

The expected characteristics of the LOL sabkha, and in particular of Section A, are:

(1) A general increase from shore towards the land of degree of evaporation, of density, and of concentration of conservative chemical elements in the brines permeating the sabkha.

(2) A consequent shift from carbonate mineralogy to higher evaporite mineralogy.

(3) The alteration of HMC and probably aragonite to dolomite.

The unusual or unexpected aspects of this sabkha are:

(4) The widespread and permanent deposition of halite, in contrast to only minor or evanescent (or absent) halite in many other Recent sabkhas.

(5) The failure of gypsum and halite to separate as distinct depositional facies.

(6) The high concentration of the brines, and the consequent alteration of gypsum to polyhalite, and deposition of (alteration of HMC to?) magnesite.

(7) The absence of alteration of gypsum to anhydrite (or of direct precipitation anhydrite), in spite of the low activity of H_2O in the brines of high concentration and temperature.

(8) The absence of substantial sulfate reduction. While this may be unexpected in the face of common ad-hoc models of sabkha chemistry, it is in complete accord with the fact that most gypsum rocks in the sedimentary section, including most ancient sabkhas, have sulfur isotope ratios that correlate synoptically around the world. Such sulfates are therefore presumed to be samples of sea water and not of sulfates variably increased in $\delta^{34}S$ by local basinal sulfate reduction (e.g., Claypool et al., 1980).

MINERALOGY AND GEOCHEMISTRY OF SABKA SEDIMENTS AND BRINES:
SECTIONS E₁ & E₂ (C.P., L.O.)

Sedimentary Deposits

(See Figs. 9, 10 and 11)

Along the southern coast of Ojo de Liebre lagoon, extensive salt-marshes cut by numerous tidal-channels, are backed by supratidal flats, covered by a thin algal-mat. This algal-mat shows polygonal structures and progressively disappears landwards. The surface is strewn with dried algae (*Zostera marina*) and marine shells (mainly *Melampus olivaceus*) which are carried by the wind and the highest spring tides. Sections E₁ and E₂ were studied at, respectively, 10 and 11 Km from the pump station, across the supratidal flats.

Carbonate sediments have a maximal thickness of 45 cm (Fig. 10,11); they are located in the external part of the flats below the algal-mat. Two macro-facies may be distinguished: the upper layer with a yellowish colour is pelloid, while the lower light grey layer is finely grained. Drift shells and plant remains are observed in these deposits.

Carbonate sediments are underlain by gypsum in section E₁ (Fig.11) or by sand in Section E₂ (Fig. 10); landwards, they laterally grade to gypsum deposits. The gypsum crystals have a lenticular habit which indicates a crystallization by diagenetic emplacement in the host sediments (Shearman, 1971). They frequently present dissolution features and, just below the surface, nodular recrystallizations. At the landward edge of the flat, gypsum crystals are accumulated by winds and form gypsum dunes.

Mineralogy of Carbonates

The carbonate fraction of these deposits consists of a complex assemblage of rhomboedral carbonates, sometimes with aragonite (Pierre and Person, in press). This aragonite, detected by X-ray analysis, is mainly related to the presence of shell fragments, but may also be represented by needles (observed in smear slides) of primary precipitation. The chemical composition of rhomboedral carbonates was determined by X-ray diffraction measurements, using the idealized curve of Goldsmith and Graft (1958). In some samples this composition may be difficult to determine, particularly when the diffraction picks are not well separated.

The main carbonate phase corresponds to magnesian calcites with 14 to 16 moles % MgCO₃. It may be associated with a low-magnesian calcite (6 moles% MgCO₃) and with a high-magnesian calcite (up to 25 moles% MgCO₃) (Table 5).

Other magnesium-rich rhomboedral carbonates are associated with magnesian calcites; they preferentially occur in sediments below and just above the water table. Their chemical compositions range from (Ca₇₀Mg₃₀) to (Ca₅₅Mg₄₅); therefore they may be considered as calcian dolomites.

Interstitial solutions: physico-chemical data

(See Table 4)

Interstitial solutions show increasing densities in a landward direction ($1,047 < d < 1,092$); these densities correspond to dissolved salt contents ranging from about 80 to 180 g/Kg and furthermore indicate that solutions are probably under-saturated relative to gypsum, since gypsum saturation of a marine brine is reached for densities close to 1,110. This interpretation is in agreement with dissolution features observed in gypsum crystals.

Temperatures of the interstitial solutions are quite constant ($t = 20 \pm 1,5^\circ\text{C}$) all along the sections.

E

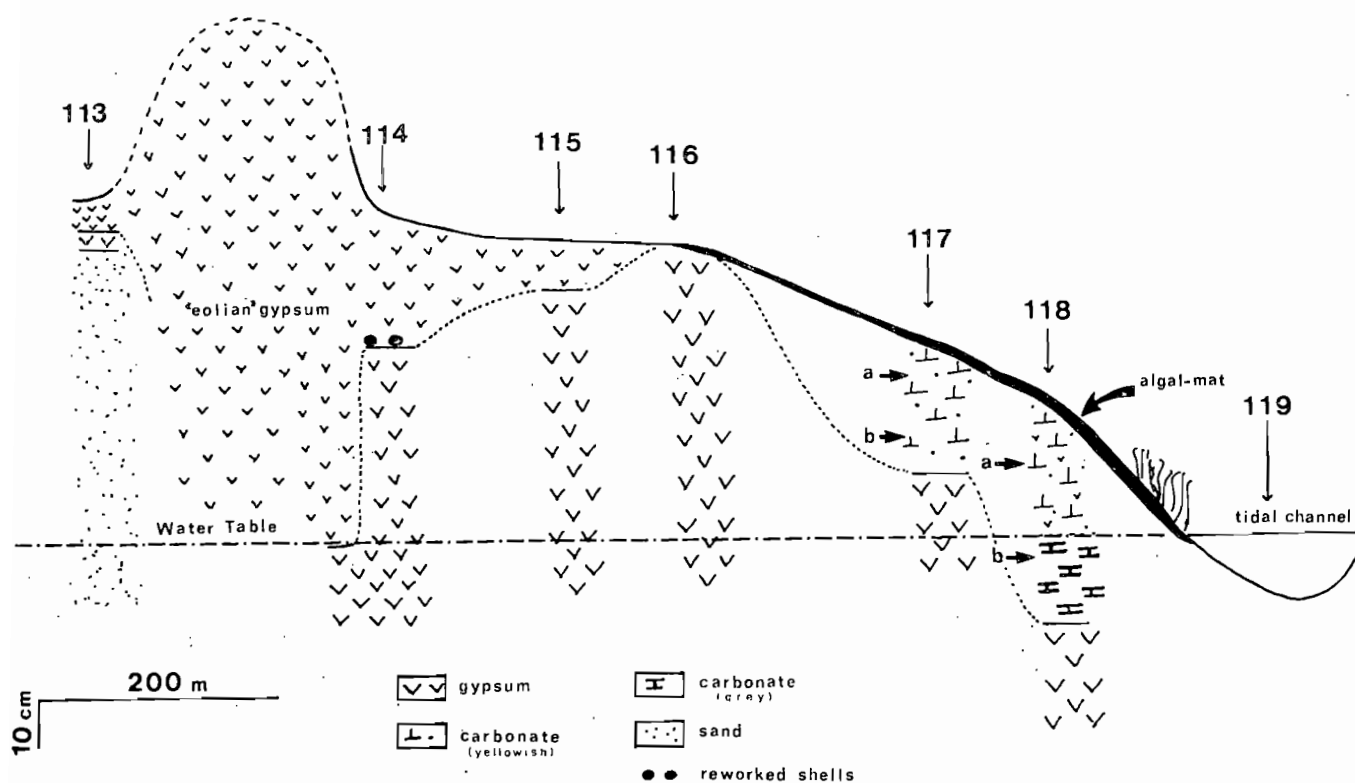


Figure 10. CROSS-SECTION E₁ OF THE SUPRATIDAL FLAT ON THE SOUTHERN SHORE OF OJO DE LIEBRE LAGOON: SAMPLES 113 TO 119 (SEE FIG. 9 FOR LOCALIZATION).

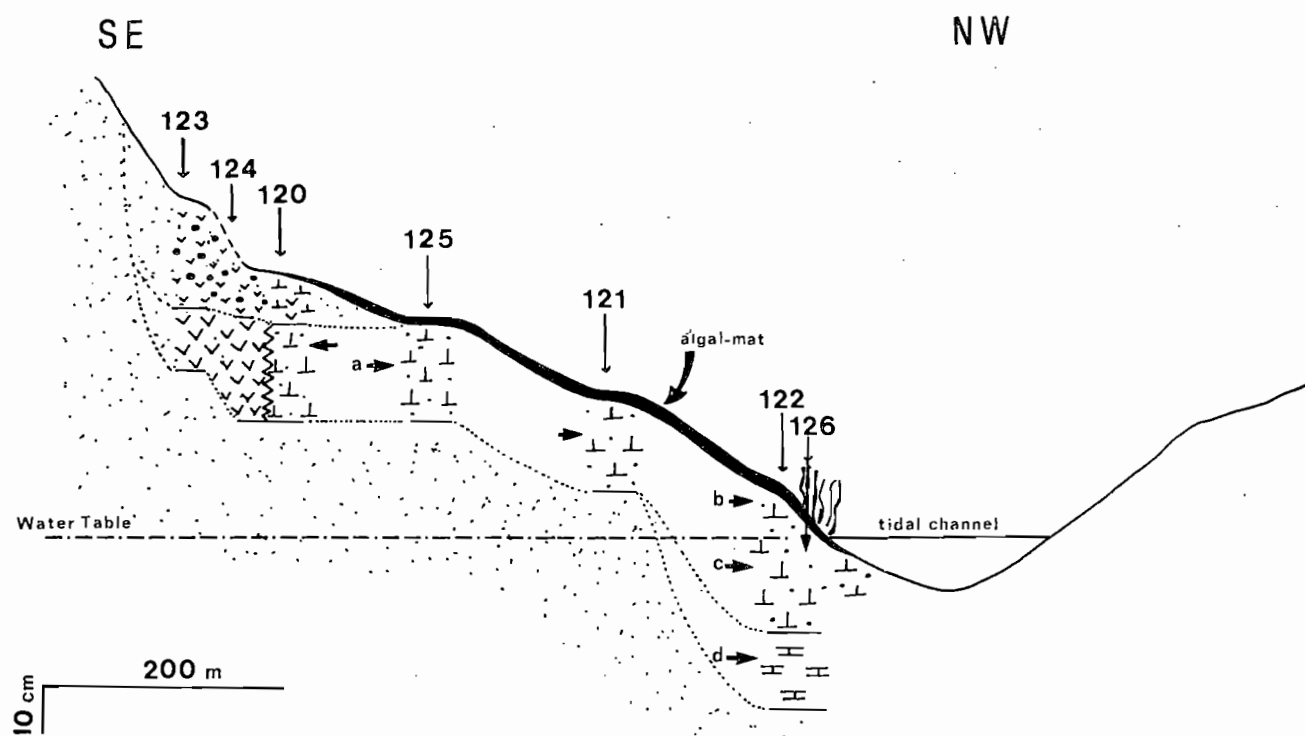


Figure 11. CROSS-SECTION E₂ OF THE SUPRATIDAL FLAT ON THE SOUTHERN SHORE OF OJO DE LIEBRE LAGOON: SAMPLES 120 TO 126 (SEE FIG. 10 FOR LEGEND, AND FIG. 9 FOR LOCALIZATION).

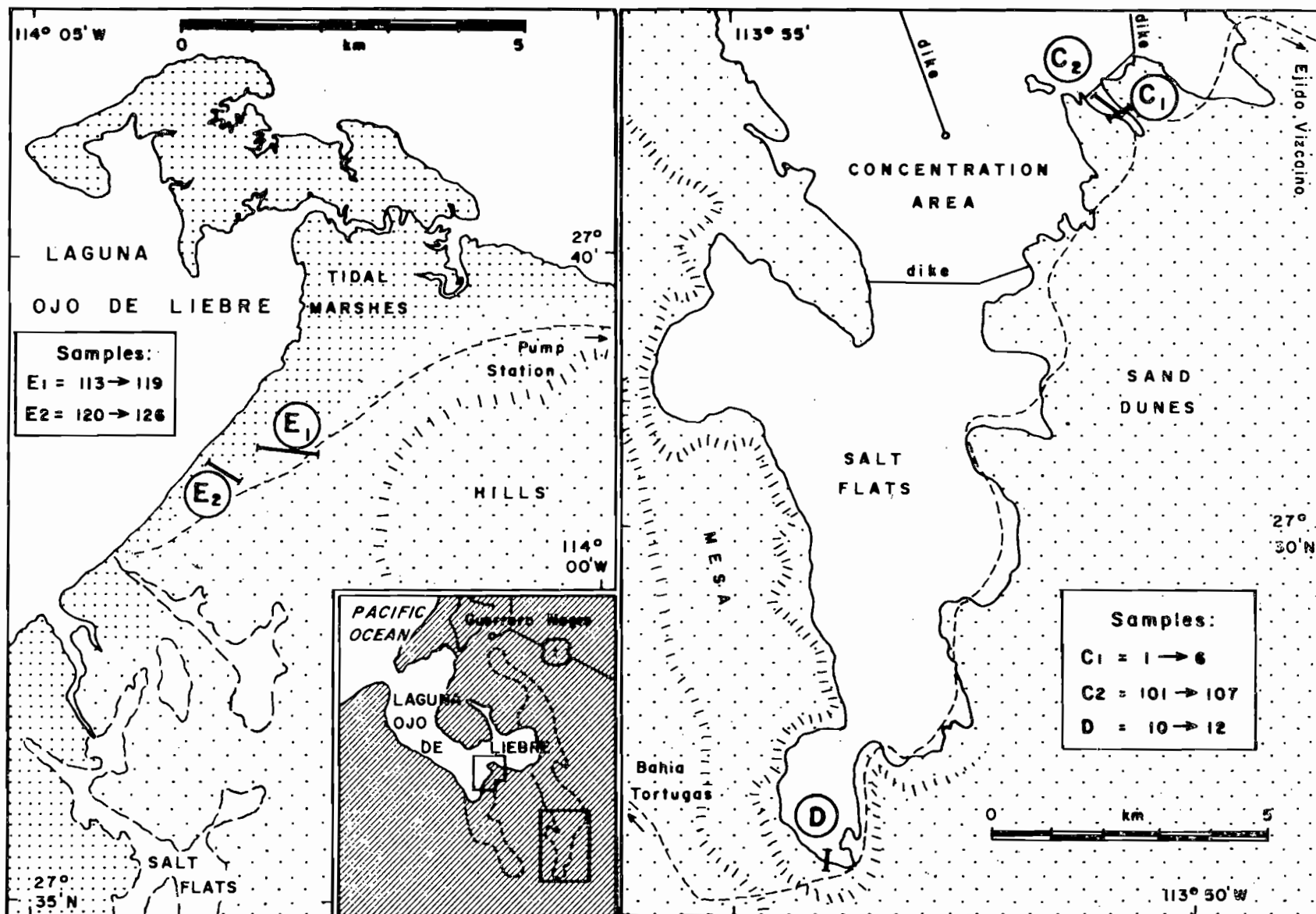


Figure 9. Localization of the analyzed samples in sections C₁, C₂, D, E₁ and E₂ in the southern area of Ojo de Liebre complex.

Table 4 = Physico-chemical and isotopic data on interstitial brines from supratidal flats on the southern shore of Ojo de Liebre Lagoon

	density	pH	Eh (mV)	t°C	$\delta^{18}\text{O}/\text{SMOW}$
L 113	1.078	7.2	+350	20.5	+1.4
L 114	1.090	7.75	+330	21.5	+1.57
L 115	1.092	7.75	+260	21	+1.74
L 116	1.091	7.6	+350	21.5	+2.47
L 117	1.086	7.6	+310	21.5	+2.27
L 118	1.069	7.15	+270	21	+2.57
L 119	not measured (collected in a tidal-channel)				+3.60
L 123	1.079	7.25	+370	21.5	+2.20
L 124	1.069	7.4	+320	20	+2.20
L 120	1.087	7.3	+430	21.5	+2.50
L 125	1.080	7.3	+330	21.5	+2.60
L 121	1.058	7.5	+320	19	+2.40
L 122	1.047	7.1	+40	19	+2.60

Table 5 = Mineralogy and isotope contents of carbonates from supratidal flats on the southern shore of Ojo de Liebre Lagoon

Sample	Associated minerals	Calcite (mole % MgCO_3)	Dolomite (mole % MgCO_3)	$\delta^{18}\text{O}$ / P.D.B. (bulk carbonate)	$\delta^{13}\text{C}$ / P.D.B. (bulk carbonate)
L 117a		Mg ₁₅	Mg ₃₆	+1.8	+2.0
L 117b	gypsum	Mg ₁₅	--	+1.4	+1.5
L 118a	gypsum aragonite	Mg ₁₄	Mg ₃₅ Mg ₃₈ Mg ₄₄	+2.0	+2.2
L 118b	aragonite	Mg ₁₄ Mg ₁₅ Mg ₂₃	Mg ₄₆	+2.2	+2.2
L 120		Mg ₁₂	--	+0.9	+1.7
L 125a	aragonite	Mg ₁₅	Mg ₃₁	+2.6	+2.7
L 121	aragonite quartz feldspath	Mg ₃ Mg ₁₄	--	+1.8	+3.0
L 122b	quartz feldspath	Mg ₁₃ Mg ₂₄	Mg ₃₈	+1.6	+0.2
L 122c	quartz feldspath	Mg ₁₄ Mg ₂₈	Mg ₄₃	+2.4	+0.1
L 122d	quartz feldspath	Mg ₁₄	Mg ₄₃	+2.3	-0.3
L 126	quartz feldspath	Mg ₅ Mg ₁₁	Mg ₄₅	+1.8	-0.9

Variations in pH values are slight (7,20 pH 7,75). Redox potentials (Eh) have very high values (+ 260 mV < Eh < + 430 mV) which characterize oxidizing conditions. At the salt-marsh edge, the low Eh value (Eh = + 40 mV) shows that slightly reducing conditions are occurring; this is also marked by the presence of free H₂S in the water.

Stable isotopes geochemistry

(See Tables 4 and 5)

Isotopes contents were measured in interstitial solutions for ¹⁸O, and in bulk carbonate for ¹⁸O and ¹³C.

Although densities are increasing landwards, $\delta^{18}\text{O}$ values of solutions are almost constant, with an average value of + 2, 5‰ (Fig. 23). At the landward edge of the flat, both ¹⁸O and salt contents are decreasing. These data indicate that marine derived brines may be diluted by continental groundwaters. Thus, part of the mineralization of solutions is probably related to gypsum dissolution.

A direct correlation is observed between the highest $\delta^{18}\text{O}$ values in the bulk carbonate and the occurrence of dolomite. Therefore, dolomitization was probably accompanied by an oxygen isotope enrichment relative to calcite which might crystallize in the same conditions (Pierre and Person, in press.). Many authors (Clayton & Epstein, 1958; Clayton et al; 1968; Engel et al; 1958; Fontes et al, 1967, 1969, 1970; Fontes & Perthuisot, 1971; Fritz & Smith, 1970; Matthews & Katz, 1977) have shown the existence of a dolomite-calcite ¹⁸O fractionation (by 3 to 4 ‰ at sedimentary temperatures) during the process of early dolomitization. It hence appears that dolomite crystallization at Ojo de Liebre is due to a very early diagenesis of a magnesian precursor; it is carried out via intermediate phases of high magnesian calcites.

¹³C contents of the carbonates indicate that their crystallization occurred either under the control of the oceanic reservoir in equilibrium with atmospheric CO₂ ($\delta^{13}\text{C} = + 2$ ‰), or in a semi-restricted environment where HCO₃ is partly biogenic (- 0, 9 ‰ < $\delta^{13}\text{C}$ < + 0,2 ‰). Furthermore, carbonate samples with low $\delta^{13}\text{C}$ values correspond to the lowest redox potentials measured in interstitial solutions.

Conclusions

Carbonate sediments from the supratidal flats of the southern shore of Ojo de Liebre lagoon are submitted to a very early diagenesis which leads to the formation of magnesian calcites and dolomites. Calcite represents the dolomite precursor, and interstitial solutions provide the magnesium required for the mineral transformation.

MINERALOGY AND GEOCHEMISTRY OF SABKA SEDIMENTS AND BRINES: SECTION D (C.P., L.O.)

Sedimentary deposits

(See Figs. 9 and 12)

The inner salt pan of the southernmost part of Ojo de Liebre complex is surrounded by a mesa about 20 metres high (Ortlieb and Pierre, in prep.) The Section D is located on the southern margin of this southernmost supratidal area of the Ojo de Liebre complex (Fig.9). The distribution of evaporites follows a circular pattern (bull's eye structure): halite deposits occupy the centre of the pan and gypsum deposits are located along the basin edge (Fig.12).

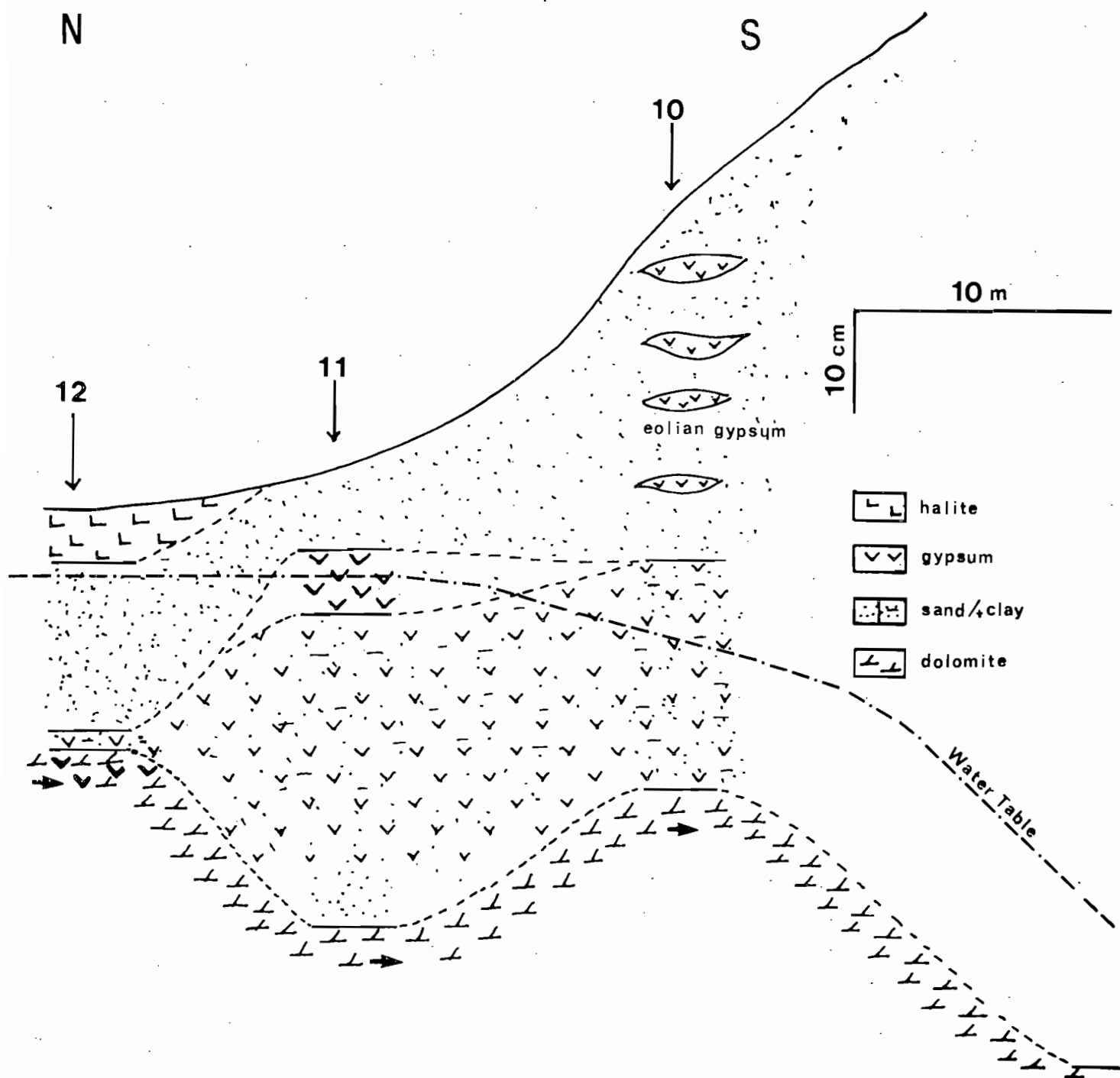


Figure 12. Cross-section D at the southern edge of the southernmost salt pan in Ojo de Liebre complex: samples 10, 11 and 12 (localization Fig. 9).

Gypsum crystals commonly have a lenticular habit which indicates that they have displacively grown in the host detrital sediment (Shearman, 1971). These crystals may be reworked by the wind and are locally accumulated at the base of the cliff. The gypsum deposits are underlain by a marly green sediment which contains a few large lensoid crystals of gypsum.

Mineralogy of sediments

Diagenesis in sediments from this locality is clearly demonstrated by their mineralogy. X-ray diffraction analysis show that the lower marly layer is almost entirely constituted by pure dolomite, with 5 moles % excess MgCO_3 in the crystal lattice. In spite of the lack of CaCO_3 in the sedimentary sequence, dolomitization was probably induced by the complete transformation of a CaCO_3 precursor with surrounding solutions enriched in magnesium. The dolomitization is accompanied by the crystallization of large gypsum crystals.

Interstitial solutions: physico-chemical data

(See Figs. 13 to 19 and Table 5)

Salt contents in interstitial brines increase while pH values decrease (in response to lower water activities) towards the centre of the pan. Redox potentials are positive although more reducing conditions predominate in the centre of the basin (where a sharp decrease in Eh, of more than 200 mV, is registered).

Concentrations in ionic species of the interstitial solutions do not correspond to normal marine brines. (K^+) and (Mg^{2+}) concentrations sharply increase in the centre of the basin where solutions have the highest densities. Total alkalinities are very low. (SO_4^{2-}) and (Ca^{2+}) concentrations respectively decrease and increase when densities are increasing. These results are in good agreement with the diagenesis expressed in the sediments: (Mg^{2+}) is used during dolomite formation while released (Ca^{2+}) reacts with (SO_4^{2-}) to form gypsum.

Stable isotopes geochemistry

(See Figs. 20 and Tables 6 and 7)

Isotopic analysis were carried out on both waters and dolomites. $\delta^{18}\text{O}$ values in waters are lower than for evaporated marine waters ($-0,2 \text{ ‰} < \delta^{18}\text{O} < +1,3 \text{ ‰}$); they are most probably due to a dilution of marine brines by continental waters.

Heavy isotopes contents in dolomites are very high for ^{18}O ($+3,0 \text{ ‰} < \delta^{18}\text{O} < +3,8 \text{ ‰}$) and are low for ^{13}C ($-5,3 \text{ ‰} < \delta^{13}\text{C} < -3,7 \text{ ‰}$). $\delta^{18}\text{O}$ values indicate that dolomitization was accompanied by an oxygen isotope enrichment (close to $3,7 \text{ ‰}$) relative to calcite which might crystallize in the same environmental conditions (temperature and $\delta^{18}\text{O}$ of water); as a result, these dolomites appear to have crystallized in an early stage of diagenesis.

Low $\delta^{13}\text{C}$ values show that large inputs of biogenic CO_2 were provided during the dolomitization process. Lippmann (1973) previously emphasized the importance of biogenic activity in dolomite formation since increasing alkalinities enhance the reaction.

Conclusions

The diagenesis which arises in the evaporitic sequence of the southernmost part of the Ojo de Liebre complex corresponds to an early dolomitization. The reaction leads to the complete transformation of an inferred CaCO_3 precursor by magnesium-rich solutions. The mineral transformation was accompanied by an oxygen isotope enrichment and was probably supported by inputs of biogenic CO_2 .

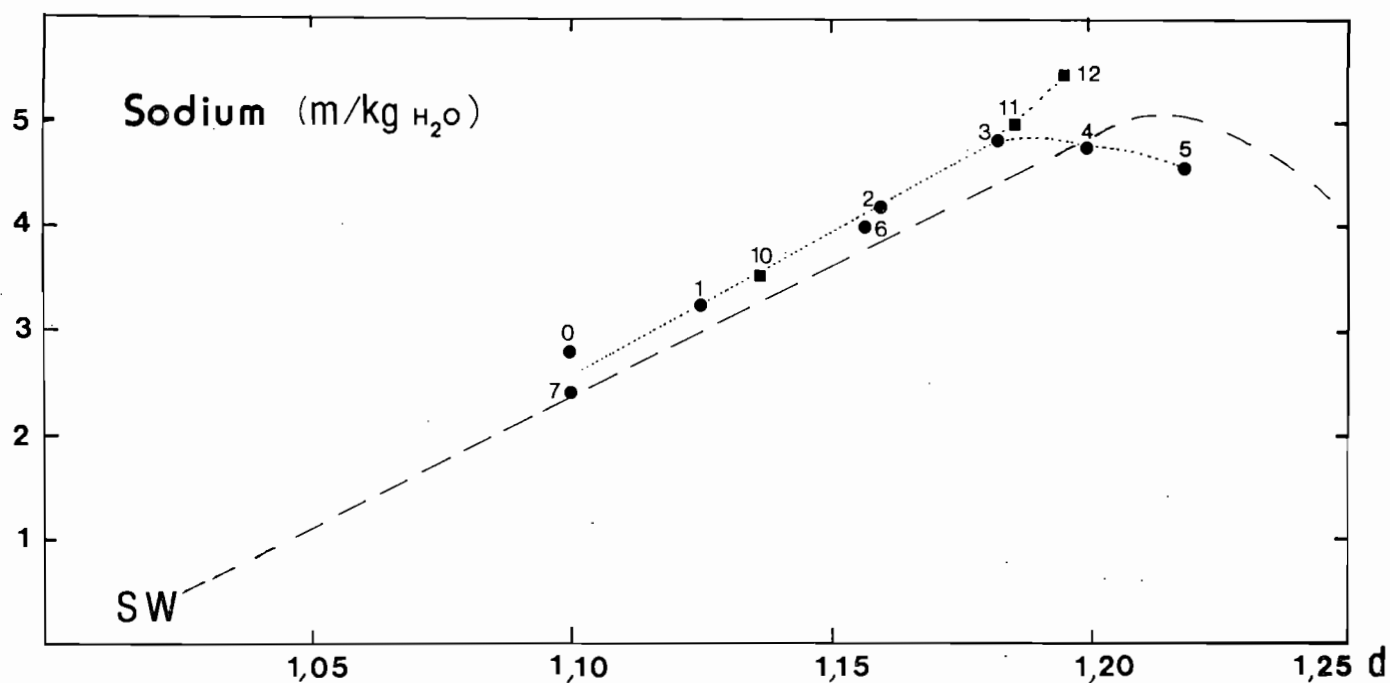


Figure 13. VARIATIONS IN (Na^+) CONCENTRATIONS, EXPRESSED IN MOLALITIES, RELATIVE TO DENSITY IN INTERSTITIAL SOLUTIONS (SECTIONS C_1 AND C_2); EVOLUTION OF SEA WATER (SW) DURING EVAPO- RATIONS IS GIVEN BY COMPARISON.

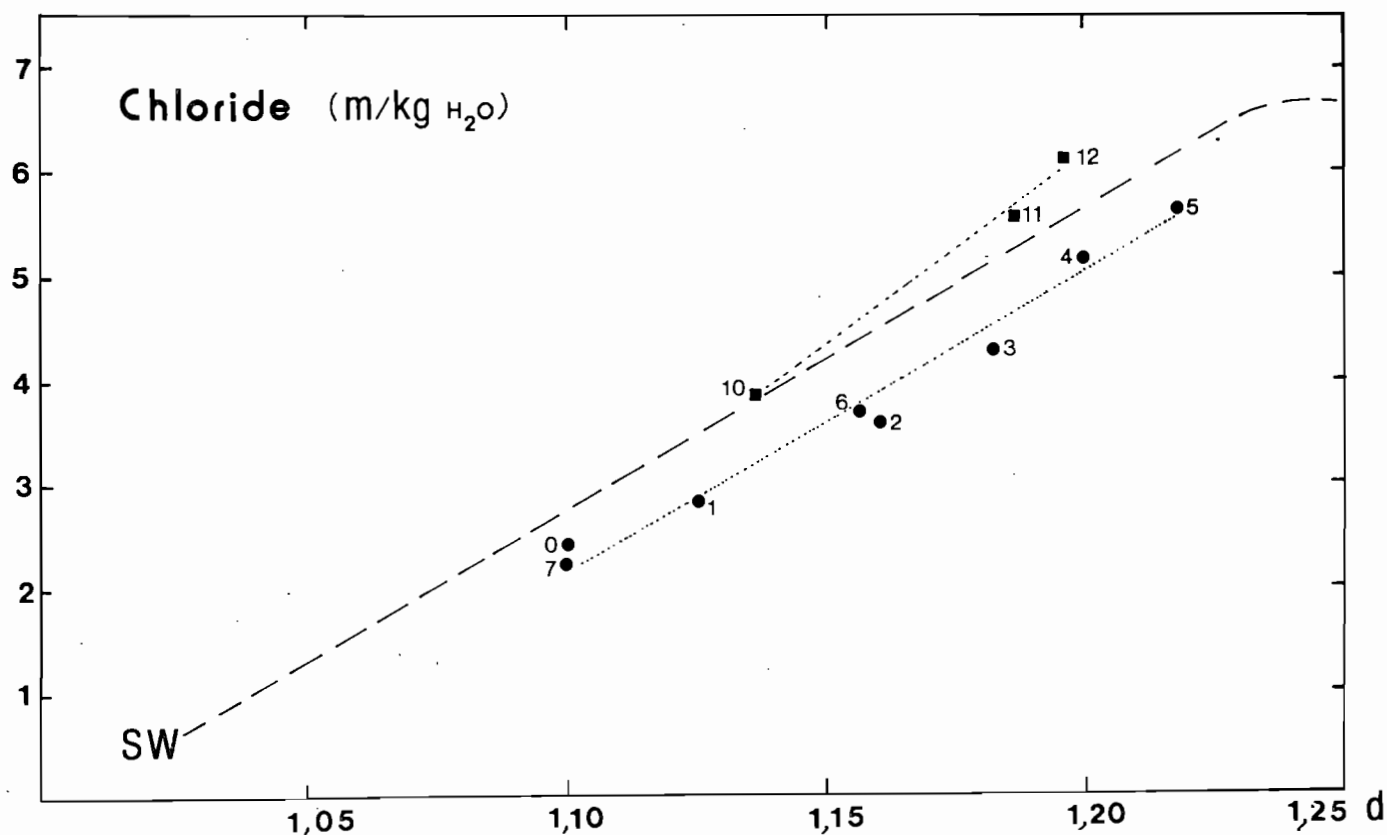


Figure 14. VARIATIONS IN (Cl^-) CONCENTRATIONS, EXPRESSED IN MOLALITIES, RELATIVE TO DENSITY IN INTERSTITIAL SOLUTIONS (SECTIONS C_1 AND C_2); EVOLUTION OF SEA WATER DURING EVAPORATION IS GIVEN BY COMPARISON.

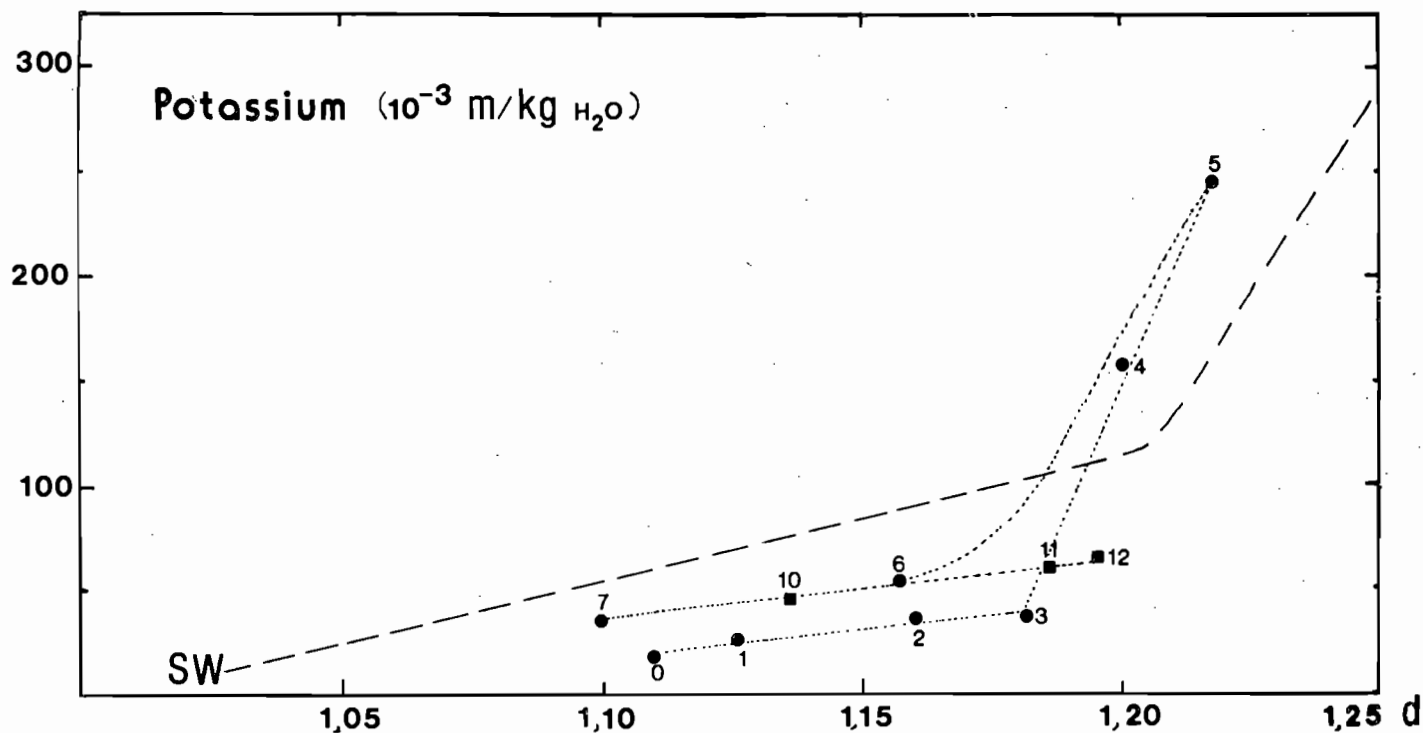


Figure 15. VARIATIONS IN (K^+) CONCENTRATIONS, EXPRESSED IN MOLALITIES, RELATIVE TO DENSITY IN INTERSTITIAL SOLUTIONS (SECTIONS C_1 AND C_2); EVOLUTION OF SEA WATER DURING EVAPORATION IS GIVEN BY COMPARISON.

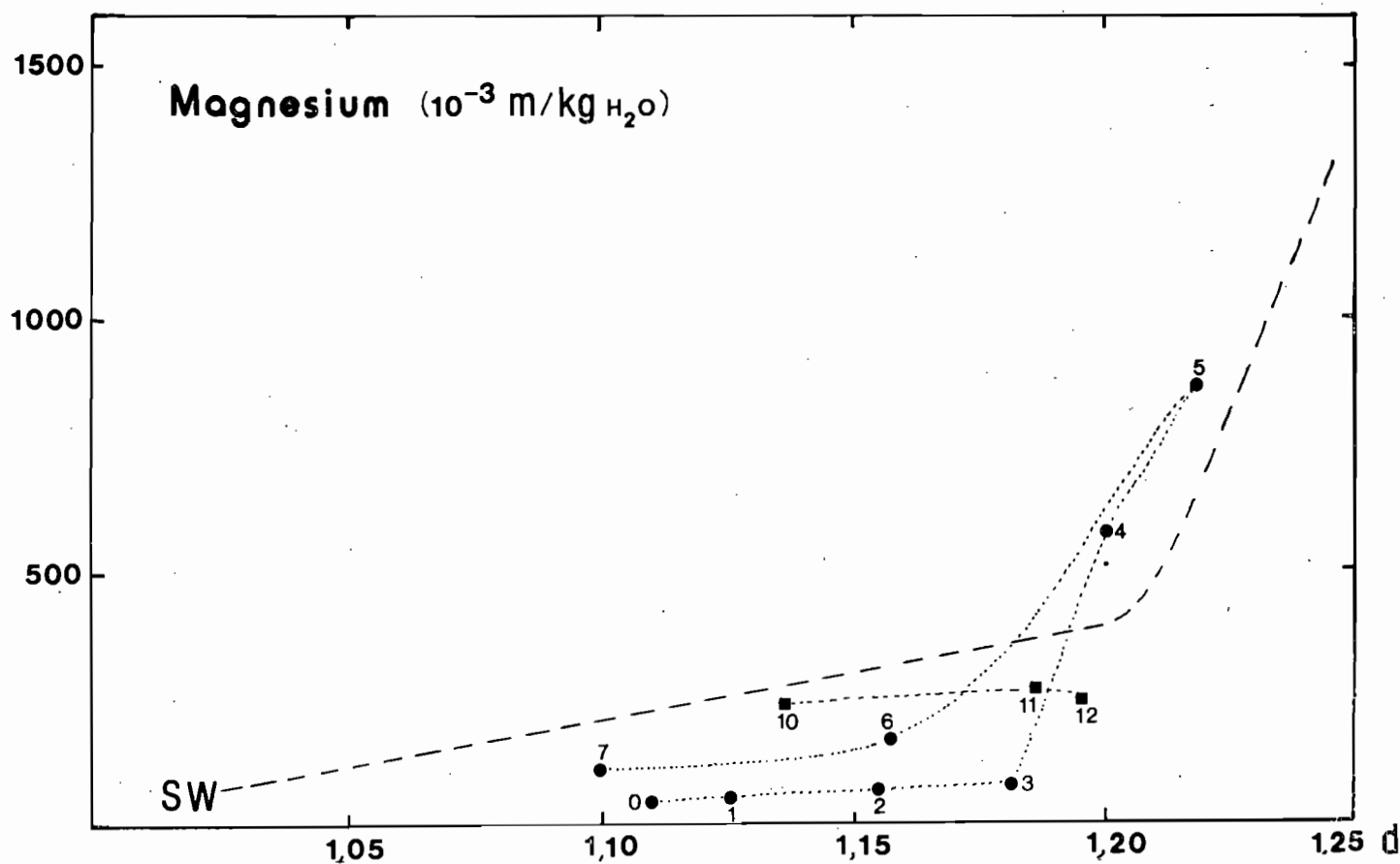


Figure 16. VARIATIONS IN (Mg^{2+}) CONCENTRATIONS, EXPRESSED IN MOLALITIES, RELATIVE TO DENSITY IN INTERSTITIAL SOLUTIONS (SECTIONS C_1 AND C_2); EVOLUTION OF SEA WATER DURING EVAPORATION IS GIVEN BY COMPARISON.

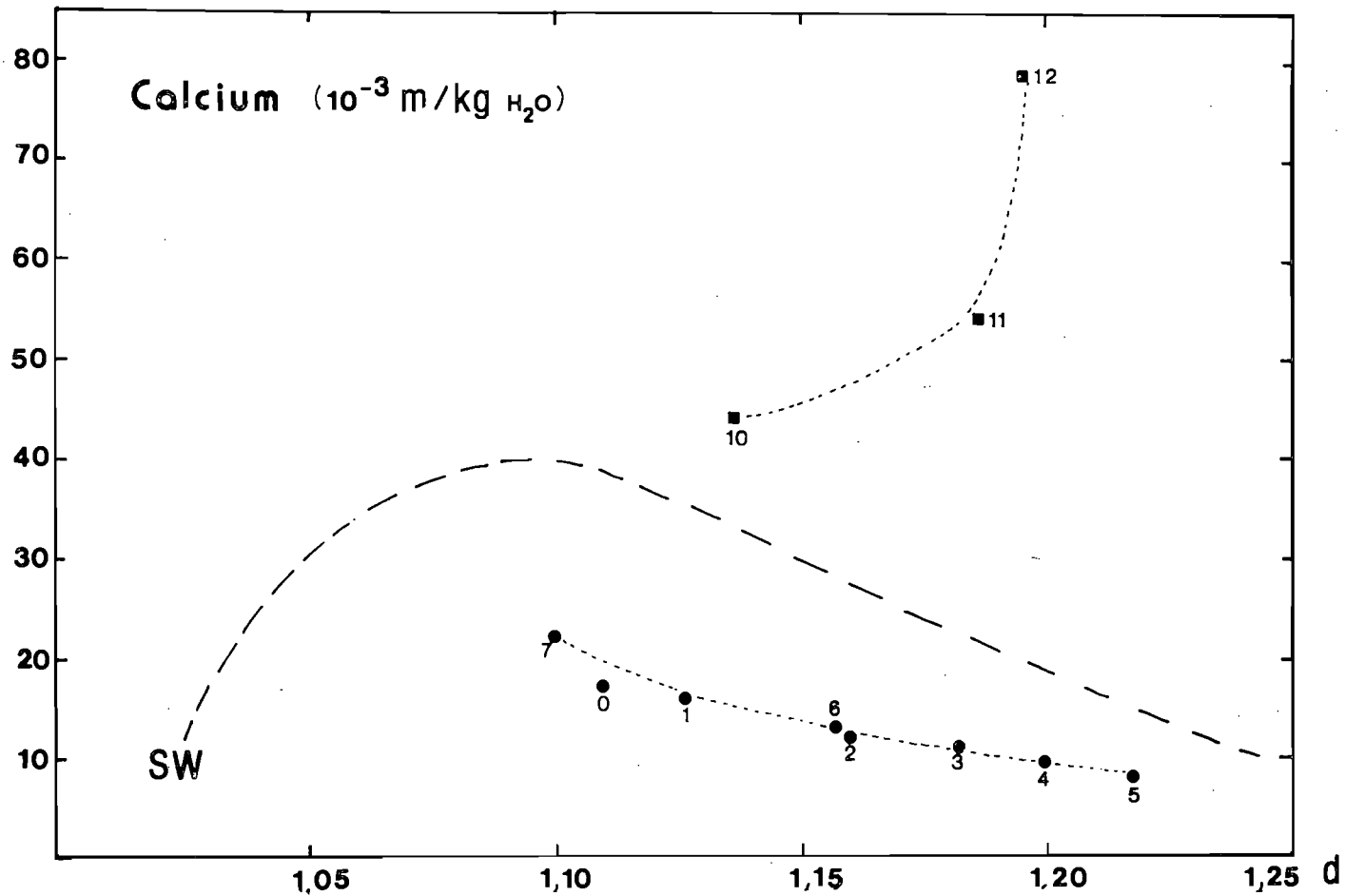


Figure 17. VARIATIONS IN (Ca^{2+}) CONCENTRATIONS, EXPRESSED IN MOLALITIES, RELATIVE TO DENSITY IN INTERSTITIAL SOLUTIONS (SECTIONS C_1 AND C_2); EVOLUTION OF SEA WATER DURING EVAPORATION IS GIVEN BY COMPARISON.

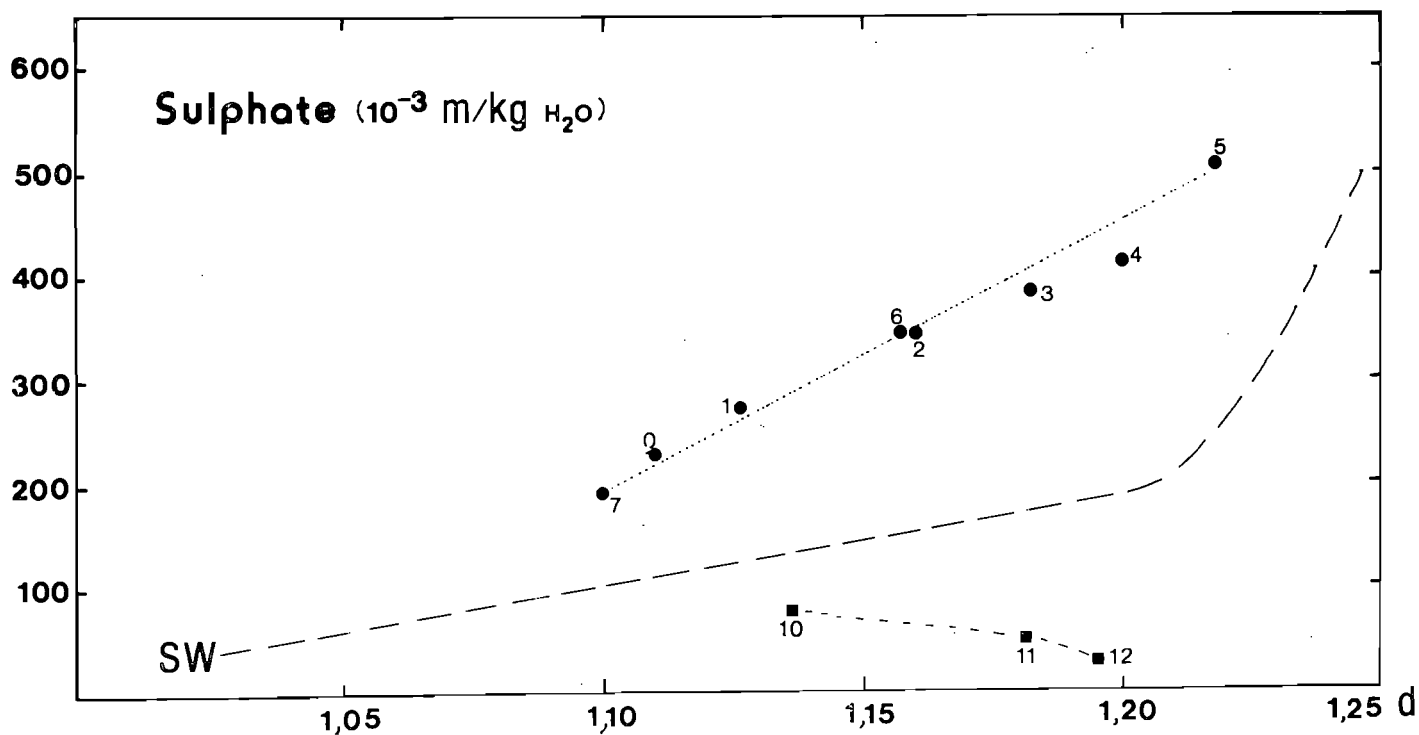


Figure 18. VARIATIONS IN (SO_4^{2-}) CONCENTRATIONS, EXPRESSED IN MOLALITIES, RELATIVE TO DENSITY IN INTERSTITIAL SOLUTIONS (SECTION C_1 AND C_2); EVOLUTION OF SEA WATER DURING EVAPORATION IS GIVEN BY COMPARISON.

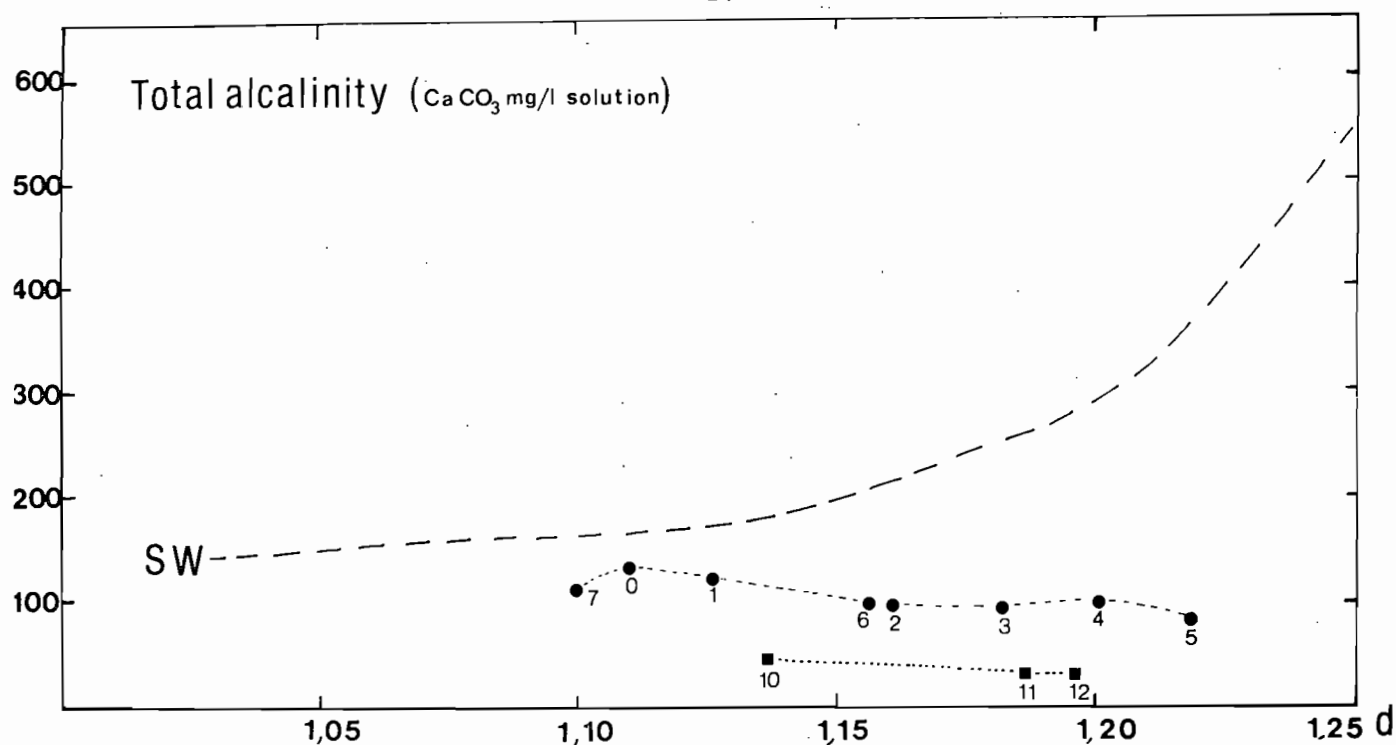


Figure 19. VARIATIONS IN TOTAL ALCALINITY, EXPRESSED IN CaCO_3 MOLARITIES, RELATIVE TO DENSITY IN INTERSTITIAL SOLUTIONS (SECTIONS C_1 AND C_2); EVOLUTION OF SEA WATER DURING EVAPORATION IS GIVEN BY COMPARISON.

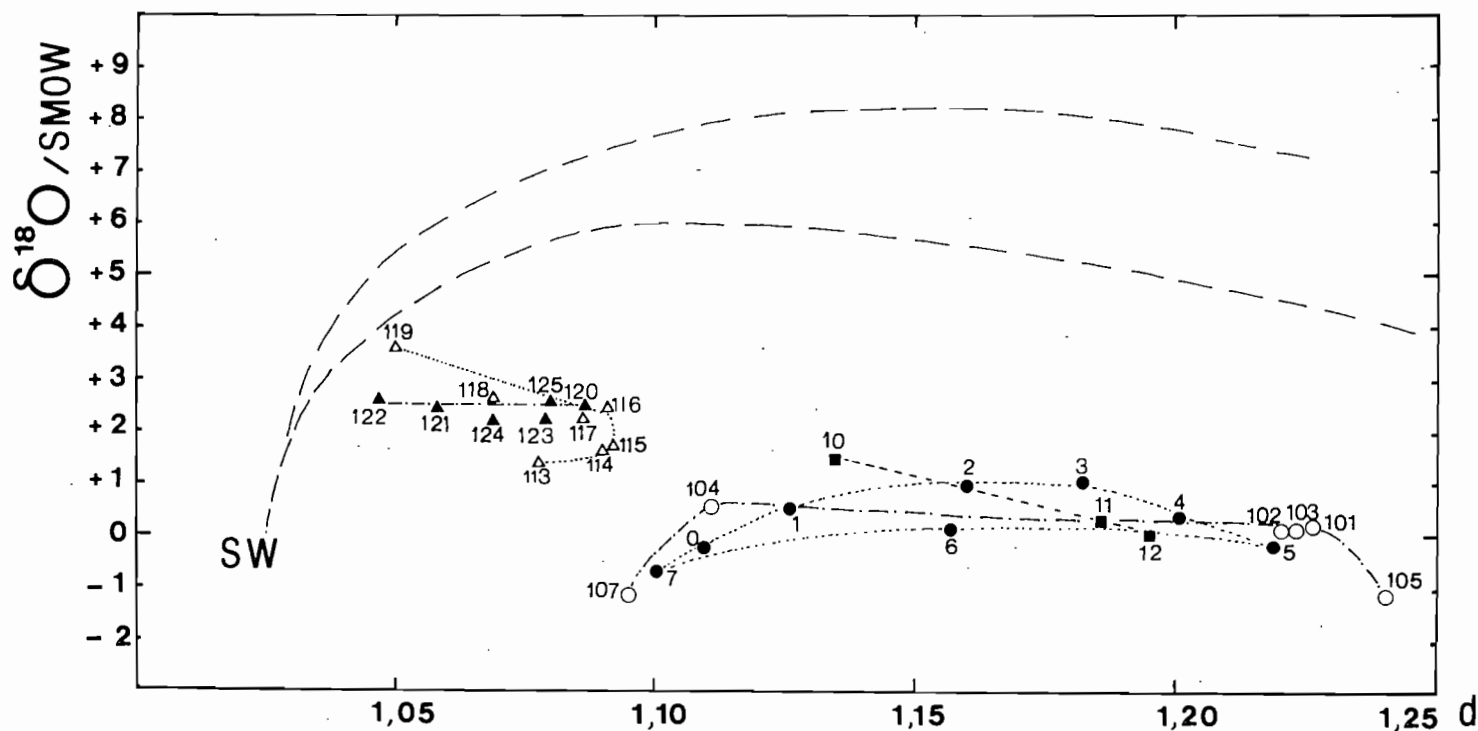


Figure 20. OXYGEN ISOTOPE COMPOSITIONS OF INTERSTITIAL SOLUTIONS (RELATIVE TO DENSITIES) FROM ALL THE STUDIED SAMPLES IN THE SOUTHERN PART OF THE OJO DE LIEBRE COMPLEX; EVOLUTION OF SEA WATER BODY DURING EVAPORATION IS GIVEN BY COMPARISON.

Table 6 = Physico-chemical, chemical and isotopic data of interstitial solutions from the salt-pan of the southernmost part of Ojo de Liebre complex

	density	pH	Eh (mV)	t°C	Na ⁺ m/kg	Cl ⁻ m/kg	K ⁺ 10 ⁻³ m/kg	Mg ²⁺ 10 ⁻³ m/kg	Ca ²⁺ 10 ⁻³ m/kg	SO ₄ ²⁻ 10 ⁻³ m/kg	T.A.C. CaCO ₃ mg/l	δ ¹⁸ O/SMOW (H ₂ O)
L 10	1.136	6.80	+460	22.5	3.53	3.87	46.7	227.4	44.1	78.1	40	+1.30
L 11	1.186	6.55	+440	23	4.94	5.54	61.5	265.7	54.2	50.3	31	-0.05
L 12	1.195	6.30	+200	22	5.40	6.09	65.7	243.5	78.6	30.6	33	-0.20

Table 7 = Chemical and isotopic compositions of dolomites from the salt-pan in the southernmost part of Ojo de Liebre complex

	Mole % MgCO ₃	δ ¹⁸ O/P.D.B.	δ ¹³ C/ P.D.B.
L 10	45	+4.61 (+3.81*)	-4.40
L 11	45.5	+4.20 (+3.40*)	-3.71
L 12	46	+3.82 (+3.02*)	-5.27

* ¹⁸O values corrected for the isotopic fractionation during the H₃PO₄ reaction (Sharma and Clayton 65)

MINERALOGY AND GEOCHEMISTRY OF SABKA SEDIMENTS AND BRINES:
SECTIONS C₁ & C₂ (C.P., L.O.)

Sedimentary deposits

(See Figs. 9, 21, 22 and 23)

After the first discovery of polyhalite ($K_2MgCa_2(SO_4)_4 \cdot 2H_2O$) by Holser (1966), and when the supratidal areas of Ojo de Liebre complex already had been transformed for the brines concentration, more occurrences of this evaporitic mineral were searched for, peculiarly along the eastern and southeastern margins of the Ojo de Liebre complex. Polyhalite was finally found in the southeastern part of the complex, in a salt pond recently isolated from the evaporitic flats by a new dike. The section C₁, across the basin, was studied in 1979 and 1980. The section C₂, oriented along the main axis of the pond and ending near the dike, was studied in 1980. Location of these two sections is given in Fig. 9. Descriptions of the geological setting and of some sedimentological aspects of this area will be published promptly (Ortlieb and Pierre, in preparation).

In section C₁, the evaporite sediments are mainly represented by gypsum with a maximal thickness of 30 cm in the centre of the basin (Fig.21). The gypsum crystals have a lenticular habit which indicates that they have grown displacively in the host detrital sediment (Shearman, 1971); they become larger (up to 1 cm) at depth and may form hard concretions (Pierre and Ortlieb, in press). In may 1979, polyhalite was only present, as small white nodules, at a few centimetres below the surface and about 10 cm above the water table. The formation of polyhalite obviously resulted from the diagenetic alteration of former gypsum (which may be observed in the core of some nodules).

In may 1980, the evaporitic sequence was strongly modified, since the lowest gypsum layers were completely replaced by a few centimetres thick polyhalite layer, somewhat indurated (Fig.22). The polyhalite was lying just below the water table, the level of which was slightly lower than a year before. These observations argue for the diagenetic replacement of gypsum by polyhalite; this replacement appears to be a very rapid process due to variations in brine chemistry (Pierre, 1981).

In section C₂ at the northwestern margin of the studied basin where solutions are seeping from the adjacent basin, a hard layer (more than 10 cm thick) of halite associated with gypsum was found at a 15 cm depth (Fig.22). Halite and gypsum crystals are large (up to 1 cm) and entirely invaded by sand grains; this argues for the diagenetic emplacement of both halite and gypsum in the sandy host sediment. Gypsum crystals may form clusters, and appear as the first diagenetic generation since some gypsum crystals are included in halite crystals.

Interstitial solutions: physico-chemical data

(See Figs. 13 to 19 and Table 8)

Brine densities are increasing and pH values are decreasing (due to decreasing water activities) from the edge towards the centre of the pond. Very high redox potential values indicate that environmental conditions are oxidizing.

Ionic concentrations of solutions collected in 1979 were not representative of normal marine derived brines. The main variations were related to higher (SO_4^{2-}) and lower (Ca^{2+}) concentrations, and to low total alkalinities. These data may be interpreted as the result of gypsum dissolution, and of a possible $CaCO_3$ crystallization.

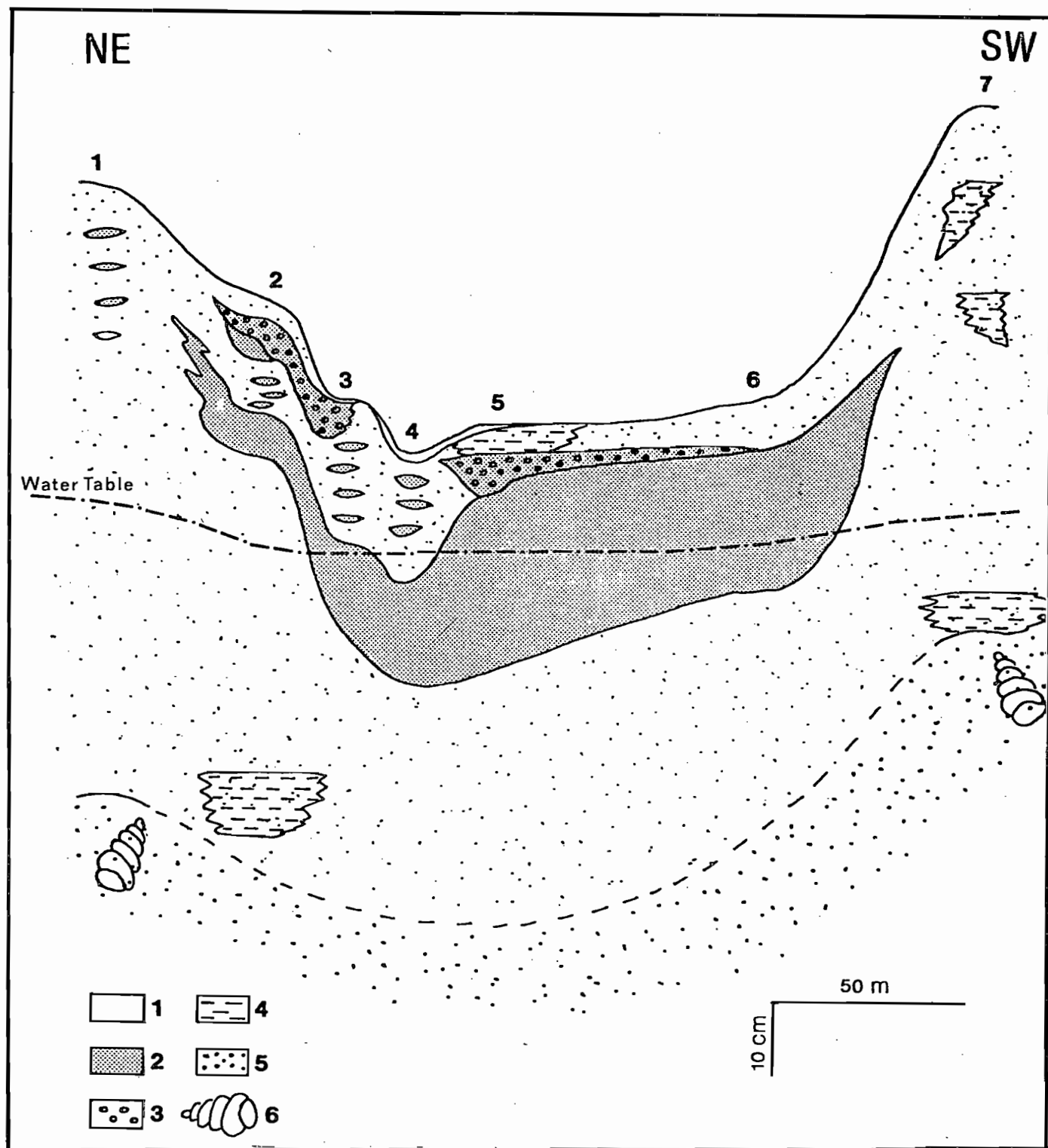


Figure 21. Cross-section C₁ of a salt pond located in the southeastern part of Ojo de Liebre complex (in may 1979): samples 1 to 7 (localization Fig. 9). Legend: 1: halite; 2: gypsum; 3: polyhalite; 4: clay; 5: sand; 6: marine shells.

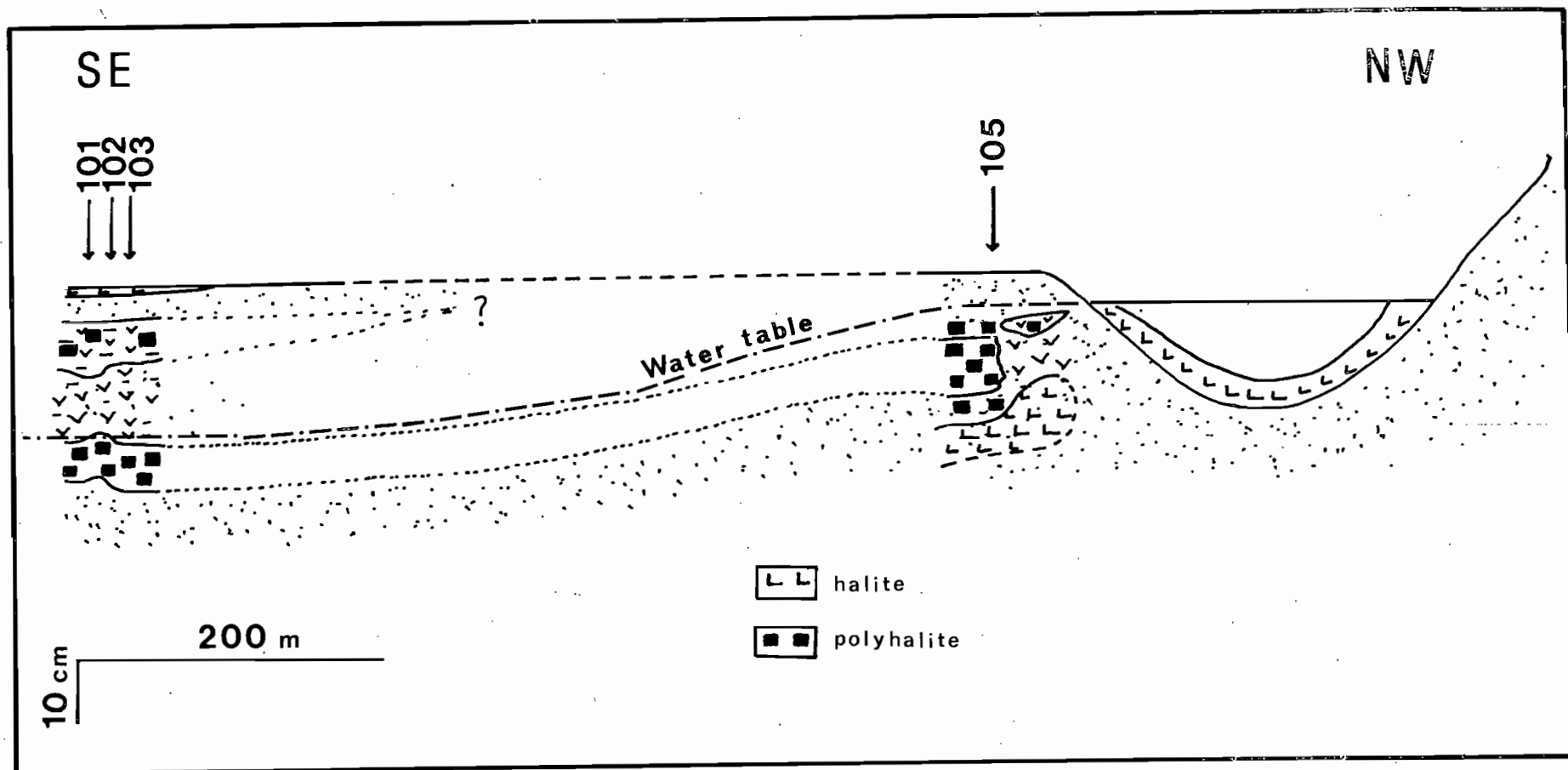


Figure 22. CROSS-SECTION C₂ IN THE NORTHWESTERN END OF A SALT POND IN THE SOUTHEASTERN PART OF OJO DE LIEBRE COMPLEX (IN MAY 1980): SAMPLES 101 TO 105 (LOCALIZATION FIG. 9).

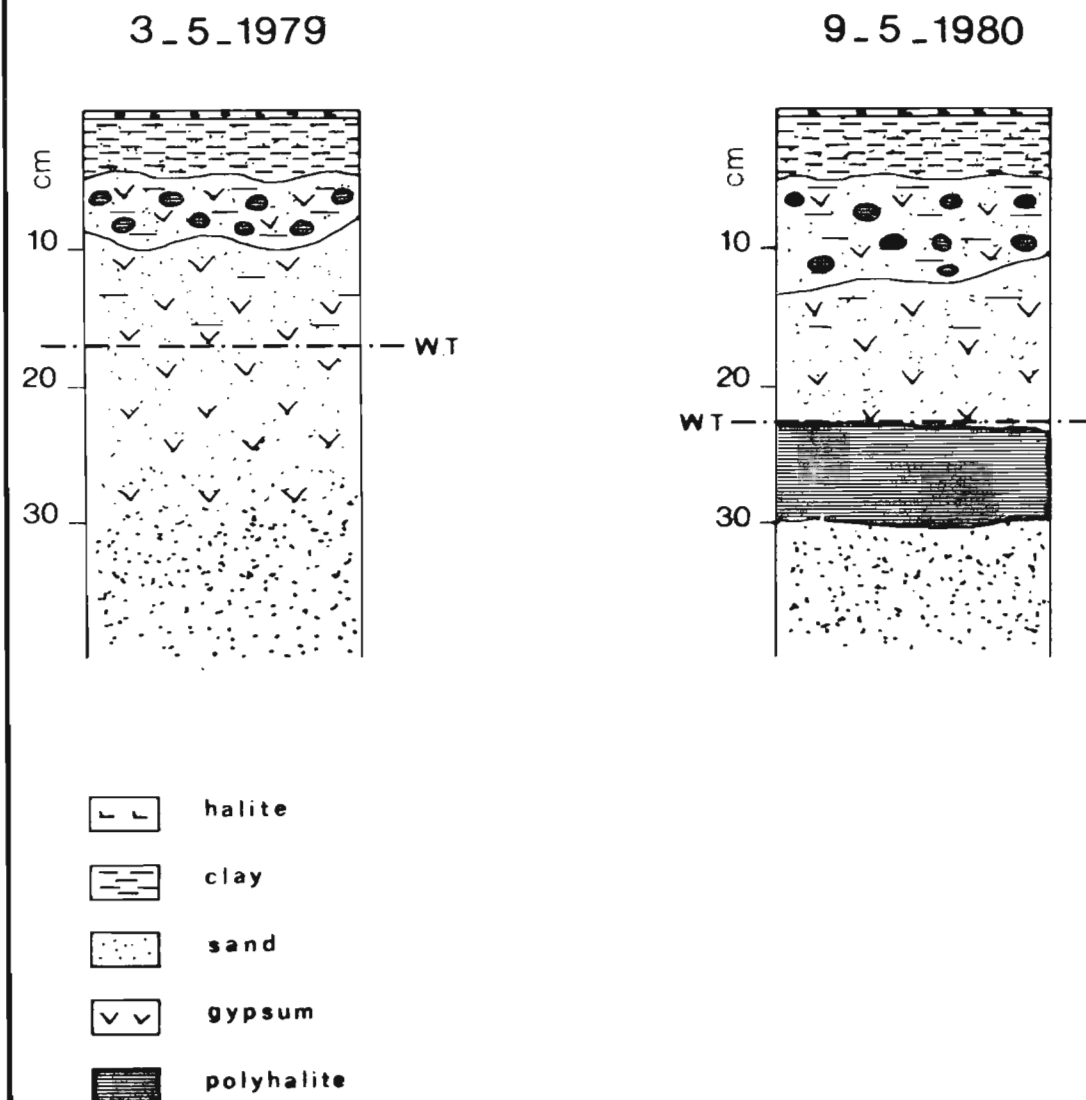


Figure 23. EVOLUTION BETWEEN MAY 1979 AND MAY 1980 OF A SEDIMENTARY SEQUENCE, IN THE CENTRE OF THE STUDIED SALT POND IN THE SOUTHEASTERN PART OF OJO DE LIEBRE COMPLEX.

Table 8 = Physico-chemical, chemical and isotopic data on interstitial solutions from supratidal areas in the southeastern part of Ojo de Liebre complex

	density	pH	Eh (mV)	t°C	Na ⁺ m/kg	Cl ⁻ m/kg	K ⁺ 10 ⁻³ m/kg	Mg ²⁺ 10 ⁻³ m/kg	Ca ²⁺ 10 ⁻³ m/kg	SO ₄ ²⁻ 10 ⁻³ m/kg	T.A.C. CaCO ₃ Mg/l	δ ¹⁸ O/SMOW (H ₂ O)	δ ¹⁸ O/SMOW (SO ₄)	δ ³⁴ S/C.D. (SO ₄)
L 0	1.110	not	measured		2.79	2.39	17.4	47.6	16.8	231.6	135	-0.2	+11.3	+18.5
L 1	1.126	7.45	+210	23.5	3.24	2.84	27.7	51.5	16.3	275.8	122	+0.5	+10.2	+18.1
L 2	1.160	7.30	+370	25	4.18	3.62	34.6	65.5	13.0	347.8	95	+0.9	+10.1	+17.3
L 3	1.182	7.15	+360	26	4.83	4.32	38.1	73.7	11.0	385.5	92	+0.9	+10.5	+18.3
L 4	1.200	6.75	+370	26	4.74	5.13	157.2	577.4	9.4	416.4	100	-0.4	+ 9.5	+18.1
L 5	1.218	6.15	+410	25.5	4.59	5.66	244.6	873.6	8.0	512.5	78	-0.1	+10	+19.0
L 6	1.157	7.10	+330	26.5	4.02	3.68	56.1	169.0	13.7	343.6	102	+0.1	+ 9.5	+17.8
L 7	1.100	7.15	+390	23	2.39	2.24	36.6	106.6	22.4	191.1	112	-0.7	+ 9.7	+17.4
L 101	1.226	6.40	+340	23								+0.2		
L 102	1.220	6.50	+285	23.5								+0.1		
L 103	1.223	5.55	+280	23								+0.1		
L 104	1.111	8.00	+340	23								+0.6		
L 105	1.240	6.30	+400	24								-1.1		
L 107	1.095	7.80	+390	24								-1.2		

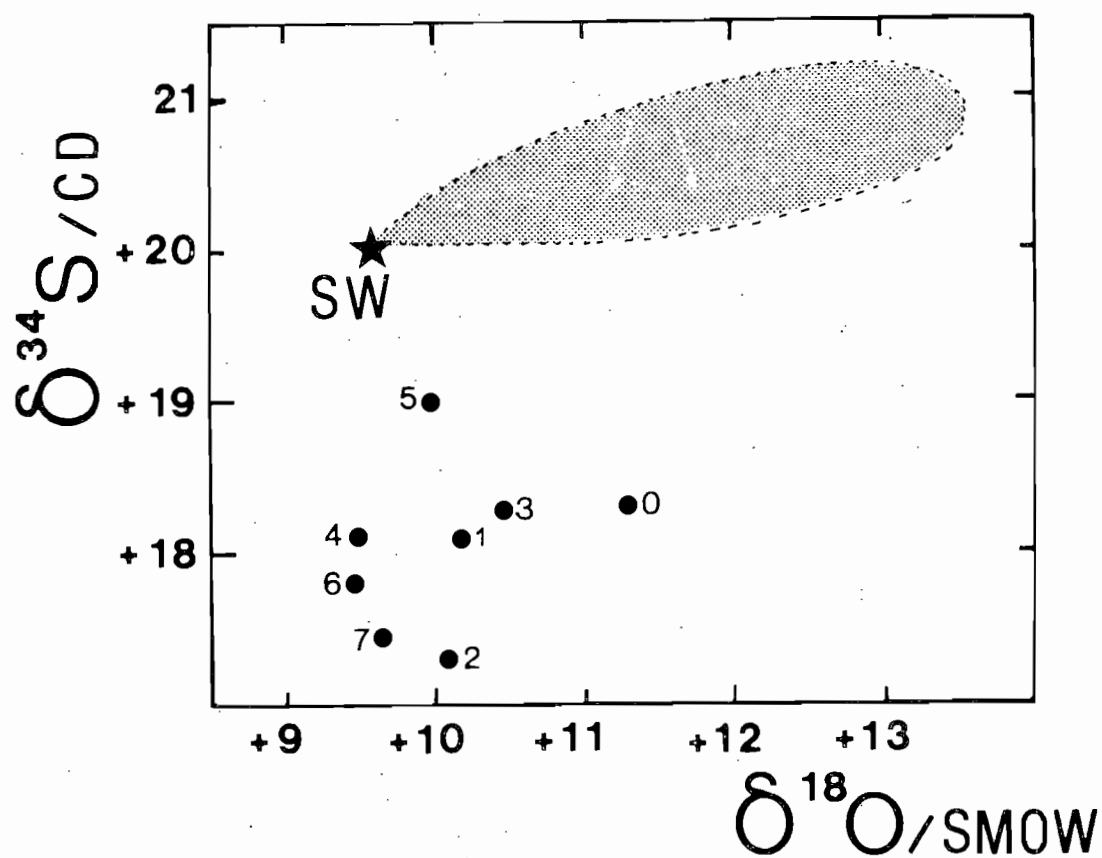


Figure 24. Oxygen and sulphur isotope contents in aqueous sulphate of interstitial solutions from the southeastern supratidal flat. The shaded area represents values observed in salt ponds from the south of France (Fontes & Pierre, 1978; Pierre, unpublished).

Stable isotopes geochemistry

(See Figs. 20 and 24 and Table 8)

The heavy isotopes contents were measured for ^{18}O in waters (collected both in 1979 and 1980), and for ^{18}O and ^{34}S in dissolved sulphate from waters collected in 1979.

$\delta^{18}\text{O}$ values of interstitial brines collected in 1979 and in 1980 are quite similar; they have a small range of variations ($-1,2\text{‰} < \delta^{18}\text{O} < +1,0\text{‰}$) and do not correspond to evaporated marine waters. It appears that a large fraction of meteoric water is responsible for the negative $\delta^{18}\text{O}$ values. It hence follows that solutions are resulting from a mixing of marine brines (seeping from the adjacent basin) and of continental waters (Pierre and Ortlieb, in press).

$\delta^{18}\text{O}$ and $\delta^{34}\text{S}$ contents in dissolved sulphate have a wide range of variations ($+9,5\text{‰} < \delta^{18}\text{O} < +11,3\text{‰}$; $+17,3\text{‰} < \delta^{34}\text{S} < +19,0\text{‰}$). These values are far from those of the sea water sulphate ($\delta^{18}\text{O} = +9,5\text{‰}$; $\delta^{34}\text{S} = +20\text{‰}$). Since oxidizing conditions preclude any bacterial sulphate reduction in this basin, the δ values are interpreted as resulting from the mixing of two different sulphate reservoirs. One fraction of sulphate might be provided by marine brines while the other part might come from the dissolution of gypsum by continental waters.

Conclusions

The sulphate diagenesis is expressed in the evaporite deposits of supratidal flats from the southeastern part of the Ojo de Liebre complex by crystallization and dissolution of gypsum and by the replacement of gypsum by polyhalite. Isotopic data show that interstitial solutions are a mixing of marine derived brines and of continental waters. Aqueous sulphate has a double origin; one part is marine derived; the other part comes from the dissolution of sulphate minerals by continental waters. It is probable that in this basin, both high contents in magnesium and potassium in marine brines, as well as large inputs of sulphate ions (by gypsum dissolution) are enhancing the formation of polyhalite.

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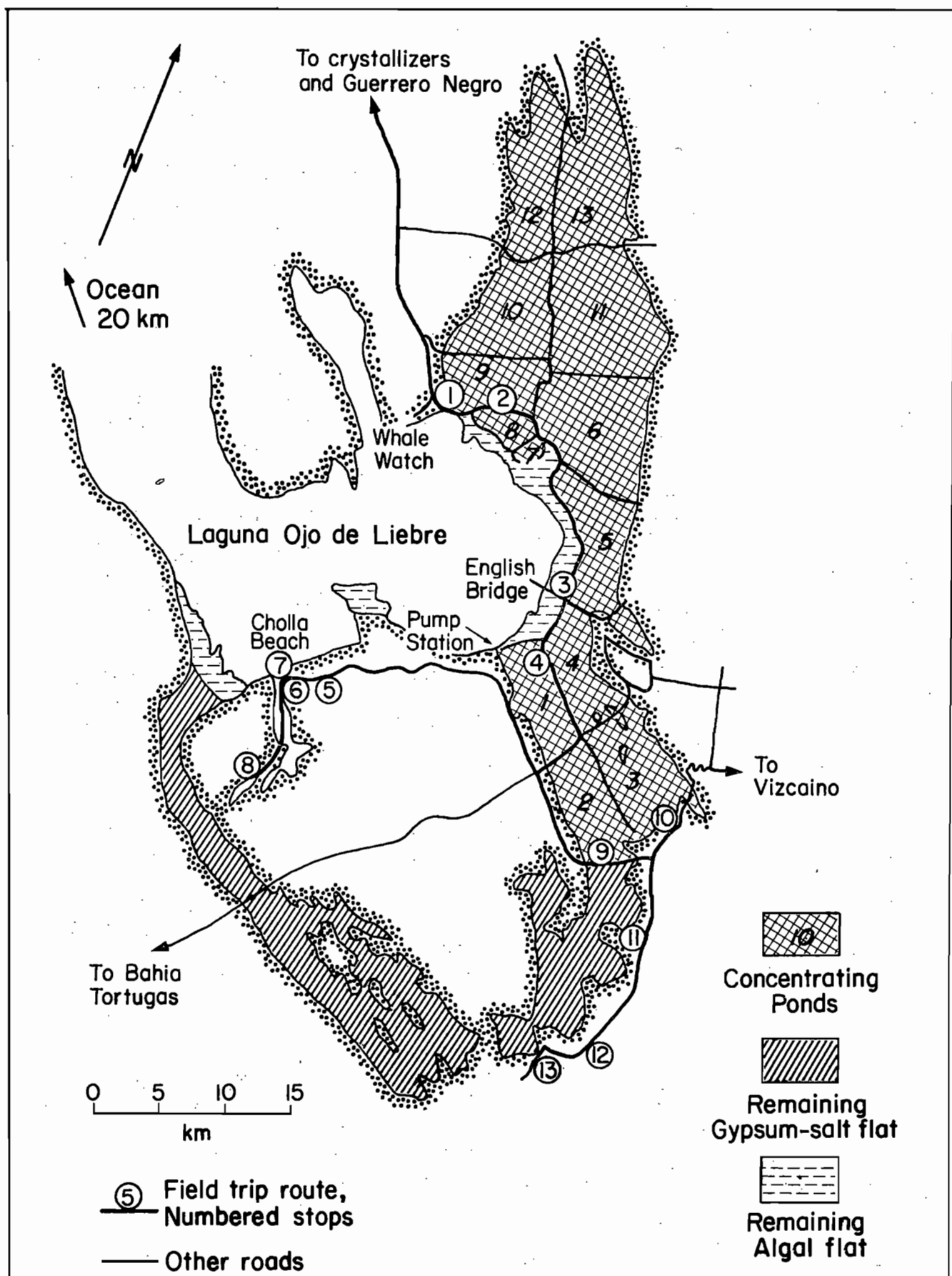
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GEOCHEMISTRY AND ECOLOGY OF SALT PANS AT GUERRERO NEGRO, BAJA, CALIFORNIA

km
(mi)

ROAD LOG

- 0
(0) Leave gates through the salt company facilities at Guerrero Negro. The first several miles lead through NaCl crystallizing ponds followed by low, stabilized dunes of quartz sands, shells, and detritus from an earlier lagoon cycle. Sea water concentrating ponds (gypsum stage) are visible in the distance on the left (east).
- 23.8
(14.9) Road junction. At the right is the turn-off to the whale-watching park. Straight ahead are low dunes, blue-green algal flats (Lyngbya aestuarii and Calothrix crustacea), salt marshes, and Laguna Ojo de Liebre. To the left is sea water concentrator Area 9 which is characterized by a floor of gypsum. Note the difference in the crystal habits on the undersides (subaqueous) and surfaces of the gypsum crusts, as well as the "teepee" structures. The underside of the crust is colonized by blue-green algae (coccoid, Aphanothece halophytica), below which are purple photosynthetic bacteria. Ephydra fly pupal cases are common.
- 24.8
(15.5) Follow the road east past the former seawater pumping station.
- 25.0
(15.6) Turn left on dike 9 just before arriving at a former pumping station. On the left is concentrator 9 (annual range of density 1.103-1.133). On the right is concentrator 8 (annual range of density 1.094-1.111). Concentrator 8 is characterized by a soft algal mat floor at the entrance (largely coccoid blue-green algae and diatoms), with gradually more and more sand-size crystals of gypsum at the base of the photosynthetic zone (about 1 cm below the mat surface) in the higher salinity areas of the pond. Cores of the mat (several cm thick) demonstrate that gypsum also precipitates at the base of the older photosynthetic layers. Area 9 is characterized by algal and gypsum-sand mats at the entrance but most of this concentrator is covered with a firm gypsum crust. The algal mat covers the surface of the crust at lower salinities and eventually only the undersides of the crust at higher salinities. Note the "hummocky" nature of the gypsum crust in Area 9 along the dike (lithified ripples).
- 28.5
(17.8) Junction of dikes 8 and 9 (Areas 6, 8, and 9). Turn right. Concentrator 6 (annual range of density 1.066-1.090) is characterized by the same mat as in 8, but gypsum precipitation does not occur at the lower densities in this pond. The orange color of the algae is caused by the predominance of carotenoid pigments.
- 29.3
(18.3) Junction of dikes 7 and 8 (concentrators 6, 7, and 8). Concentrator 7 was filled in Winter, 1979, but fauna, flora, and sediments have not yet stabilized.



- 30.1
(18.8) Former pumping channel. This is the beginning of the route characterized by natural salt marshes and algal flats on the right (west) which are truncated by the dike and sea water concentrators on the left. This area corresponds to the toe of the boot-shaped Laguna Ojo de Liebre.
- 31.4
(19.6) Corner of Areas 5 and 6. Continue to the right.
- 33.0
(20.6) "Patch" in dike where old gypsum floor was broken during construction. Because so much brine leaked out from the pond, a new retaining wall was built using the nearly impermeable gypsum crust as the floor. Concentrator 5 algal mats are dominated by filamentous blue-green algae (largely Microcoleus chthonoplastes). Diatoms and photosynthetic bacteria also occur (annual range of density 1.047-1.070).
- (possible additional stop)
- 34.7
(21.7) Note the presence of large dark clumps in the dike wall. These are old algal mats excavated during dike construction.
- 37.0
(23.1) Note the old pilings from a former salt-loading railroad on the right. Note that the Salicornia-dominated salt marshes have given way to blue-green algal flats as we move further leeward on the lagoon shoreline. The associated sediments include more aragonite and less siliceous sand further landward from the lagoon. Some recrystallization to dolomite may be occurring within the sediment (results from X-ray diffraction analyses).
- 39.0
(24.4) Former pumping station.
- 41.0
(25.6) "English Bridge" on the right (a former salt-loading facility). Cores through the sediments here show alternating layers of dead sea grass (Zostera), blue-green algal mats, quartz sand, gypsum, carbonates, and clay. Note the association of lensoid gypsum crystals (sand to cm size, not cemented) with algal layers. Also note the lensoid gypsum crystals in the dike walls. This locality is not accessible during or following high tides. If not, continue to mile 27.7. Section A (Figs. 2-8), sampled by Holser before construction of the concentrators, started where the "Bridge" crosses the shore, intersected the route of the field trip at this stop, and continued to the landward end of the small salina to the east.
- STOP 3
- 42.4
(26.5) Concentrator 4 on the left. Large (up to several decimeters), laminated (stromatolitic but not lithified) blocks occur near the edge of the concentrator (annual range of density 1.036-1.066). The species are basically the same as those in Area 5. Due to competition by other algae and grazing by fish and invertebrates, algal mats (stromatolites) do not occur in the ponds of lower concentration.
- (possible additional stop)

- 44.3
(27.7) About 100-200 m before the gate to pump station 3A, a finger of the lagoon reaches the road. Within a short walking distance, middle intertidal blue-green algal mats (Lyngbya aestuarii) grade into upper intertidal Calothrix crustacea mats, above which are evaporite flats. Sections through these sediments show stratigraphy similar to those at the English Bridge (mile 25.6). However, at this locality a rather impermeable layer of coquina lies at about 0.5 m depth.
- 44.5
(27.8) Gate to the present pumping station 3A. Turn to the left and cross the natural island in the former tidal flats. Note the former beach fauna and Indian midden remains (obsidian chips, shells, bones).
- 44.8-
46.4
(28.0-
29.0) Sediments on the shore of the former island are covered with shells. Just below the surface is a layer of coffee-colored carbonate-siliceous sands. Between 10-20 cm below the surface is a layer of gypsum, some is nodular, like powder sugar, and white- to buff-colored. Below the gypsum is about 30 cm of light colored carbonate-siliceous sands with shells.
- STOP 4
- 53.6
(33.5) Junction of concentrators 9, 2, 3, and 4 (center of former "Salina Grande"). Turn right and follow sign to Bahía Tortugas and Asunción. Holser (1966) found polyhalite also in this section of the original sabkha (Fig. 2, Section B).
- 56.5
(35.3) At the end of the dike is a 3-way junction. Turn right and follow the road closest to the shore of the evaporator.
- 64.8
(40.5) Gate to the pump station. Turn left and follow the road closest to the lagoon.
- 66.6
(41.6) Junction to a fisherman's camp on the shore at the right. Take the left and continue following the shore of the lagoon. The track may be very poor in places.
- 74.2
(46.4) This marks the beginning of a very long (several km), high intertidal-supratidal flat. Sediment cores in this area have a mixture of gypsum, carbonate, and siliceous sand. A gypsum layer is especially noticeable below buried layers of sea grass. Pierre and Ortlieb found dolomite in this area (Fig. 2, Section E).
- STOP 5
- 77.1
(48.2) Just before the fisherman's camp at Cholla Beach. The surface carbonate muds are immediately underlain by gypsum. Abruptly at ca. 20 cm depth is a rust-colored layer of lensoid gypsum crystals (about 2 mm diameter crystals). This layer is about 20 cm thick and is underlain by carbonate sediments of the same color.
- STOP 6

- 77.8
(48.6) Cholla Beach. Shore sediments are siliceous and carbonate sands and muds with gigantic windrows of sea grass washed in from the lagoon. Note the windrows of foraminifera on the beach below the sea grass.
- STOP 7
- 80.5
(50.3) Continue following the track inland and across the flats. Do not leave the track. A large natural salt flat can be seen on the left (east).
- 81.1
(50.7) Junction with another track and a broken road sign to Guerrero Negro. In front is a large natural salt flat. The salt is underlain only by coffee-colored carbonate-siliceous sand and occasional gypsum lenses.
- STOP 8
- 105.8
(66.1) Return to junction at 56.5 km. At the junction, continue straight (southeast) along the shore of concentrator 2.
- 110.2
(68.9) Cross the canal. The canal was to be used to flood another salt flat.
- 112.2
(70.1) Intersection with a dike. Turn left and follow the dike.
- STOP 9
- 115.2
(72.0) End of the dike. At this point follow left to locate the site of Pierre and Ortlieb's sampling locality, on a part of the sabkha separated from concentrator 3 by a short dike (Fig. 2, Section C). Return to intersection and continue south (distance not logged).
- STOP 10
- 115.5
(72.2) Crossroad (look for marker stakes). Turn right and follow the track at the edge of the salt flats 12.8 km (8.0 mi).
- 120.3
(73.4) Most of this extensive salt flat is covered by gypsum sand with some quartz and carbonate sand abruptly underlain at about 10 cm depth by sand-size, lenticular crystals of gypsum. At about 30 cm depth blocky gypsum begins. In some places below the blocky gypsum lies a layer of siliceous-carbonate sand (ca. 10 cm) which is again underlain by gypsum. Some of the blocky gypsum, which is generally a dusky color, contains "cores" (small local seams) of uncemented, white, sand-size crystals.
- STOP 11
- 123.7
(77.3) The road climbs about 3 or 4 meters on a low dune. Just ahead the road branches. Keep to the right, staying near the salt flat.

- 124.3
(77.7)
(possible additional stop) The road returns to the edge of the salt flat. Notice the sediment here is covered with pebbles (predominantly basalt). Sediment cores yield a mixed gypsum-carbonate mud. At about 50 cm lies a fairly impermeable carbonate layer. Within the top 50 cm are occasional rounded cobbles of diverse lithologies.
- 126.2
(78.9)
STOP 12 Large gypsum roses occur in the road cut. They appear to be recrystallization products of the large, blocky gypsum crystals that outcrop here. Note the erratic basalt boulders on the salina floor. Halite occurs only as a thin veneer. These are the furthest inland salt flats of the Laguna Ojo de Liebre sabkha.
- 128.3
(80.2)
STOP 13 Continue on the track around the base of the large mesa. Note the magnesite along the road cut. At the base of the head of the mesa is an old sandstone reef similar to one found on Isla de Piedra in the lagoon. Climb the mesa for a good viewpoint. From here one can walk across the salina floor to examine more large basalt boulders. Their origin and mode of movement poses a curious geologic question. The sabkha near here was sampled by Pierre and Ortlieb (Fig. 2, Section D).
- 147.4
(92.1)
STOP 14 Return to the canal at 110.2 km (68.9 mi). Turn left (south) 1 km (0.6 mi) up the grade following the canal. Note the sediment cross section in the canal walls as the road rises up the mesa. A viewpoint of a large salina is at the top. Sediments underlying the salt are the same as those in the salt pan near Cholla Beach.
- 209.3
(130.8) Return to Guerrero Negro.

GEOLOGY OF COASTAL SONORA BETWEEN PUERTO LOBOS
AND BAHIA KINO

FIELD TRIP No. 2 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 22-24

TRIP LEADERS:

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MEXICO.

ROADLOG FOR COASTAL SONORA FIELD TRIP--SUNDAY, MARCH 22nd

Time	Increment in Kilometers	
7:00AM	00	Depart from Hotel Valle Grande, Hermosillo.
9:00AM	169	North from Hermosillo to Santa Ana (Figure 1). The hills west of the highway are largely unfoliated granitic rocks of late Cretaceous age. Host rocks show only minor contact metamorphism. Near Santa Ana the host terrane is highly fossiliferous Albian-Aptian limestone in contact with older Precambrian metamorphic rocks (Salas, 1968).
STOP 1 10:15AM	96	Pitiquito. Between Santa Ana and Altar we cross presumably Jurassic volcanoclastic rocks. Near Altar we cross older Precambrian metamorphic rocks. At Pitiquito the rocks are Late Precambrian carbonates intruded by granitic rock of Cretaceous age (Longoria and Perez, 1979). Fill vans with gasoline; get out the coffee jug for mid-morning wakeup.
11:30AM	40	Rancho Bamori We drive south from the old town of Pitiquito, immediately crossing the Rio Asuncion. The prominent hills to the east of the road are the Cerro Pitiquito, Cerro Chino, and Cerro Rajon. These have recently been described by Longoria and others (1978) and Longoria and Perez (1979). The Cerro Pitiquitos consists of Late Precambrian quartzite and carbonate rock. The Cerro Chino consists largely of Jurassic andesite-rhyolite flows and tuffs, with older Precambrian granodiorite exposed near the southern end. The Cerro Rajon consists of Late Precambrian quartzite and carbonate rock overlain by Jurassic sandstones. The Cerro Berruga, just north of Bamori is an area of older Precambrian rocks investigated by Damon and Mauger (1962) and Anderson and Silver (1971).
	9	Enter the hills southwest of Bamori, following the pass between Cerro San Clemente and Cerro Calaveras. The quartzite and carbonate rock to the north of the road is the lowest part of the Late Precambrian sequence, the rocks south of the road are Lower Cambrian (Eells, 1972).
	5.3	Faint tracks lead to the left.
STOP 2 11:45AM	1.0	Reach a fence; time for lunch. Here we will hike about two kilometers up the canyon to view the Cambrian and Late Precambrian section (Figure 2). Our traverse is essentially

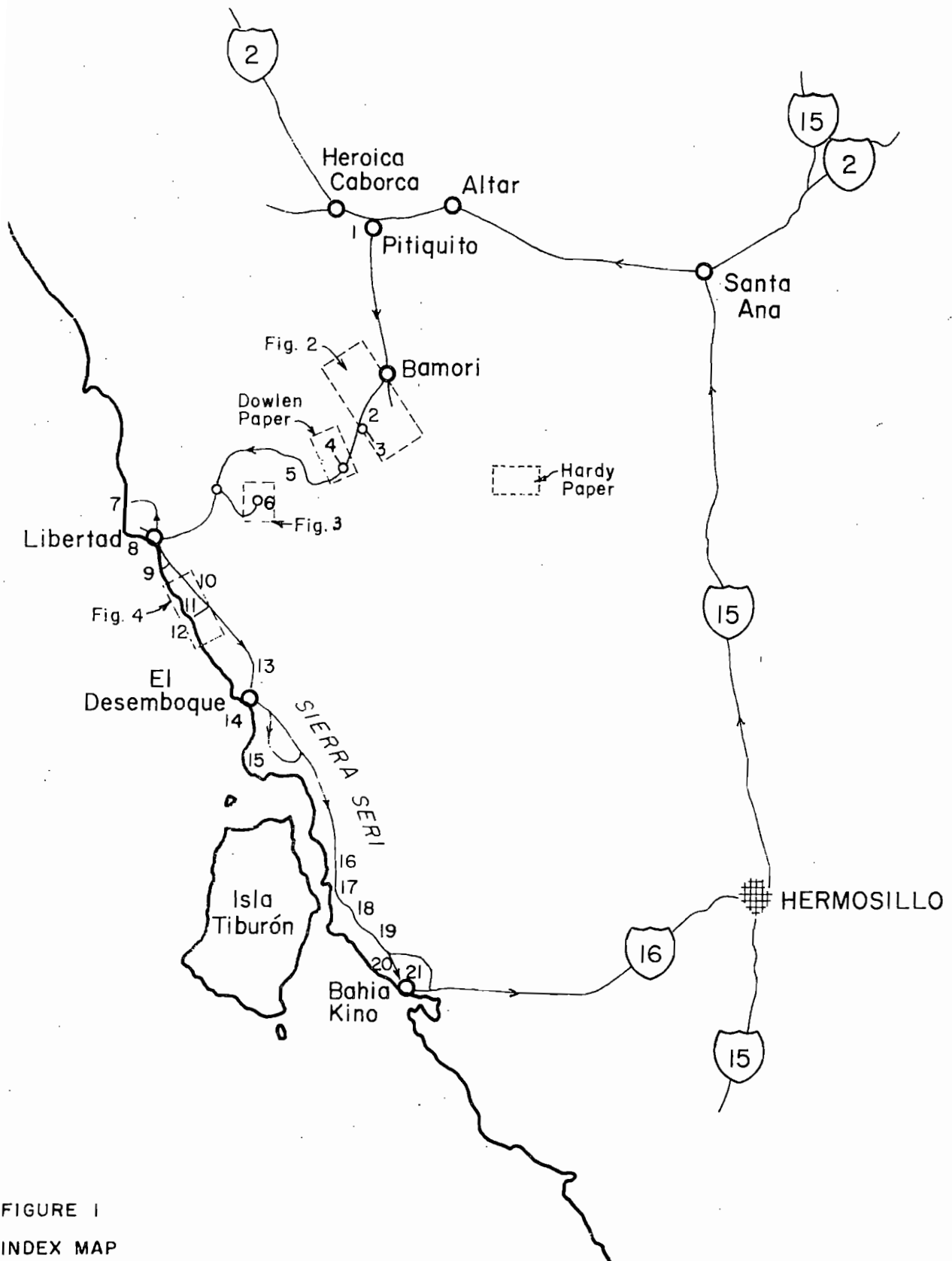
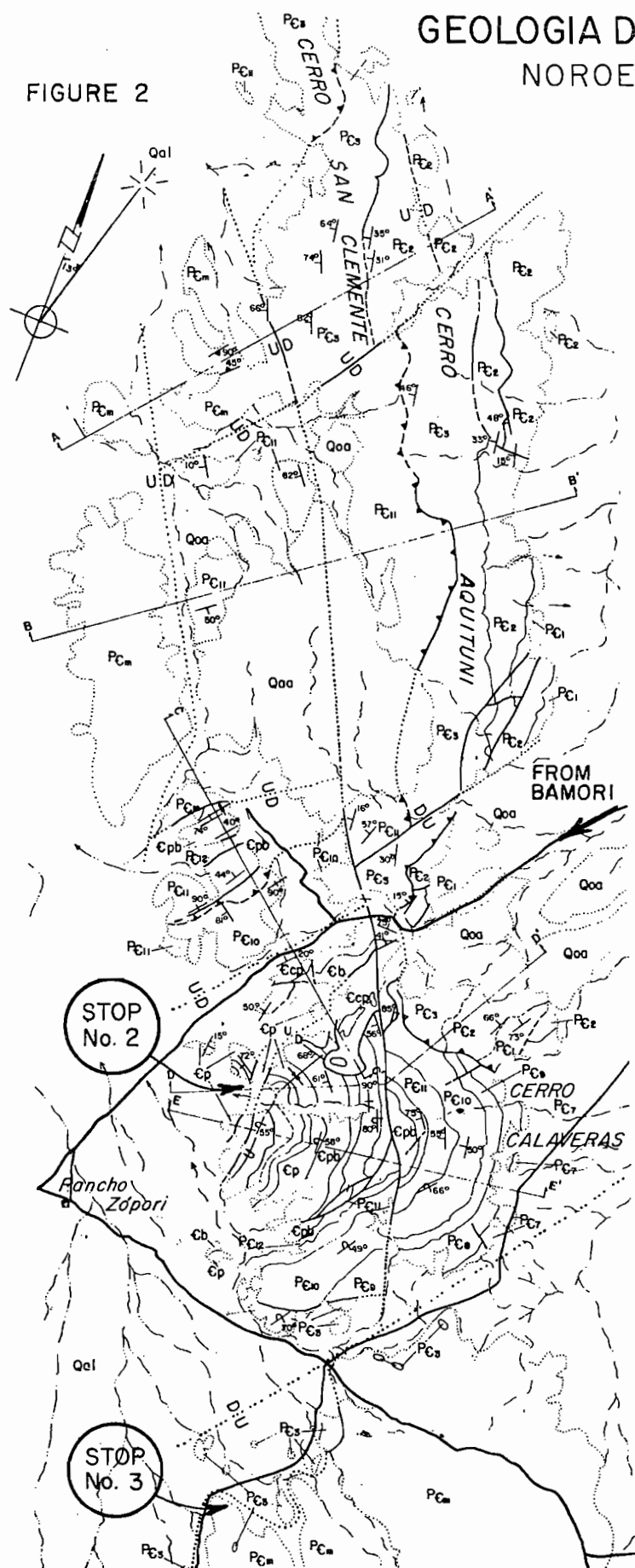


FIGURE 1
INDEX MAP
SONORA COAST FIELD TRIP

GEOLOGIA DE LA SIERRA DE LA BERRUGA NOROESTE DE SONORA, MEXICO

Por John L. Eells. (1972)

FIGURE 2



Qal

Aluvión

Qaa

Aluvión Antiguo

Ql

Depósitos Lacustres

T1

Riolita

M10

Brecha de Bloques

E1

FORMACION TREN
Dolomita gris. Espesor > 100 m.

Ea

FORMACION ARROJOS
Lutita y Caliza. Espesor ≈ 300 m

Ecp

FORMACION CERRO PRIETO
Caliza masiva. Espesor 88 m.

Eb

FORMACION BUELNA
Caliza, lutita y cuarcita. Espesor 68 a 104 m.

Ep

CUARCITA-PROVEEDORA
Cuarcita vitrea. Espesor 224 m.

Epb

FORMACION PUERTO BLANCO
Lutita, caliza y cuarcita. Espesor 456 m.

PE12

UNIDAD 12
Conglomerado volcanoclastico
y arenisco.
Espesor ≈ 90 m.

PE11

UNIDAD 11
Dolomita cafe y cuarcita
Espesor ≈ 250 m.

PE10

UNIDAD 10
Dolomita gris masiva
Espesor ≈ 490 m.

PE9

UNIDAD 9
Dolomita mateada
Espesor ≈ 140 m.

PE8

UNIDAD 8
Conglomerado de fragmentos finos
Cuarcita y lutita
Espesor ≈ 280 m.

PE7

UNIDAD 7
Caliza blanca y dolomita
Espesor > 150 m

PE6

UNIDAD 6
Dolomita gris
Espesor > 450 m

PE5

UNIDAD 5
Cuarcita vitrea
Espesor 184 m.

PE4

UNIDAD 4
Cuarcita y lutita
Espesor 139 m.

PE3

UNIDAD 3
Dolomita gris que contiene material
algaceo y aloquímicos
Espesor > 1400 m.

PE2

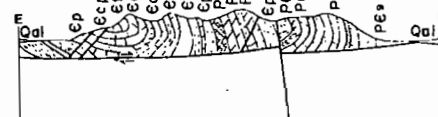
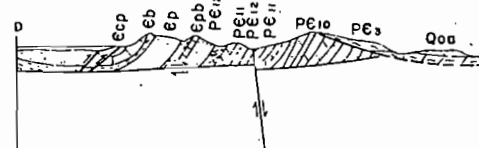
UNIDAD 2
Caliza, lutita y cuarcita. Espesor 68 a 104 m. Dolomita estramatolítica
Espesor ≈ 525 m.

PE1

UNIDAD 1
Cuarcita cafe. Espesor > 30 m

PE0

ROCAS
METAMORFICAS



cross-section DD' down through the Lower Cambrian to the conformable Precambrian contact. Of particular interest is unit 12 the basic volcanic unit at the top of the Late Precambrian (Eells, 1972).

2:15PM

depart

1.0 Back to main road, continue south.

3.5 Rancho Zópari, turn left (to east).

5.4 To road junction, turn right (south) by ruins of a building.

STOP 3 2:45PM

3.3 To the point of the hill.
Mylonitized older Precambrian granite, with inclination at low angle to the east.

3:15PM

depart

8.8 Back to Rancho Zópari, continue west.

0.8 Back to main road, continue southwest.

3.6 Gate to the right of the road, sign says Puerto Libertad, take it.

8.1 Rancho Pozo Serna, turn right (take right hand fork immediately).

STOP 4 4:00PM

2.1 Rancho Tilin, white buildings and windmill: time for afternoon beer. See note in this guidebook by Robert Dowlen. The paleontology of the Jurassic rocks has been described by Beauvais and Stump (1976). This little valley is a graben with overturned Cambrian strata both to the west and east. The rocks we will look at are volcanoclastic Jurassic sandstones which have been preserved in the center of the graben.

This is where we will prepare supper and camp for the night.

MONDAY, MARCH 23rd

6:00AM

wake up

6:30AM

breakfast

7:15AM

depart

2.2 Back to the Libertad road at Rancho Pozo Serna, continue southwest. On both sides of the road are andesitic volcanic rocks believed to be of Mesozoic age.

5.8 Junction, follow main road to right.

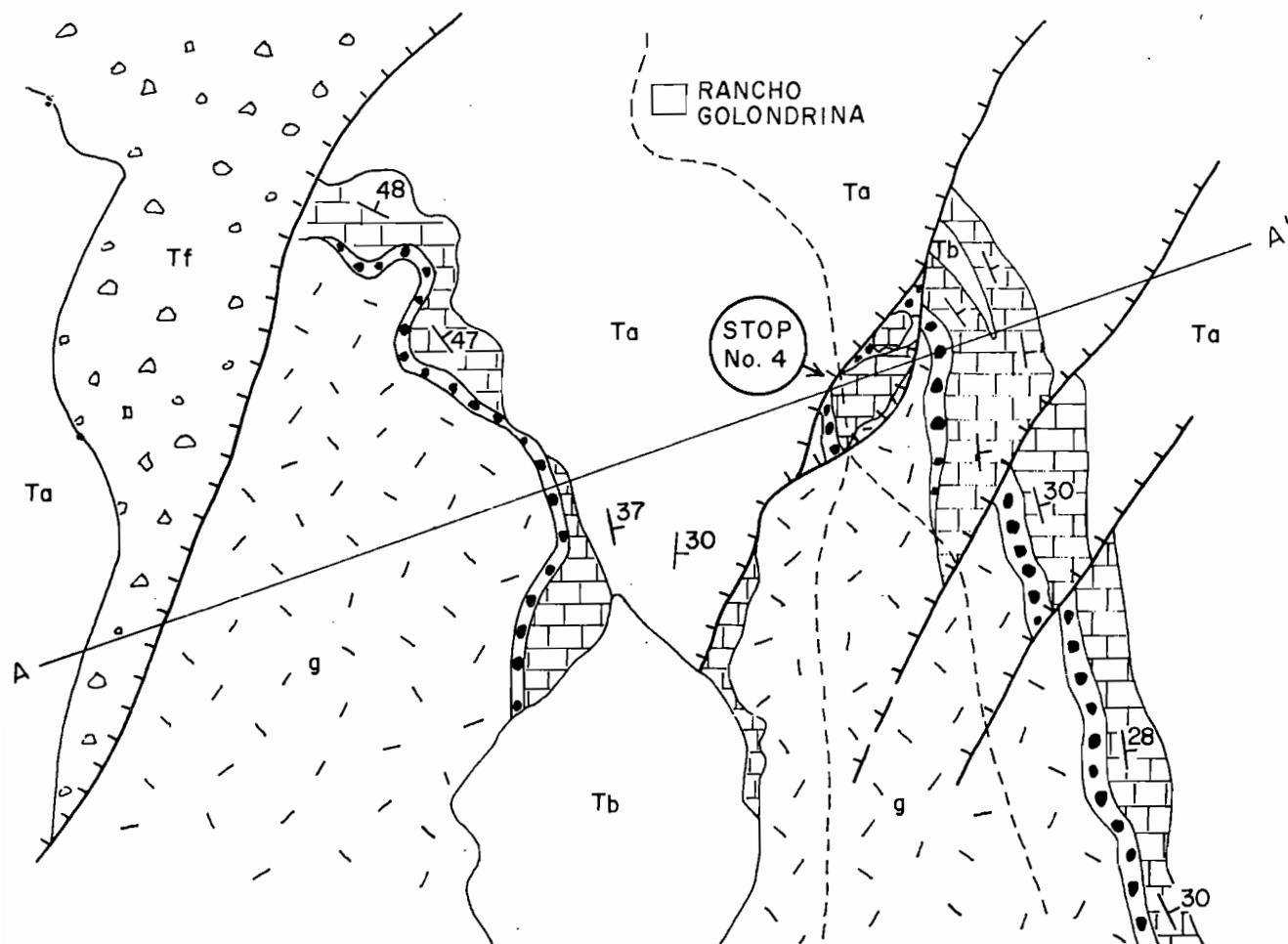
STOP 5 7:40AM

5.6 Granite outcrops.
We have just crossed a boundary between two apparently fundamentally different types of crust. To the east rocks with very little regional metamorphism rest on older Precambrian basement. To the west the pervasive basement rock is Cretaceous

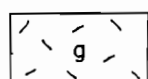
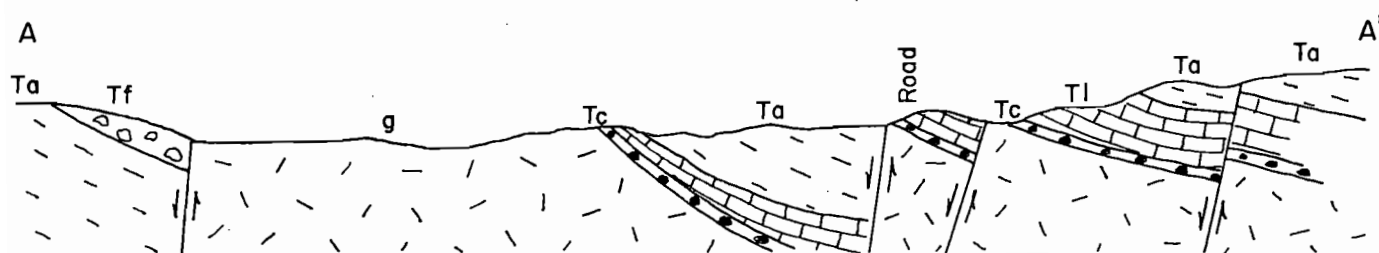
granitic rock, with all pre-Cretaceous rocks appearing to be "roof pendants." In 1977 Gastil and Krummenacher hypothesized a major fault (Figure 2) just to the northeast. Neogene (?) sedimentary rock exposed in the arroyos have been highly deformed. Any important faulting appears to pre-date the older alluvium. We now wonder if the boundary between two apparently different basement terranes is not a matter of erosional depth. We wonder if mylonite such as seen south-east of Rancho Zópari is part of a decoupling level, above which are rocks such as those seen in Cerro Calaveras and Cerro Pozo Serna, below which are Cretaceous granites such as seen here and regionally metamorphosed rocks such as will be seen the balance of the trip.

	approx. 12	Rancho El Carbón, follow main road to the south.
	approx. 6	Rancho El Americano, just west of here is a large porphyry copper deposit called La Fortuna. It is in volcanic rocks believed to be of Mesozoic age.
	approx. 3	Rancho Caracól, turn left (east) just south of the ranch buildings.
	2.4	Gate, should be unlocked.
	4.9	Rancho Pozo Nuevo.
	5.8	Take fork to the left (north).
STOP 6	9:15AM	3.5 Limestone outcrops, just short of Rancho Golondrina. Figure 3, this area was mapped by Wayne Mattox (1972). The small sketch map shown is modified from his map. He shows 55m of basal volcanoclastic redbeds resting on Cretaceous granite overlain by 150m plus of freshwater limestone and minor red beds. Fossils consist of gastropods and algal material, overlain by more than 230m of andesite tuffs and breccias, overlain by more than 100m of indurated fanglomerate. A rock from the andesite sequence near Rancho Golondrina yielded a K-Ar age of 12.7 ± 0.4 m.y. (Gastil and Krummenacher, 1977).
		This will also be our mid-morning coffee stop.
	10:30AM	depart
		16.6 Back to Rancho Caracól turnoff.
	11:00AM	28.3 Puerto Libertad, turn right (north).
		0.8 Opposite a group of trees the road crosses the new highway and proceeds northeast toward Santa Margarita (also known as Santa María).
		9.4 Take fork to left (west) toward the beach.

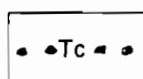
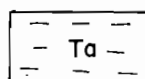
FIGURE 3 GEOLOGIC MAP OF RANCHO GOLONDRINA, SONORA



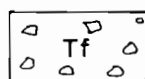
modified from W. A. Mattox, 1972



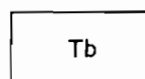
granite

basal
redbed
conglomeratelacustrine
limestone

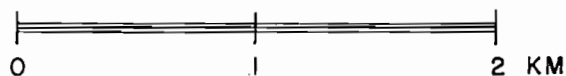
andesite



fanglomerate



basalt



STOP 7 11:20AM

3.1

Reach the beach, Santa Margarita.

Interbedded argillite and argillaceous sandstone with thin carbonate rock units, disharmoniously folded around northeast trending variably plunging fold axes. The axial plane strikes around N85°W, dipping 20°N. Greenschist facies regional metamorphism.

The low wave-cut terrace here has a maximum elevation of + 6m. In the first embayment south of the outlet various bench-like surfaces, veneered with pebbles of probable marine origin, are found at + 11m, + 16m, and + 20m. These accumulations are only a few centimeters thick and lack significant fossil shells. These are the highest marine terraces found up to now on the coast of central Sonora. They might have resulted from brief surges, as described by Hollin in several parts of the world.

This is our lunch stop.

12:30PM

depart

12.5

Return to Puerto Libertad, turn right toward entrance to Comisión Federal Electricidad generating plant.

STOP 8 12:50PM

Here we hope to visit the outcrops designated T3 on the map by Gastil and Krummenacher (1977). These steeply inclined conglomerate beds are the principal evidence for the graben mapped across Punta Bola.

On west edge of CFE enclosure is the Sangamon coquina (*Tagelus* and *Chione* sp. predominant) described by Stump (1975), Malpica and others (1978), Celis and Malpica, 1981, and Ortlieb and Malpica (1978, Fig. 8). The maximum elevation of marine deposits here is around 5m.

1:45PM

Depart, return to Puerto Libertad.

3.6

Turn off road toward the beach.

2.0

At fork turn right.

STOP 9 2:00PM

1.1

Reach the beach north of Punta Cirio. Conglomeratic sandstone with in situ echinoderms (Malpica and others, 1978, Fig. 5); presumed age is early or middle Pleistocene (or Pliocene?); maximum elevation is around +5m. These deposits are covered by old dunes (layers of calcified root molds and caliche), and locally by marine littoral sands, Sangamon (?).

2:30PM

depart

3.1

Back to the highway, continue southeast.

- STOP 10 2:40PM 17.1 Highway cuts through a small granite hill.
 This is the afternoon beer stop. Here you can see many fault surfaces. There are strike-slip faults striking N65°W, and a prominent fault striking north with slip surfaces dipping 35°E, and others. The movement planes exposed in this small hill are a hint of the movement surfaces that are presumably hidden beneath this broad alluviated pediment.
- 3:10PM Depart, continue south on highway.
- 4.4 Turn right onto dirt road leading west.
- 2.5 Volcanic hills.
 An andesite from unit T4x yielded a K-Ar age of $11.9 \pm .5$ m.y.
- STOP 11 3:25PM 1.3 Stop to look at the mylonitic gneiss exposed on the east side of the Falla Bacha (Figure 4). Note that the map by Gastil and Krummenacher (1974) erroneously labels the area as Mv (Mesozoic volcanic). The inclination of the mylonite is around 65° to the NE. Rotation of the Miocene volcanic rocks back to horizontal would leave the older structure inclined at a small angle to the east. Thus it is possible that we are looking at a rotated fragment of the same mylonitized older Precambrian granite observed southeast of Rancho Zópari.
- 4:25PM Depart, continue west.
- STOP 12 4:35PM approx. 4 Punta Cuevas.
 Wave-cut platform, fossiliferous littoral deposits at the base of the coastal cliff, between +3 and +6m.
 The fold of Miocene volcanic and clastic strata mapped in this area is atypical of the region, and presumably was caused by compression along the strike-slip Falla Bacha.

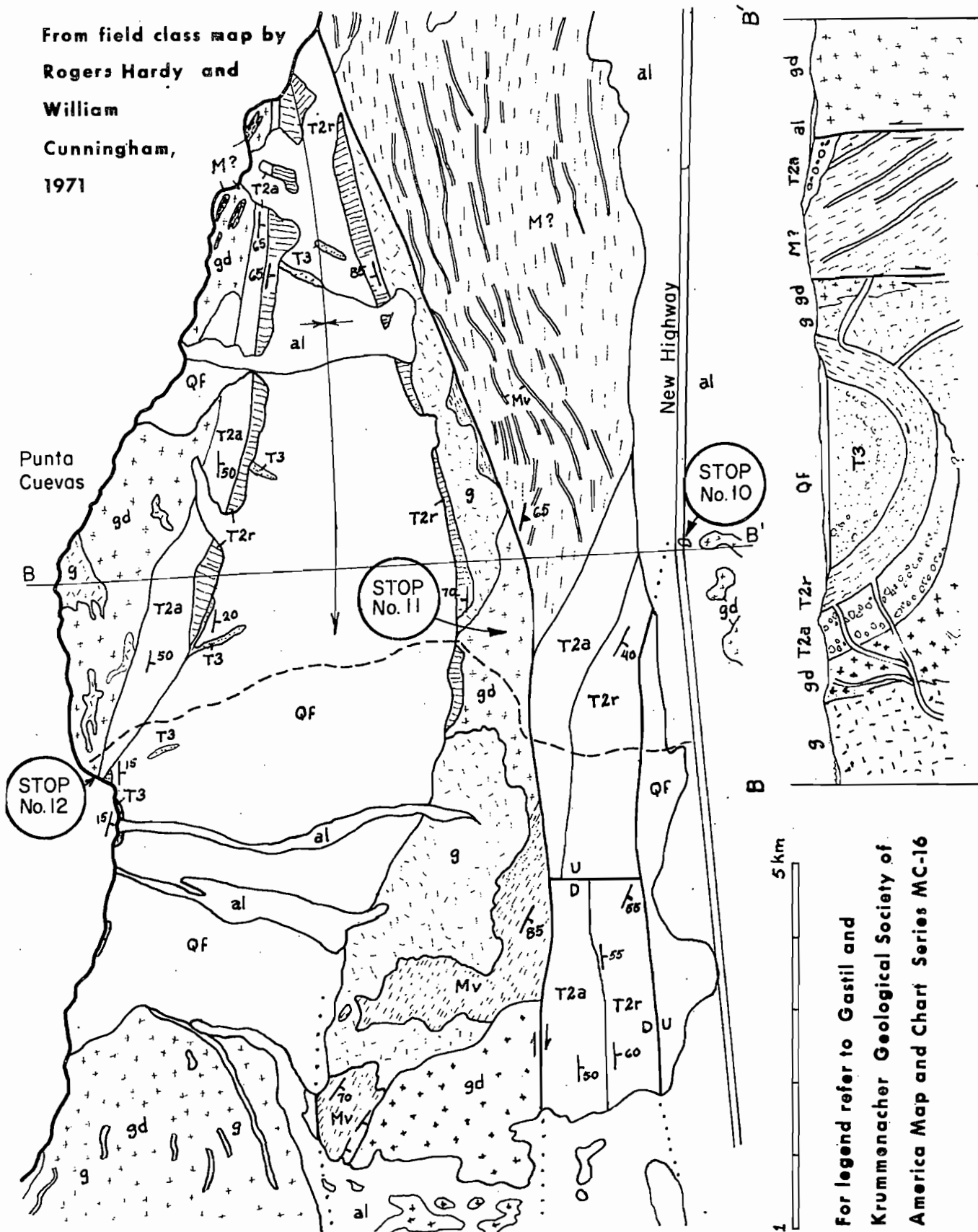
This is where we shall prepare supper and camp for the night.

TUESDAY, MARCH 24th

- 6:30AM wake up
- 7:00AM breakfast
- 7:45AM depart, return toward the highway
- approx. 12.2 back to the highway, continue south
- 8:15AM 16.9 Cerro Prieto basalt (K-Ar date 6.4 ± 1.9 m.y., Gastil and Krummenacher, 1977).
- 2.1 Road to east leads to Rancho Arroyo Seco.
- 1.8 Another road to east.
- 1.5 Rancho Pozo Coyote.
- 1.7 Cross Rio San Ignacio.

FIGURE 4 GEOLOGY OF THE PUNTA CUEVAS AREA, SONORA

From field class map by
Rogers Hardy and
William
Cunningham,
1971



- STOP 13 8:25AM 0.9 Massive quartzite.
This could be a fragment of the Cambrian-Late Precambrian terrane visited in the Sierra Calaveras.
- 8:40AM depart
- STOP 14 9:00AM 12.3 Seri Village of Desemboque Del Rio San Ignacio.
If you wish to barter for handicrafts we shall stop here for that purpose.
- 9:30AM depart
- 1.0 Road junction east of village, continue southeast on highway.
- 11.6 Take a fork to the right, road leads southwest toward the coast.
- 2.7 Turn onto a fork that leads east of south.
- 3.9 Take a left turn, the other road leads to Campo Dólar.
- STOP 15 10:00AM 1.0 "Laguna Tepoca": an area of playa morphology located between the piedmonts of Cerro Tepoca and Sierra Seri. Here sandy deposits with Chione sp., Protothaca sp., and Cerithium sp. have been dated by ^{14}C at about 30,000 years B.P. (but are probably older: Sangamon). The maximum elevation is $+3\text{m} \pm 2\text{m}$ (LeColle and Ortlieb, 1978; LeColle, 1979; Ortlieb and LeColle, in preparation; Richards, 1973; theses of B. Luna and R. Gonzalez, 1981).
This will be our morning coffee stop.
- 10:30AM depart
- 5.9 Turn to the left, away from the salt marsh.
- 10:45AM 2.3 From here we can see the Estero Sargento with its mangrove vegetation. For the vegetation see the thesis of Sherwin (1971), for the geomorphology Lancin (1979), and the sedimentology Nichols (1965).
- 0.3 Take the straight road toward the northeast.
- 1.8 Back to the highway.
- 3.7 Pass a junction on the right leading toward Estero Sargento.
- 19.8 Punta Viboras: pull off onto a small parking area.
- STOP 16 11:25AM This is an opportunity to look at the mangrove.
- 11:40AM depart

- STOP 17 12:00PM 11.6 This is a double stop. Up the hill on the left-hand side of the road are the metamorphosed banded cherts which have been correlated (by lithology) with rocks on Isla Turner, Isla Angel de la Guarda, and localities on the coast of Baja California.
- On the right-hand side of the road we will climb down to the beach where an infra-littoral coquina deposit and an intertidal sandy conglomerate are exposed in the sea cliff. The maximum elevation of the littoral material is +9m. The age is presumed to be Sangamon (this is an unusually high elevation for Sangamon material along this coast and may be disturbed by movement on the hypothesized fault: Gastil and Krummenacher, 1974).
- This is also our lunch stop.
- 1:10PM depart
- 1:17PM 4.5 Village of Punta Chueca: Seri
- STOP 18 1:20PM 3.0 Stop to look at the metavolcanic and metasedimentary rocks exposed along the left-hand side of the road. These include andesite, rhyolite, and both carbonate and clastic sedimentary rocks. Similar rocks are exposed immediately across the channel on Isla Tiburón. The age of these rocks is not known. Similar metavolcanic strata are not found along the eastern coast of northern Baja California.
- 1:45PM depart
- STOP 19 1:55PM 6.4 Stop briefly to look north and see the uplifted old conglomerate deposits.
- 2:15PM 8.1 Turn off of the highway to the right, road leads toward Punta Blanca.
- 1.1 Fork in road, keep left.
- STOP 20 2:25PM 2.3 Stop briefly for view. In the bay north of Punta Ignacio there is a Wisconsinian Bajada deposit overlying Sangamon marine strata. The former is several meters of light brown sandy gravel of sheet flood origin (Malpica and others, 1978; Ortlieb and Malpica, 1978, Figure 3).
- 2:35PM 2.5 Kino Bay Trading Post. Turn left and follow the Mar de Cortez Blvd. This is the "new town" of Bahia Kino, built on Holocene sand dunes.
- 2.8 Turn left onto Av. Punta Checa.
- 0.9 Turn left onto the "Periferico."
- 0.5 Turn right through hole in the fence.

STOP 21	2:45PM	2.0	Stop at fence and walk to steeply inclined Miocene clastic and volcanic strata. A dacite from this locality has been dated at 17.8 ± 0.8 m.y. (Gastil and Krummenacher, 1977). This is also our afternoon beer stop.
	3:35PM		depart
		3.4	Back to Mar de Cortez Boulevard.
	5:00PM	117	Hermosillo. We will be crossing a broad coastal plain of Pliocene-Quaternary alluviation. This alluviation has resulted in the partial burial of Siete Cerros and other hills. The gradual slope extends from sea level to Hermosillo at an elevation of 180 m.

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GEOLOGY OF THE CENTRAL SIERRA DE SANTA ROSA

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ABSTRACT

An allochthonous Lower Jurassic marine sedimentary and andesitic volcanic sequence at least 2,360 m thick is exposed in the Sierra de Santa Rosa in northwest Sonora, Mexico. The Jurassic sequence is tectonically emplaced over older Precambrian regionally metamorphosed rocks and autochthonous younger Precambrian sedimentary rocks.

The Lower Jurassic Sierra de Santa Rosa Formation (new name) consists of 1,460 m of alternating andesitic lithic wacke, calcareous arkose, sandy biomicrite and argillite. The formation is interpreted to have been deposited in a shallow, nearshore marine environment with tidal shelf, lagoon and barrier bar facies. The source area for the formation is possibly an island arc proposed to have existed along the western edge of Sonora. The Lower (?) Jurassic Cerro San Luis Formation (new name) lies conformably on the Sierra de Santa Rosa Formation, is at least 900 m thick, and consists of 0 to 70 m of conglomerate containing quartz arenite and dolomite clasts, grading upward into approximately 860 m of andesitic agglomerate, lapilli tuff, lava flows and interbedded channel conglomerate and arenite. The Cerro San Luis Formation is interpreted to have been deposited subaerially. The tectonic events necessary for the deposition of the conglomerate at the base of the formation and the andesite volcanism could also be related to the presence of an island arc. The Precambrian and Jurassic rocks are intruded by a Late Mesozoic or Early Tertiary rhyolite porphyry.

The Jurassic sequence occurs as a set of north-south striking allochthons exhibiting a low-angle undulating fault contact with the underlying Precambrian rocks. The strata within the allochthons strike north-south, dip steeply to the west, and are repeated as imbricate sets along reverse bedding-plane faults. Sevier-age imbricate thrust faulting within the Jurassic sequence, and Laramide age gravity sliding of these allochthons might explain the observed geometry. Gently dipping beds of the Miocene (?) Baucarit Formation indicate that the central portion of the range was tilted to the west, but normal faulting has not altered the configuration of the allochthons in a major way.

INTRODUCTION

Physiography and Location

In the Sierra de Santa Rosa area, 70 km southeast of Caborca (Figures #1 and 2), a sequence of Lower Jurassic shallow water marine sedimentary and volcanic rocks crops out as a set of imbricate allochthons. The sequence is also exposed in the adjacent Cerro San Luis, and in the low lying foothills, including the Cerro Garambullo, to the west of the Sierra de Santa Rosa, Cerro San Luis and Sierra Yaqui. Throughout this paper, the location of this sequence will collectively be referred to as the Sierra de Santa Rosa.

This linear range of rolling hills is 3 to 10 km wide, 35 km long, and exhibits 1,000 m of relief. The eastern front of the range is bordered by a flat pediment valley, 300 m higher than the valley to the west. The west front of the range is less well defined, breaking up into a series of foothills and ridges toward the valley. The mapped area is bounded by $30^{\circ} 12' N$ on the north, $30^{\circ} 3' N$ on the south, $111^{\circ} 37' W$ on the east and $111^{\circ} 50' W$ on the west, comprising 357 square kilometers.

Previous Work

Pioneering geologic expeditions through the State of Sonora early in this century determined many of the stratigraphic ages and field relationships observed in the Sierra de Santa Rosa. Most notable are by Jose Aguilera (1896, in Dumble, 1900), E.T. Dumble (1900), Teodoro Flores (1929), Carlos Burckhardt (1930), and an excellent regional synthesis by R.E. King (1939).

Since World War II, correlative rocks and structures similar to those in the Sierra de Santa Rosa have been described in the immediate area. Cooper and Arellano (1946) described the younger Precambrian and Cambrian strata in the Caborca area. White and Guiza (1948) described an Upper Triassic and Lower Jurassic section in El Antimonio mining district, 103 km west of the Sierra de Santa Rosa. This locality is the closest measured section of correlative rocks to the Sierra de Santa Rosa, and the depositional base is exposed. Gloria Alencaster De Cserna (1961) studied the paleontology and stratigraphy of the Upper Triassic Barranca Formation and elevated it to group status. Damon and others (1962) and Anderson and Silver (1971) have mapped and dated older Precambrian metamorphic rocks between Caborca and Rancho Bamori. Guillermo Salas, Jr. (1968) mapped similar older Precambrian metamorphic and Mesozoic volcanic rocks in the Santa Ana region. J.L. Eells mapped the younger Precambrian and Cambrian geology of the Sierra de La Berruga, (Anderson, Eells, and Silver, 1978) immediately west of the Sierra de Santa Rosa. T. E. Stump and R. Dowlen (Beauvais and Stump, 1976) mapped Upper Jurassic strata at Cerro Pozo Cerna. R.H. Merriam (1972) published a paper on the reconnaissance geology of the Sonoyta 1:250,000 quadrangle.

Active mining operations in the 1920's attracted four geologists to the Sierra de Santa Rosa itself. They published only very generalized stratigraphic observations and collected fossils. W.T. Keller (1929), with F.E. Wellings, made an east to west reconnaissance traverse through the central area

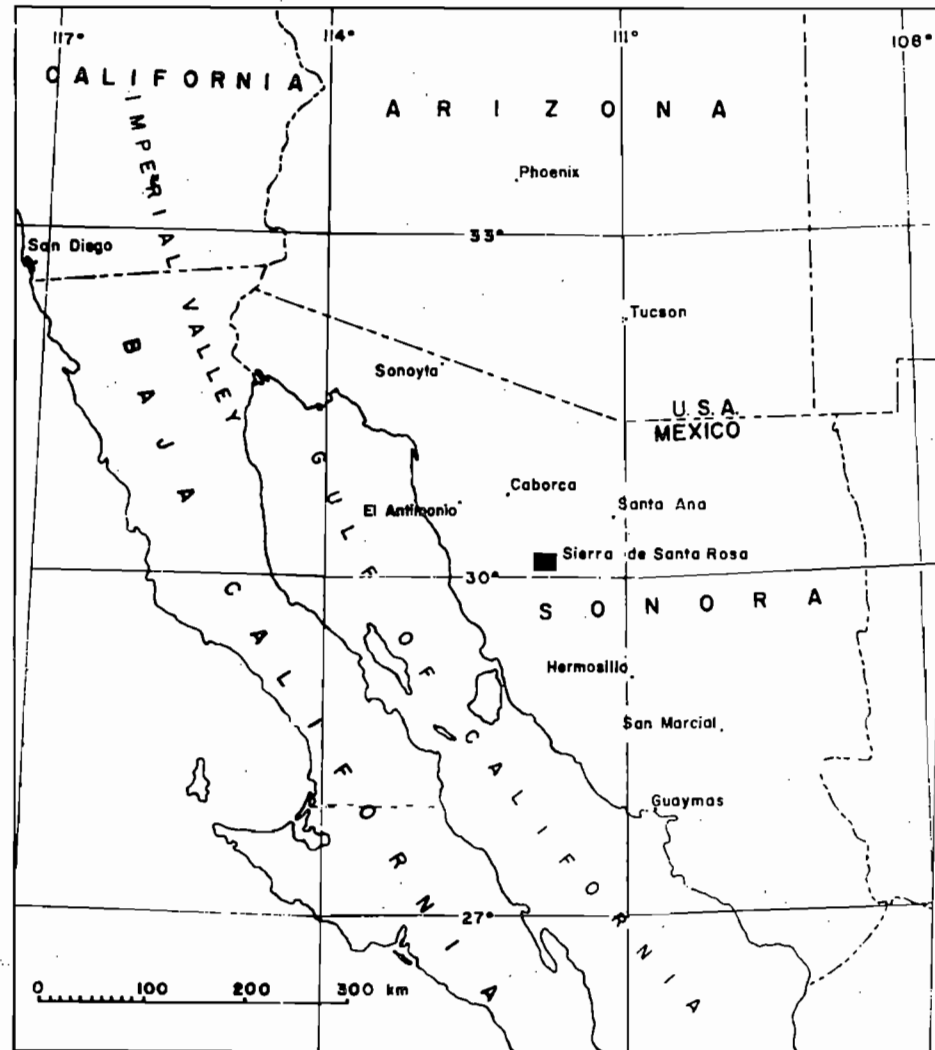


FIGURE 1 - Index Map of the state of Sonora.

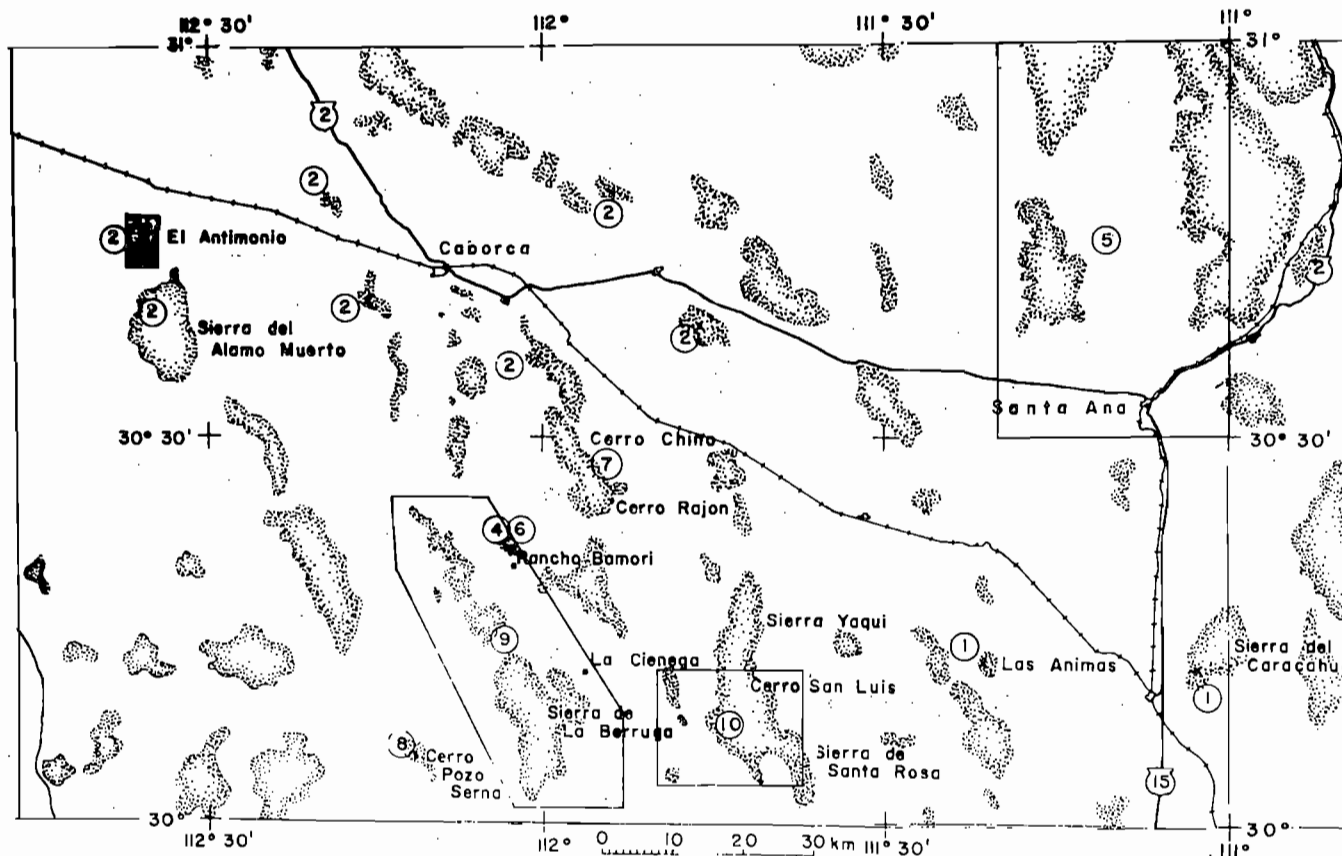


FIGURE 2 - Map of the Caborca region showing previous work and known Lower Jurassic outcrop locations: 1) Flores (1929), 2) Cooper and Arellano (1946), 3) White and Silvers (1971), 4) Damon and others (1962), 5) Salas (1968), 6) Anderson and Silvers (1971), 7) Merriam and Chandler (1971), 8) Beauvais and Stump (1976), 9) Anderson and others (1978), and 10) this paper. Jurassic strata crop out at 1, 3, 8, and 10.

of the range. They described four very general stratigraphic units totaling an estimated 2,350 m in thickness. The sequence was observed to be metamorphosed and highly fractured, but no major low-angle faulting was recognized. Keller and Wellings gave their fossil collection to Erich Jaworski (1929), who published a faunal description and assigned an age of Lias (Early Jurassic) to the sequence. Charles L. Baker also visited the area and supplied stratigraphic and paleontologic information to Charles Schuchert (1935, p. 145) and Burckhardt (1930, p. 23). Burckhardt reprinted the stratigraphic description of Keller and Baker, reviewed Keller's faunal collection and concurred with Jaworski's Lias age. Flores (1929, p. 150) included a brief summary of the major rock types in the vicinity of the mine and collected fossils. Baker's, Burckhardt's and Flores' comments support, but add little to, the observations and conclusions of Keller and Jaworski. R.E. King (1939) used Keller's lithologic information and Jaworski's age conclusions as evidence assigning the sequence to the widespread Barranca Formation.

Base Map Construction

The most detailed topographic map available of the region is the Cananea 2° quadrangle with a scale of 1:250,000. The areal geology of the area, therefore, was plotted on a drainage map constructed planimetrically from air photos with a nominal scale of 1:62,500 (Figure #3).

PRE-JURASSIC ROCKS

Older Precambrian Metamorphic Rocks (pCM)

The oldest rocks in the Sierra de Santa Rosa are a complex assemblage of schist, gneiss, metaquartzite and mildly gneissose metaplutonic rocks. This assemblage forms a structural basement upon which younger sedimentary and volcanic rocks have been tectonically emplaced, and is exposed throughout the mapped area. The assemblage is similar petrologically, and exhibits the same field relationships, as older Precambrian rocks studied in the region by Anderson and Silver (1971; Anderson, personal communication) and Damon and others (1962).

The basement of the Cerro Colorado and western Cerro San Luis is predominately a medium to coarse-grained, chloritized quartz monzonite. The basement north and northwest of the mine at Cajon de las Amarillas consists of a quartz-feldspathic gneiss porphyry. The porphyritic texture is due to large potassium feldspars. The basement terrane east of the Mina de Cajon de las Amarillas is an assemblage of dark gray, highly fractured metaquartzites and black hornfels. The basement in the southwest portion of the mapped area, in the vicinity of Rancho Tierras Blancas and Rancho San Carlos, is dominated by a medium-grained, gneissose pink granite with lesser amounts of gray muscovite schist. East of Rancho San Carlos and east of the crest of the Sierra de Santa Rosa, the basement is predominately a buff, medium-grained, foliated muscovite quartzite.

Younger Precambrian Sedimentary Rocks (pCs)

Unconformably overlying the metamorphic complex is at least 900 m of dolomite with lesser thicknesses of quartz arenite and shale. This sequence exhibits petrologic and stratigraphic relationships similar to the younger Precambrian rocks studied by Cooper and Arellano (1946) in the Caborca area, and Eells (Anderson, Eells, and Silver, 1978) in the Sierra de La Berruga. No formal stratigraphic name has yet been assigned to this sequence, although its correlation to the latest Precambrian rocks of the Nopah Range in southeastern California has been suggested (Livingston and Damon, 1968).

The sequence within the mapped area is best exposed at Cerro Colorado and in the northwest portion of the area and on the western slope of the Cerro San Luis. In both localities the sequence strikes N 10° W and dips steeply to the west. The sequence is unmetamorphosed and much less fractured than the allochthonous Jurassic sequence. The nonconformity with the older Precambrian metamorphic rocks is exposed along the eastern side of Cerro Colorado, and 1 km due east of Rancho San Luis in a stream canyon. The nonconformity shows very little relief, and the weathered zone extends only a few meters into the older metamorphic complex. The basal bed is a pink, medium to coarse-grained, subrounded arkose averaging 3 m in thickness. The rest of the section is dominated by a light grey, mottled dolomite. The dolomite contains planar or cross-laminae sandstone beds and crenulated structures resembling algal mats. The dolomite is also massive, and intraclastic beds occur. Light pinkish gray, fine to coarse-grained, well rounded quartz arenite comprise about 10 percent of the section, in .5 to 3 m beds. In the Cerro Colorado a 40 m thick interval of reddish black laminated shale forms a series of low saddles in the resistant dolomite.

One small, highly fractured klippe composed of pinkish gray quartz arenite resting entirely on older Precambrian metamorphic rocks is exposed in the west central portion of the mapped area. This is interpreted to be the only allochthonous block of younger Precambrian strata in the Sierra de Santa Rosa region.

PETROLOGY, STRATIGRAPHY AND ORIGIN OF THE JURASSIC ROCKS

General Statement

The Jurassic strata of the Sierra de Santa Rosa consist of a continuous section of at least 2,360 m of sedimentary and volcanic rocks distributed on the Precambrian basement in an assemblage of imbricated allochthons (Figure #4). They comprise the central portion of the Sierra de Santa Rosa and Cerro San Luis, and the foothills to the west, covering an area 11 by 15 km. The complex is intruded by post-thrusting rhyolite dikes of Late Cretaceous or Early Tertiary age, and is overlain by the moderately inclined continental Tertiary Baucarit Formation and flat lying Quaternary gravels.

The lower 1,460 m of the section is alternating volcanic and calcareous lithic wacke, calcareous arkose, sandy biomicrite and argillite with lesser thicknesses of volcanic conglomerate, shale and sandy oosparite. This portion of the section is interpreted to have been deposited in a shallow, nearshore marine environment with barrier bar, lagoon and tidal shelf facies. This marine sequence is conformably overlain by a lensing quartz arenite and dolomite cobble conglomerate up to 70 m thick, which

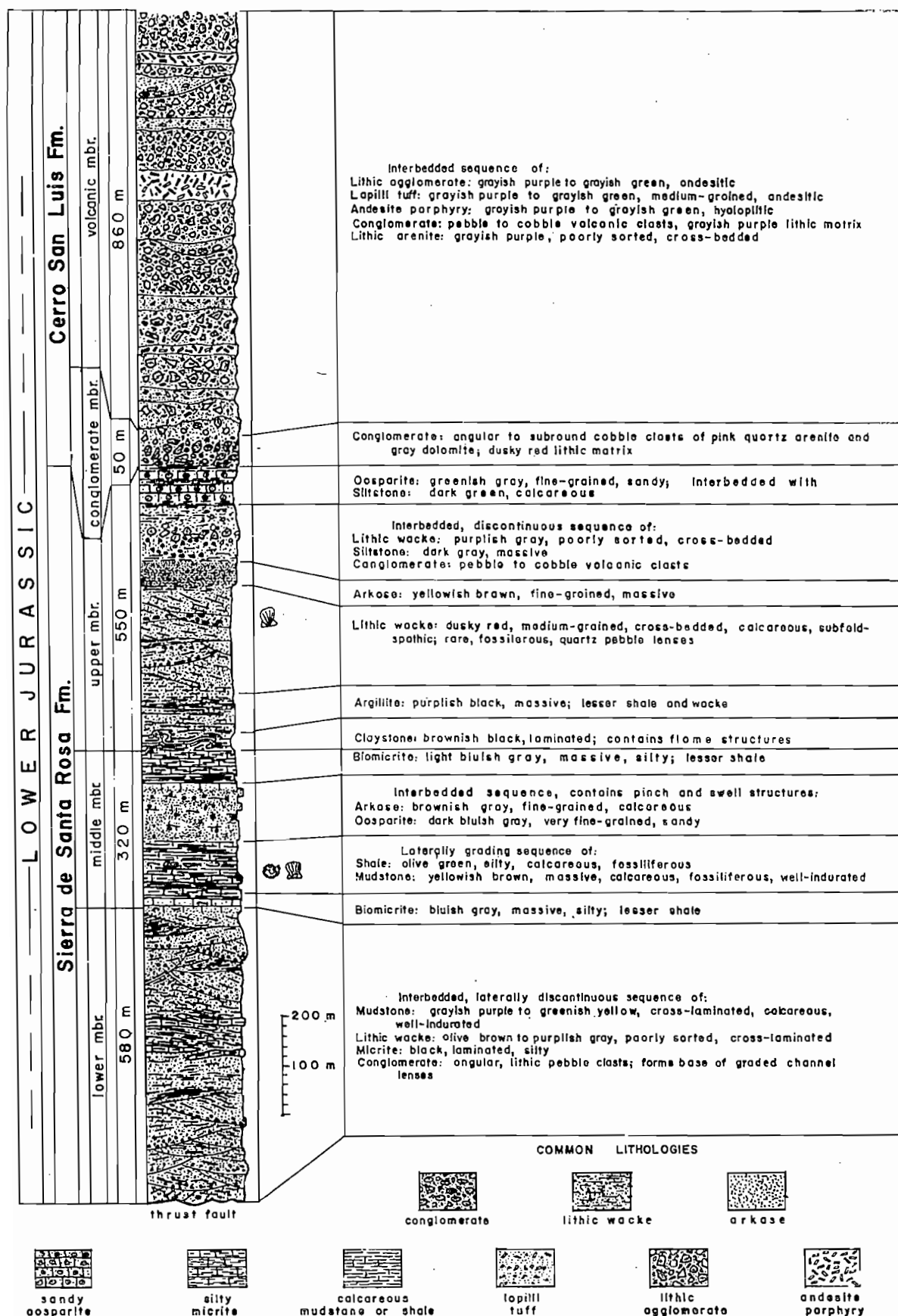


FIGURE 4 - Composite columnar section of the Sierra de Santa Rosa and Cerro San Luis Formations. See text for type localities.

grades upward into a sequence of andesitic agglomerate, lapilli tuff, lava flows and interbedded volcanic conglomerate and wacke. The thickest exposed section of the volcanic sequence is 860 m. The absence of any fossils or fossil fragments in the quartz arenite-dolomite conglomerate and the interbedded conglomerate of the volcanic sequence infers subareal deposition, although a nearshore marine environment is possible.

Hydrothermal alternation occurs throughout the Jurassic rock mass, but is concentrated in the volcanic rocks along contacts with the rhyolite intrusions, along minor fractures, and along the major low-angle faults. These zones are commonly fractured and show epithermal mineralization (Wisser, 1965). The entire Sierra de Santa Rosa is laced with diggings. Mina de Santa Rosa and Mina de Cajon de La Amarillas produced precious metal ore from the late 1800's to the 1940's.

Sierra de Santa Rosa Formation (Jsrl)

The new name Sierra de Santa Rosa Formation is here proposed for the Lower Jurassic and possibly Upper Jurassic marine sedimentary rocks exposed in the central Sierra de Santa Rosa, and in the low-lying hills west of the range.

R.E. King (1939, p. 1655) included the Lower Jurassic strata of the Sierra de Santa Rosa in the Barranca Formation of Dumble (1900, p. 139). King grouped all the Upper Triassic-Lower Jurassic continental and marine strata spread over thousands of square kilometers of Sonora into this one formation. King's reason in so doing was to delineate a distinct, regional cycle of sedimentation, the deposits of which are separated by unconformities from the underlying and overlying strata. G.A. De Cserna (1961) elevated the Barranca to group status, dividing it into the Upper Triassic brackish water marine Santa Clara Formation lying between the continental Arrayanes and Coyotes Formations. According to De Cserna, these formations are restricted to the "San Marcial Basin" east of Hermosillo, but at times were connected to a marine embayment to the northwest where sedimentation was continuous into the Jurassic. Therefore, the Sierra de Santa Rosa Formation is contained in the upper portion of the Barranca Group.

The formation is informally divided into three members based on the predominance of a particular lithologic type, lateral continuity of beds and sedimentary structures. Because of the structural complexity of the Sierra de Santa Rosa, measurement of a continuous type section through the entire formation is impossible. Therefore, each member has its own section location, picked to show good, continuous exposure, and give a representative indication of the internal characteristics (Figure #3).

Lower member (Jsrl). The lower member of the Sierra de Santa Rosa Formation is volcanic wacke and hard, calcareous mudstone, with lesser volcanic pebble conglomerate and sandy micrite. The member is distributed within several allochthons, primarily in the east-central Sierra de Santa Rosa. The thickest section is approximately 580 m, but the total thickness is unknown as the depositional base is not exposed. The rock types occur repeatedly throughout the section in laterally discontinuous, commonly cross-bedded lensoidal deposits, but no major facies change was noted in the member across structural boundaries. The type section was chosen to represent the relative abundances of the different lithologies, and does not show the stratigraphic position of any one lithology

throughout the area. It is located along the east-west ridge between the Cerro San Luis peak and the road from La Cienega to the ruins of the Mina de Cajon de Las Amarillas, starting 1 km west of the ruins.

Approximately 50 percent of the member is hard, massive, laminated and cross-laminated calcareous siltstone and mudstone occurring in 1 to 5 m lensoidal beds. They are highly fractured, commonly weathering into dark grayish purple to greenish yellow rectangular blocks from cobble to boulder size. The average composition is 50 percent subangular, coarse, quartz and feldspar silt, 35 percent authigenic and detrital clay, primarily sericite, and 15 percent authigenic calcite. These modes vary from 80 percent silt to thin lenses of claystone or marl with shell fragments. In at least three horizons, the mudstone grades into evenly laminated, silty black micrite beds, 2 to 3 m thick, which comprise about 5 percent of the member. They are the most laterally persistent beds in the member, traceable for over 1 km.

Forty percent of the member is well-indurated, cross-laminated, light olive brown, angular to subangular, fine to medium-grained lithic wacke. The laminae, .5 to .4 mm thick, are subtly defined by grain size and color variation, and form gentle, planar cross-laminations in prisms 1 to 10 m thick. The prisms commonly show channel grading from pebble to lutite beds. The wacke contain an average of 48 percent clay matrix, and 52 percent sand grains composed of polycrystalline quartz with sericite and monocrystalline quartz with vacuoles (37 percent), feldspars (30 percent) and aphanitic volcanic rock fragments with ferromagnesian minerals (33 percent). The clay matrix probably also contains some completely decomposed rock fragments and unstable mineral grains. Interbedded with the wackes, commonly at the bottom of a graded sequence in 10 cm thick lenses, are poorly sorted, angular pebble conglomerates.

Middle member (Jrsm). The middle member, averaging 320 m, is a relatively laterally persistent sequence of silty biomicrite, sandy, fossiliferous shale and arkose, well exposed in the east-central Sierra de Santa Rosa, and in the low hills between Cerro Garambullo and Rancho San Carlos. The type section is in a stream gully crossed by the dirt road through the range from Rancho Tierras Blancas to Rancho San Diego, 400 m north of the main shaft to the Mina de Santa Rosa. This is probably the principal fossil locality of Keller (Jaworski, 1929). The lithologic sequence described in the type section is present throughout the Sierra de Santa Rosa, although the relative proportion of the limestones is greater in the allochthon located in the southwestern low hills of the mapped area.

The bottom of the member is marked by the first occurrence of a light bluish gray, silty biomicrite. The micrite is commonly massive and contains 3 percent quartz silt and abundant bivalve shell fragments with geopetal spar. Interbedded in the micrite are 1 to 5 m beds of green, silty shale and rare sandy lenses with current-oriented bivalve fragments. This limestone-dominated assemblage ranges from 80 to 140 m in the thickness and is present at this stratigraphic position throughout the Sierra de Santa Rosa. At the type section, the limestone-dominated assemblage grades into a calcareous silty shale bed 20 to 80 m thick. In the allochthon located in the low lying hills southwest of the type section, this stratigraphic interval is occupied by a calcareous siltstone. The apparent facies gradation between the two is not readily observable in the area. The shales, similar to that interbedded with the biomicrite is pale olive green, contain 19 percent quartz silt, 33 percent calcite and are extremely fissile, weathering to unconsolidated, pebble size

blocks forming gentle slopes. The siltstone is moderate yellowish brown weathered and grayish black on fresh faces. It is massive, well indurated, highly fractured and weathers into cobble size rectangular blocks. The siltstone contains 70 percent subangular to round coarse silt, 30 percent authigenic spar and recrystallized fossil allochems and traces of sericite. The shale and siltstone contain the best preserved fossils in the formation. Well preserved specimens of Weyla (Weyla) mexicana Jaworski are abundant in the siltstone, along with other bivalves. Ammonite impressions are more common in the shale, although occasional bivalves are found.

At least two planar, 5 to 20 m beds of dusky red, laminated tuffaceous mudstone occur interbedded with limestones, shale, and siltstone, in the allochthon in the southwestern low hills, in the vicinity of Rancho San Ambrosio and Rancho San Carlos. The rocks are very porous and extremely soft, only cropping out in deep stream gullies. The bedding is denoted by 20 mm layers of devitrified volcanic ash interbedded with 5 mm laminae of subround quartz silt, with ammonite impressions and rare glauconite.

The upper 160 m of the middle member consists of fine-grained, calcareous arkose and fine-grained, sandy oosparite that show a sequence of sedimentary and penecontemporaneous compaction structures at this stratigraphic interval throughout the Sierra de Santa Rosa. The two rock types were originally interbedded in 1.5 to 3 cm tabular beds traceable up to 100 m laterally. Varying degrees of soft sediment grain flow contorted the oosparite beds into partially and completely detached pinch and swell, or sedimentary boudinage structures. Selective solution of the oosparite beds and spheroids gives the arkose beds a raised appearance.

The arkose weathers a brownish gray, is dark bluish gray on a fresh surface, and exhibits an even, conchoidal fracture. It contains 59 percent subangular to subround, very fine to fine sand, and 41 percent spar. The fraction contains 65 percent clear, mono- and polycrystalline quartz, and 35 percent feldspar, including orthoclase, twinned plagioclase and microcline perthite. The spar is primarily optically continuous pore filling cement, but a minor portion is very fine, round recrystallized intraclasts. The oosparite is dark bluish gray fresh and weathers to dark brownish gray. It contains 64 percent spar and 36 percent very fine to fine sand. Twenty percent of the spar is very fine and fine-grained, recrystallized oolites, many with sand nuclei. The remainder is authigenic spar matrix. The sand fraction mineralogy is similar to that of the interbedded arkose.

The uppermost 30 m of the middle member is composed of a silty biomicrite nearly identical to the biomicrite occurring in the lower portion of the middle member.

Upper member (Jsru). The upper member is a sequence of argillite, lithic wacke, volcanic conglomerate and oosparite showing poor lateral continuity, exposed primarily on the west slope of the Sierra de Santa Rosa. The thickness at the type section is 550 m, but the thickness varies, reaching 700 m in the south portion of the range. The type section is located along the second ridge south of the road from La Cienega to the ruins of the Mina de Cajon de Las Amarillas, in the same allochthon as the type section for the lower member. The section starts at the intersection of the road from Rancho San Carlos to the ruins where it crosses a ridge near Rancho Horchata.

The bottom of the member is marked in most places by a 20 m bed of brownish black claystone with white quartz silt laminae .5 mm thick, at 3 to 5 mm intervals, showing flame and bed load deformation structures not observed elsewhere in the formation. Overlying the claystone, or directly on the middle member, is an interval of purplish black argillite, weathering yellowish gray, with occasional interbedded mudstones, shales and volcanic sand beds similar to those in the lower and middle members. The argillite fractures into characteristic sharp conchoidal fragments, is typically massive and contains 60 percent very well indurated clay matrix, 25 percent quartz silt and 15 percent sand size angular andesite rock fragments. The argillite contains poorly preserved ammonites and rare bivalves. This argillite-dominated sequence averages 60 m in thickness.

The lutites that make up the bottom of the member grade vertically into a fine to medium-grained sandstone sequence dominated by a calcareous subfeldspathic lithic wacke with lesser sandy micrites and fine-grained, subangular arkose. This sandstone assemblage forms most of the middle portion of the member, averaging 220 m in thickness. The sandstone shows gentle cross-bedded, or planar, laminae .5 to 2 mm thick, but is locally massive. Rare, well-rounded quartz pebble lenses contain bivalve and ammonite shell fragments.

The calcareous subfeldspathic lithic wacke weathers to a diagnostic very dusky red, but is light brownish gray on fresh surface. Sixty-two percent of the wacke is a sand framework filled by 18 percent detrital clay matrix and 20 percent calcite cement. The sand fraction is 57 percent quartz, 2 percent feldspar and 41 percent rock fragments, black devitrified volcanic rocks and calcite-replaced plagioclase (?). The much less common fine-grained arkose is interlensing with the wacke, primarily in the upper portion of the sequence. The rock is massive, very well indurated, pale yellowish brown where weathered, and light bluish gray where fresh. It is composed of 92 percent subangular sand and 8 percent clay matrix. The sand is 52 percent monocrystalline quartz and 48 percent feldspar.

Above the sandstone sequence is a volcanic sedimentary sequence of interlensed siltstone, sandstone and conglomerate. Specific lithologic types within this sequence have poor lateral continuity, but the composite sequence is present at this stratigraphic horizon throughout the Sierra de Santa Rosa, averaging 90 m in thickness.

The siltstone and sandstone are very similar to those of the lower member in color, grain size and mineralogy, but more commonly show planar lamination rather than cross-bedding with channel grading. The angular pebble conglomerate beds are better washed than those in the lower member, sometimes containing no interstitial clay or silt, the voids being filled by spar cement. The rudite clast mineralogy contains 73 percent volcanic rock fragments, including devitrified welded tuff and fine-grained pilotaxitic andesite porphyry; 18 percent feldspar, mostly altered plagioclase; and 9 percent quartz. In the northeast and west-central Sierra de Santa Rosa, this assemblage includes cobble conglomerate beds 1 to 5 m thick. The matrix is a massive, poorly sorted, volcanic lithic wacke. The subround clasts include older Precambrian-type metamorphic rocks, welded tuffs, andesite agglomerates, and a white, hypabyssal granite. This granite does not crop out in the Sierra de Santa Rosa.

The uppermost portion of the upper member consists of a relatively continuous alternating sequence of well-indurated, dark green calcareous siltstone and massive, greenish gray, fine-grained, sandy oosparite averaging 130 m in thickness. The average composition of the oosparite is 25 percent very fine to fine, subangular arkosic sand, 25 percent fine-grained, recrystallized spar oolites and 50 percent spar matrix. The sand is concentrated in 1 to 5 mm vein-like stringers through the rock in contrast to the even alternation of

beds in the oosparite-calcareous arkose succession of the middle member. The siltstone and oosparite occur in tabular, laterally persistent beds 1 to 3 m thick.

Fossils, age and correlation of the Sierra de Santa Rosa Formation.
Table 1 contains a composite list of the identified fossil species collected from the formation by W.T. Keller (1929) with F.E. Wellings, T. Flores (1929) and the present author, with updated generic and subgeneric geologic ranges (Moore, 1957). Figure 4 shows in which stratigraphic horizons these fossils are found, and Figure 3 shows the major fossil localities. Erich Jaworski (1929) identified and described Keller and Wellings' collection and designated an age of Lias (Early Jurassic) to the fossiliferous interval. Carlos Burckhardt (1930, and in Keller, 1929) reviewed the collection and also assigned a Lias age to it. Flores (1929, p. 150) assigned a Lias age to the area based on his own collection.

The most common and best preserved fossils in the Sierra de Santa Rosa are the bivalves. The two most common bivalve species, Weyla (Weyla) mexicana Jaworski and Trigonia inexpectata Sowerby, have updated geologic ranges restricted to the Lias (Moore, 1957b, p. N371 and p. N478). They occur in the silty shale and calcareous mudstone of the middle member. Flores (1929, p. 150) collected "...moldes de Pecten n. sp., del grupo Pecten pradoanus Verneuil et Collumb del Lias Inferior...," which is probably the same species as Neithea mexicana Jaworski, whose type locality is in the Sierra de Santa Rosa (1929, p. 3). Neithea has subsequently been elevated to group status, and the name is now Weyla (Weyla) mexicana Jaworski (Moore, 1957b, p. N371). Numerous impressions of at least two ammonite species commonly occur in the calcareous shale layer of the middle member, and occasionally in the lithic wacke and argillite beds of the upper member. Only the general outline and major ribs are preserved even in the best specimens, and none collected by the author could be identified. Flores (1929, p. 150) tentatively identified one of these species as belonging to the genus Arietites. Jaworski (1929, p. 9) assigned another type to the genus Deroceras. Both of these genera are restricted to the Lias (Moore, 1957a, p. L238 and p. L247). Star-shaped Crinoid columnals commonly occur with the bivalves. Flores (1929, p. 151) assigned them to the genus Pentacrinus, also restricted to the Lias (Shimer, 1944, p. 235). Poorly preserved gastropods and unidentified corals also occur in the limestones of the middle and upper members. A lone fragment of a fern leaf was found in an argillite layer in the upper member nearly adjacent to some ammonite fragments, but it was not identified. The lower 600 m of the formation, including all the lower member, has no age control. The lower member could quite possibly extend into the Upper Triassic, as the Triassic-Jurassic contact in Sonora is not marked by an unconformity (King, 1939, p. 1654).

White and Guiza (1948) described an unnamed sequence of marine sandstone, siltstone and conglomerate of Triassic and Liassic age resting unconformably on Permian limestone in El Antimonio District, 103 km west of the Sierra de Santa Rosa. The upper portion of the Antimonio section, Triassic (?) Unit H, and Jurassic Unit I might correlate to the lower and middle members of the Sierra de Santa Rosa Formation, respectively. Very general lithologic descriptions, an absence of paleontologic data, and the distance between the various localities make this correlation tenuous.

Table 1. Composite List of the Identified Fossils
Occurring in the Sierra de Santa Rosa Formation.

Fossil	Flores (1929)	Publication		Geologic Range
		Jaworski (1929)	This report	
BIVALVIA				
<u>Camptonectes</u> (<u>Camptonectes</u>) <u>sp.</u> Cret.			X	L.Jur.-U. Subgeneric
<u>Weyla</u> (<u>Weyla</u>) <u>mexicana</u> Jaworski	X	X	X	L.Jur. Subgeneric
<u>Alectryonia</u> <u>gregarea</u> Sowerby		X		L. Jur.-U. Jur. Specific
<u>Lima</u> <u>nodulosa</u> Terq.		X		Jur.-Rec. Generic
<u>Plagiostoma</u> cf. <u>P. exaltata</u> Terq.		X		M. Trais.-U.Cret. Subgeneric
<u>Trigonia</u> cf. <u>T. inexpectata</u> Jaworski		X	X	L.Jur. Subgeneric
<u>Myoconcha</u> cf. <u>M. Valenciennesi</u> Bayle		X		L.Jur.-U.Cret. Generic
<u>Astarte</u> cf. <u>A. antipodium</u> Geibel		X		Jur.Rec. Generic
<u>Pholadomya</u> <u>sp.</u>			X	U.Trais.-Rec. Generic
AMMONIODEA				
<u>Arietites</u> <u>sp.</u>	X		?	L.Jur. Generic
<u>Deroceras</u> <u>sp.</u>		X		L.Jur. Generic
CRINOIDEA				
<u>Pentacrinus</u> <u>sp.</u>	X		?	L.Jur. Generic

The Sierra de Santa Rosa Formation is probably time equivalent to the upper portion of the original continental and brackish marine Barranca Formation in southeastern Sonora, now the Coyotes Formation of De Cserna (1961). Lower Jurassic marine fossils occur elsewhere in Sonora (See Discussion) and further mapping may determine them to be correlative to the Sierra de Santa Rosa Formation.

Cerro San Luis Formation (Jsl)

The new name Cerro San Luis Formation is here proposed for a continuous sequence of Jurassic and possibly Cretaceous conglomerate and andesitic volcanic rocks conformably overlying the Sierra de Santa Rosa Formation. The formation is exposed primarily in the low hills west of the Cerro San Luis, but is also structurally juxtaposed to the Sierra de Santa Rosa Formation in the eastern portion of the range. The thickness of the volcanic sequence is unknown and probably exceeds the 860 m of the type section. The type section is through a stream gully cutting the west slope of the Cerro San Luis. The top and east end of the section is at Rancho San Luis. The formation is informally divided into two members on the basis of lithology.

Conglomerate member (Jslc). Conformably overlying the upper member of the Sierra de Santa Rosa Formation is a lensing, 0 to 70 m layer of quartz arenite and dolomite cobble conglomerate. The conglomerate is best exposed in the north-central portion of the range. Where exposed, the contact between the formations consists of a pebble bed on top of calcareous mudstones showing penecontemporaneous sediment flow and partially detached rip ups in response to the weight of overlying clasts. The matrix of the lower few meters of the conglomerate contains intraclasts of the underlying muds. The lower pebble bed grades rapidly upward into unoriented, angular to well-rounded cobble beds with rare, isolated boulders and pebble lenses. In gully outcrops, the conglomerate commonly fractures across clasts forming a smooth, mosaic pattern.

Bluish white to grayish orange pink, fine-grained, well-rounded supermature quartz arenite comprises 62 percent of the clast percentage. The remaining 38 percent is composed of a light greenish gray to pale yellow orange, sandy, massive, laminated and algal mat dolomite. These rudite clasts are virtually identical to younger Precambrian sedimentary rocks exposed in the area. The matrix is dusky red, very fine to medium-grained, massive and composed of 52 percent mono- and polycrystalline quartz, including sedimentary and metamorphic quartzites, 40 percent dolomite rock fragments and 8 percent clay matrix.

Volcanic member (Jslv). Conformably overlying the conglomerate member where present, and elsewhere directly on top of the upper member of the Sierra de Santa Rosa Formation, is a heterogeneous assemblage of andesitic agglomerate, lava flows, lapilli tuff and flow breccia with lesser interbedded volcanic conglomerate and arenite. The maximum thickness of the member exceeds the 860 m of the type section, but is impossible to determine due to incomplete exposure, and the possibility of undetected reverse faulting within the member. The relative abundances of rock types in the type section are characteristic of the member in

the northern portion of the range, but the exposures east of Rancho San Carlos contain fewer agglomerate and flows, and a higher percentage of finer-grained tuff, pebble conglomerate and volcanic sandstone. The contact with the conglomerate member is a subangular pebble conglomerate layer a few meters thick that shows a rapid gradation from a quartz arenite and dolomite clast assemblage to a welded tuff and pilotaxitic volcanic rock fragment clast assemblage set in a lithic arenite matrix typical of the conglomerate member. This grades upward within 10m to cobble conglomerate or tuff and lava, with rare quartz arenite and dolomite clasts, and next above to an agglomerate.

The andesite lava, comprising only 5 to 15 percent of the member in the type section, is altered grayish green and grayish purple hyalopilitic cumuloaphyric hornblende andesite, occurring in flows 5 to 70 m thick. The phenocrysts are 62 percent oligoclase to andesine plagioclase, 36 percent chloritosericitic aggregates with magnetite (?) rims formed from the alteration of hornblende, and 2 percent apatite. The phenocrysts range in size from .03 to 4.5 mm. The original matrix was mostly volcanic glass which devitrified into hyalopilitic quartz, feldspars and zeolites. Eighteen percent of the matrix is now secondary calcite and lesser epidote.

The flows are interstratified in grayish purple and grayish green lithic agglomerate, comprising 70 percent to 85 percent of the member. The agglomerate blocks range from lapilli size to 50 cm in diameter and form massive beds up to 70 m thick. Over 90 percent of the blocks are composed of andesitic lava, lapilli and fine tuff, and less than 10 percent are foliated metamorphic rock of the older Precambrian type exposed in the area. The interstratified, welded lapilli tuff forming 5 to 15 percent of the member show bedding denoted by medium to coarse plagioclase crystals and deformed pumice fragments .2 to 5 cm long. The matrix of both the agglomerates and tuffs is similar to that of the lavas.

Interbedded in this pyroclastic sequence are lensoidal deposits of pebble to cobble conglomerates and poorly sorted arenite consisting of clasts of andesite of the same composition as the pyroclastic sequence in which they are included. At the type locality and in the Cerro San Luis area, these sediments comprise only 5 percent of the member, but to the south, in the vicinity of Cerro Garambullo and Rancho San Carlos, the arenites comprise up to 40 percent of the member.

The conglomerate is grayish purple and contains unoriented, subangular to round pebble to boulder clasts, 80 percent cobble size. The arenite is grayish purple and grayish green, commonly channel graded and cross-bedded in 5 m thick lenses within the conglomerate. In some beds the only criteria for distinguishing these sediments from the agglomerate and tuff are the round shape and sharp boundaries of the conglomerate clasts and cross-bedding in the arenites. Thin-section inspection reveals subround quartz in most of the arenites and rudite matrix. Most of the conglomerate contains up to 15 percent well rounded, pebble and cobble size younger Precambrian sedimentary clasts identical to those in the conglomerate member, and foliated metamorphic rocks similar to those in the older Precambrian basement.

Age and correlation of the Cerro San Luis Formation. The base of the formation rests conformably on strata containing Lower Jurassic fossils. Therefore, the conglomerate member and at least a major portion of the volcanic member probably occupies the

Lower and Middle Jurassic. Thin sections of the lavas, agglomerates and feeder dikes to the formation were inspected for suitability for K/Ar dating, but the recrystallization was too great in every case.

The volcanic member of the formation exhibits petrologic and internal stratigraphic characteristics very similar to the Lista Blanca Formation of Dumble (1900, p. 147), whose type section is near San Marcial, 220 km southeast of the Sierra de Santa Rosa. Dumble stated:

The impression made on us in the field was very strong that they (the volcanics) and the underlying Triassic beds (Barranca Group) were part of one system.

This is interpreted as meaning that an actual conformable contact was not observed, but the two rock types were in close proximity and showed no angular discordance. The conformable relationship is positively observed in the Sierra de Santa Rosa. This is rare, however, as King (1939, p. 1655) found the widespread Lista Blanca variety of volcanic rocks unconformably on Barranca strata and interbedded with fossiliferous Cretaceous rocks. Salas (1968) mapped volcanic deposits very similar to the Cerro San Luis Formation volcanic rocks interbedded with Albian-Aptian Cretaceous sedimentary rocks just east of the town of Santa Ana, 100 km northwest from the Sierra de Santa Rosa. King assigned all the pre-Tertiary volcanic rocks in Sonora to the Cretaceous, but it appears that at least isolated andesitic volcanism occurred in northwestern Sonora as early as the Jurassic.

Origin of the Sierra de Santa Rosa and Cerro San Luis Formations

Provenance. Arenite and rudite clast mineralogies indicate that the source area of the terrigenous rocks of the Sierra de Santa Rosa Formation was primarily a coarse crystalline, regionally metamorphosed, and acidic plutonic terrane, partially covered by andesitic pyroclastic and possibly lava deposits. The chemical instability, angularity and poor sorting of the volcanic detritus suggests that they were eroded a relatively short time after eruption and were transported a relatively short distance to the site of deposition. However, the absence of all but the finest grained pyroclastic deposits in the formation implies that these eruptions, if syndepositional, were not in the immediate vicinity of the area of deposition. The absence of very stable, supermature Precambrian quartz arenite rock fragments in the Sierra de Santa Rosa Formation suggests that these strata and their associated dolomites, now common in the region, were not widely exposed during deposition.

The abrupt but conformable introduction of younger Precambrian quartz arenite and dolomite cobble clasts forming the conglomerate member of the Cerro San Luis Formation is evidence for a localized tectonic event near the site of deposition. Andesitic volcanic material was abruptly re-introduced to the area during deposition of the volcanic member, first in the form of volcanic conglomerate and wacke, and then directly as pyroclastic and lava deposits. The interbedded, channelized conglomerate and sandstone, mostly derived from the agglomerate, tuff and lava they were deposited on, are not believed to mark major episodes of volcanic quiescence. Younger Precambrian clasts in the interbedded conglomerate indicate that this source terrane for the conglomerate member was emergent throughout deposition of the volcanic member. Gneiss and schist xenoliths in the agglomerate and flow breccia suggest that the depositional basement for the entire Jurassic sequence was the older Precambrian metamorphic rocks.

Environment of deposition. The lower member of the Sierra de Santa Rosa Formation is composed of channelized, mixed carbonate-clastic rocks with benthonic shallow water marine fossil fragments. The entire member is interpreted to have been deposited on a shallow tidal shelf. The different lithologies and poor lateral continuities of the member are a function of the different subenvironments on the tidal shelf. The angular, cross-bedded pebbles and sands were deposited in tidal channels meandering across the tidal shelf. The muds were deposited in channels as current energy dropped and in adjacent tidal mud flats when flow spilled over the channels. The silty micrite probably formed in these same mud flats with a more restricted circulation.

The middle member is also a mixed clastic-carbonate assemblage, but this is attributed to a more limited detrital supply in a continually transgressing sea. The member was deposited in two subenvironments offshore from the shallow tidal flat shelf facies of the lower member. The fossiliferous calcareous shale and mudstone were deposited in protected but aerated lagoons away from any channel mouths, but still receiving silt and clay. The associated biomicrite was deposited in these same lagoons during periods of very low sediment influx. The interbedded calcareous arkose and oosparite of the upper portion of the middle member were deposited in a high energy, laterally persistent environment, most probably a barrier bar complex that protected the shoreward lagoon and tidal flat in the transgressing sea. These well sorted sands and oolites probably represent the most open marine deposition in the formation.

The uppermost biomicrite of the middle member and the mudstone and argillite of the upper member were most probably deposited in a protected shallow water environment, probably in a lagoon shoreward from the barrier bar. This would indicate the end of the marine transgression in the sequence, or at least a seaward migration of the barrier bar. The fern leaf fragment found in the argillite of the upper member indicates a low current energy between this basin and land. The remainder of the upper member contains features of both the lower and middle members, and probably represents a transition between the barrier and tidal flat environments and filling of any well-developed, widespread lagoon present in the lower portion of the upper member. This might have been caused by a progradation of the tidal flat or destruction of the barrier bar. The interbedding of clean, calcareous sands and oolites with poorly sorted, angular volcanic sands could have been caused by tidal channels breaching the barrier bar.

In the southwestern low hills and the southern portion of the range east of Rancho San Carlos, the middle and upper members contain a greater percentage of biomicrite. This is possibly due to a starved sediment supply in the lagoons because of a position farther offshore from the barrier bar-lagoon complex.

The sharp contact at the base of the conglomerate member of the Cerro San Luis Formation suggests a rapid filling of the shallow sea in the depositional area. This filling was caused by the nearly adjacent uplift of younger Precambrian sedimentary rocks already mentioned. Portions of the underlying marine sediments were most probably swept away, but soft sediment deformation at the contact and bedding concordance show that the conglomerate member is not a basal deposit upon a lithified, folded terrane. The conglomerate could have been partially deposited in a very nearshore marine environment with high relief, but no positive evidence supports this. Its texture is typical of flood plain and alluvial fan deposits in close proximity to the elevated source rocks. The absence of the conglomerate member in the southwest and south-central areas further suggests that these strata were deposited farther from the source area. The interbedded conglomerate and sandstone beds of the volcanic member show channel grading, cross-bedding and lensoidal geometries. This suggests that they were deposited subaerially in meandering streams upon the volcanic terrane.

POST-JURASSIC ROCKS

Late Mesozoic-Early Tertiary (?) Intrusive Rocks (MTi)

An intrusive, light colored rhyolite porphyry of post-low angle faulting and pre-mineralization age crops out at many places in the Sierra de Santa Rosa. The porphyry was emplaced as a set of irregularly-shaped dikes, generally oriented along strike of the Jurassic strata and commonly along low and high angle reverse fault zones. The intrusive bodies are resistant to erosion, forming the peak of Cerro San Luis and other ridges in the Sierra de Santa Rosa. In outcrop appearance the porphyry is commonly fractured into angular, boulder size blocks with planar sides, and weathers to a very pale orange. A fresh exposure is light gray with white phenocrysts. The phenocrysts comprise about 37 percent of the rock and are set in a sericitized quartz and feldspar matrix. The phenocrysts are anhedral quartz and sericitized megacrystic microcline, with lesser antiperthite, oligoclase and andesine. Many of the dikes have been subjected to epithermal mineralization, and were the major production rocks for the Mina de Cajon de Las Amarillas and the Mina de Santa Rosa.

Mesozoic granitic intrusions are common in central and northwest Sonora (King, 1939), but none crop out in the Sierra de Santa Rosa. Salas (1968) observed intrusive rhyolite porphyries associated with granites in the Santa Ana region 100 km northwest of the Sierra de Santa Rosa. Fine to medium-grained granites crop out adjacent to the Sierra de Santa Rosa, out of the mapped area. They show the same sericite alteration as the rhyolite dikes, but none of the regional metamorphism characteristics of the foliated older Precambrian rocks. Therefore, they might be phaneritic equivalents of the rhyolite porphyry in the Sierra de Santa Rosa.

Field relations in the Sierra de Santa Rosa do not permit the assignment of a precise age to the dikes, and sericite alteration prevents an accurate K/Ar age determination. Salas (1968) suggests a Late Cretaceous or Early Tertiary age for the rhyolite porphyry and associated granite in the Santa Ana area, but no evidence is given. Anderson and others (1969, p. 3) obtained isotopic U/Pb zircon ages in the Sierra Seri, along the coast of Sonora 120 km west of the Sierra de Santa Rosa, that show a Late Cretaceous magmatic event of regional dimensions. The ages of emplacement for Mesozoic granitic rocks appear to show a younger gradation eastward from the coast toward the Sierra de Santa Rosa region (Anderson, 1972, personal communication). The granitic emplacements of central Sonora were also suspected of being Late Cretaceous in age by King (1939). The rhyolite porphyry dikes of the Sierra de Santa Rosa are possibly of the same age.

Late (?) Tertiary Baucarit Formation (Tb)

In the Sierra de Santa Rosa, the Baucarit Formation (Dumble, 1900, p.151) occurs as a poorly consolidated, thin-bedded, fine to medium-grained, moderate pink sandstone with beds of subround cobble conglomerate, cropping out in a low area east of Rancho San Carlos. The formation dips up to 25 degrees to the west forming an angular unconformity with the underlying, more steeply dipping Jurassic strata, and is partially covered by flat-lying older Quaternary boulder conglomerate beds. The conglomerate beds in the formation are composed of cobbles entirely derived from the sedimentary and

volcanic rocks of the underlying Jurassic strata, Late (?) Cretaceous granitic rocks and the older Precambrian basement. The sandstone is a volcanic lithic wacke containing primarily andesitic tuff and siltstone rock fragments, altered plagioclase and quartz. The sandstone beds contain mud cracks indicating subaerial deposition, probably on an alluvial flood plain during initial emergence of the Sierra de Santa Rosa. Exposure of the formation is restricted to this one valley, where subsequent structural tilting has preserved the sequence, and overlying gravels have been removed.

The Baucarit Formation was defined by Dumble to include poorly consolidated, evenly bedded pink sandstone with conglomerate that occurs throughout east-central Sonora. King (1939, p. 1681) refined the definition to include rocks of this type laid down as alluvial fan and flood plain deposits in intermontane depressions, commonly preserved from erosion by subsequent down faulting. King (1939, p. 1962) observed that the formation overlies volcanic deposits of Early Tertiary age and is covered by undeformed gravel beds of Quaternary age, so a Late Tertiary age is inferred by King. He considers the Baucarit Formation to be nearly the same age as the Miocene Gila Conglomerate of southern Arizona.

Older Quaternary Alluvium (Qoa)

Extensive deposits of flat-lying, unconsolidated, boulder to cobble conglomerate, presently undergoing dissection, occur on the flanks of the Sierra de Santa Rosa. These cap many of the low ridges west of the range crest and cover a portion of the west-central Sierra de Santa Rosa itself, 300 m higher than the valley floor 6 km to the west. The capping conglomerate is at the same elevation as the dissected alluvium on the valley floor to the east. Therefore, west of the range divide they are interpreted to be erosional remnants of extensive alluvial deposits that buried a major portion of the west slope of the range. These conglomerates show no evidence of structural deformation. Their occurrence at high elevations and their present dissection are attributed to a lowering of the regional base level by widespread crustal uplift, or the "undamming" of intermontane valleys.

Quaternary Alluvium (Qal)

Unsorted gravels, sands and muds lie in stream gullies flanking the range, in localized "dammed depressions" and on the valley floor to the west of the range. These are the only areas not presently undergoing dissection in the mapped area.

STRUCTURAL GEOLOGY

Low Angle Faults and Folds

The structure of the Sierra de Santa Rosa is dominated by a series of imbricated and locally folded allochthons (Figure #3). They are entirely composed of rocks of the Sierra de Santa Rosa and Cerro San Luis Formations, except for

rare, isolated klippen of younger Precambrian sedimentary rocks. The structural basement is composed of the older Precambrian metamorphic complex and the autochthonous younger Precambrian sedimentary sequence. Older Precambrian rocks were not found to be emplaced over younger sedimentary rocks, as was observed in the Cerro Rajon and Cerro Chino 50 km to the north (Merriam and Chandler, 1971). However, the metamorphic terrane was not studied in enough detail to rule out the possibility of additional low angle faults within the older Precambrian complex. In the Cerro Calaveras area 30 km east of the Sierra de Santa Rosa, Eells, (Anderson, Eells, and Silver, 1978) mapped an extensive system of younger Precambrian and Cambrian strata emplaced eastward over similar age strata, and over the older Precambrian metamorphic complex.

The rocks adjacent to the fault planes are highly fractured and, in many cases, mineralized or intruded by the white rhyolite dikes. In some areas the fault planes are actually defined by a zone of brecciation or a series of minor detachment surfaces up to 50 m thick. However, the major fault planes themselves are generally unexposed, making it necessary to determine their locations by the recognition of stratigraphic units and abrupt changes in structural trend.

A set of bedding plane thrust faults within the Jurassic sequence in the Sierra de Santa Rosa define a set of imbricate allochthons (informally named in Figure #5). The lowermost is in the east front of the range and the uppermost is in the western foothills. The eastern and central allochthons, forming the crest and west front of the central range, are bounded by successive, overriding thrusts with surface dips averaging 40 degrees to the west. These contain Jurassic strata with dips nearly parallel to the fault planes.

East of Hill 1315, the "eastern imbricate allochthon" is bound by a frontal, mildly folded low-angle fault, which dips gently to the west. Two successive, overriding thrust faults dip up to 40 degrees to the west. This allochthon was overthrust by an allochthon containing the entire Jurassic section, homoclinally dipping to the west. This "central homoclinal allochthon" forms most of the eastern front and crest of the range. A major reverse fault along the western front of the range defines the base of the "western allochthon," which overthrust the central homoclinal allochthon. The western allochthon contains a repetition of the entire Jurassic section contained in the central homoclinal allochthon and the strata show nearly the same attitudes, striking N 5° - 15° W and dipping 30° - 80° to the west. The stratigraphy shows good correlation across the structural boundaries of these two "homoclinal allochthons," which suggests the strata forming the allochthons were deposited in close proximity to one another. Careful inspection of sedimentary structures and vertical lithologic sequences through the "imbricated" and "homoclinal" allochthons indicate no beds are overturned, precluding the possibility of a isoclinal fold complex. Also, the possibility that the repetition of beds through this sequence is due simply to a cyclic nature of a given set of sedimentary environments is highly unlikely, as the vertical succession of strata from fault block to fault block is virtually identical.

The "southwest folded allochthon" contains a series of east-west striking folds, steeply plunging to the west, consisting of correlatable, but less similar strata than the homoclinal allochthons, which suggests a tectonic juxtaposition of this allochthon from a more distant depositional location. The tight folding of the strata does not extend into the basement, and was possibly caused by a north-south compression prior to eastward emplacement.

The imbrication attitudes and drag fold axis measurements along faults within the allochthons indicate an eastward component of thrusting. The entire Jurassic allochthon

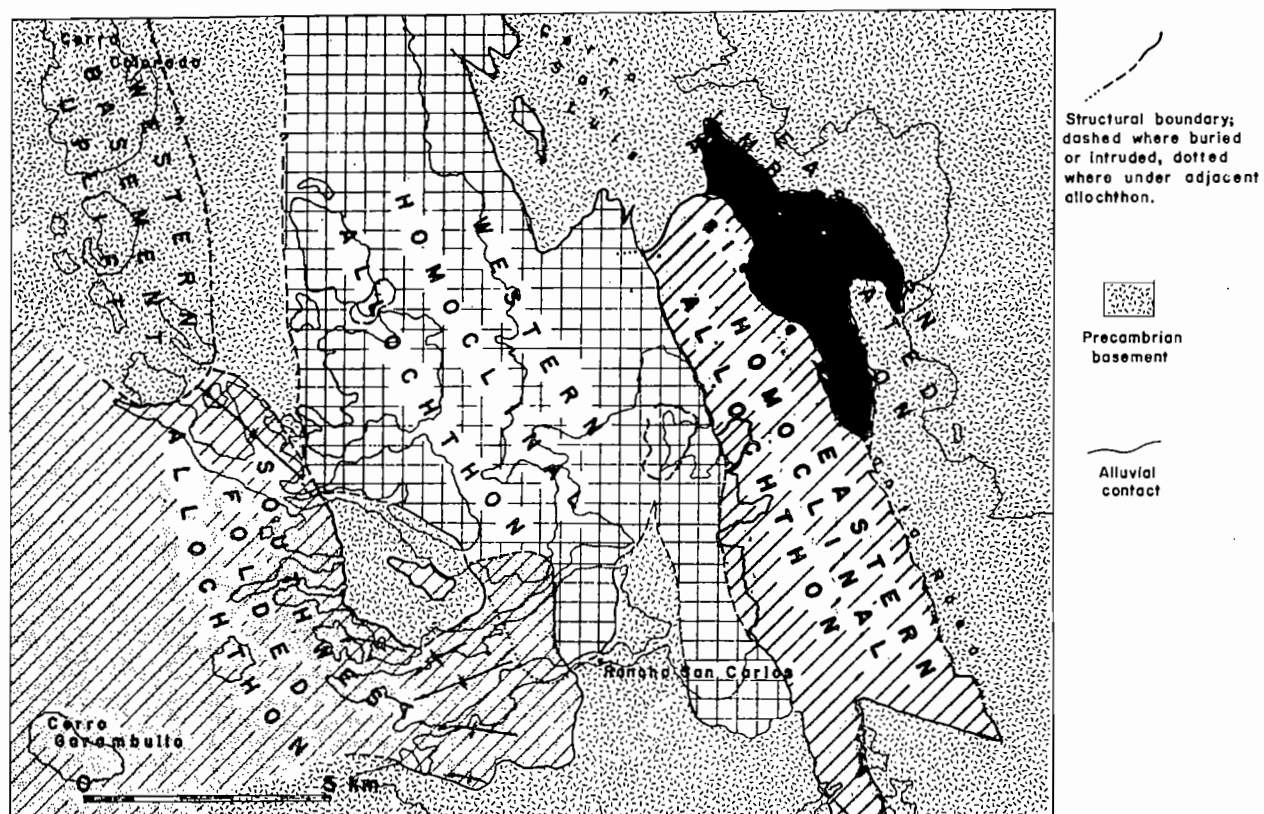


FIGURE 5 - Generalized structural map of the Sierra de Santa Rosa showing the distribution of the allochthons. Tertiary normal faults and mesozoic intrusions omitted.

complex was also probably emplaced from the west (Gastil and Krummenacher, 1977). This is consistent with Merriam and Chandler (1971) and Anderson, et. al., (1978) for movement in the ranges to the west. Regional geologic mapping is not sufficient to determine the distance of movement with certainty, but no autochthonous Triassic or Jurassic exposures are known for at least 50 km to the west. The low angle faults have displaced andesitic volcanic rocks of Jurassic or possibly Early Cretaceous age, but are cross-cut by rhyolite dikes of probable Late Cretaceous or Early Tertiary age. Therefore, the age of tectonic emplacement is generally correlatable to the "Mid" Cretaceous "Sevier Orogeny" of the Western United States. This age predates the Late Cretaceous Laramide age suggested by Merriam and Chandler in the Cerro Rajon, and by King (1939) for thrusting in central Sonora, 250 km east of the Sierra de Santa Rosa.

The angular relationship between the moderately dipping allochthonous Jurassic strata and the low angle fault plane juxtaposing this complex against Precambrian basement argues against a thrust fault emplacement. Either this thrust would have had to cut steadily down section for at least 10 km, or it would have had to cut through previously tilted intervals across strata of varying competency. Neither mechanism is typical of Western North America thrust fault style (Armstrong, 1968), or basic structural geologic principle. More likely the Jurassic allochthon complex was emplaced by gravity sliding from the west, as suggested by Gastil and Krummenacher (1977). The strata and imbricate thrust faults were probably rotated to their present moderate westward dip during emplacement.

An alternate explanation for the present geometry of moderately dipping Jurassic strata truncated by a low angle fault is described by Proffett (1977). In much of the western Basin and Range province of Nevada, originally high angle listric normal faults have been rotated to nearly horizontal attitudes by subsequent movement on younger, successive sets of high angle normal faults. However, evidence of such extensive subsequent normal faulting is lacking in the Sierra de Santa Rosa.

Normal Faults

Post-thrust and post-intrusion normal faults occur throughout the Sierra de Santa Rosa, but no major, range-forming longitudinal faults are exposed in the mapped area. The normal faults do not show a predominant orientation throughout the range, nor do they alter the configuration of the allochthons in a major way.

In the north-central portion of the mapped area, in the vicinity of Cerro San Luis and Hill 1315, a set of normal faults trends approximately N 40° W. The displacement at the center of these faults is as great as 200 m and is observed to die out within 500 m laterally. A single fault striking N 30° E cuts the western homoclinal allochthon and the metamorphic complex in this same area. In the southeastern portion of the range, within the central homoclinal allochthon, a set of transverse normal faults of minor displacement striking N 80° W cut the intrusive rhyolite and the members of the Sierra de Santa Rosa Formation. The southernmost of these faults defines a portion of the boundary of the central homoclinal allochthon because erosion has removed the entire Jurassic section exposing basement rock on the southern, upthrown side. Vertical displacement on this fault need not be more than a few hundred meters because the allochthon is less than 300 m thick on the northern, downthrown side.

The south-central portion of the range has undergone approximately 25 degrees of tilting to the west since deposition of the Baucarit Formation in the Tertiary. This tilting does not extend west or south to where basement rock and the untilted southwest folded

allochthon crop out. A normal longitudinal fault concealed under the alluvial gravels along the front of the Sierra de Santa Rosa may define the westward limit of this localized tilting within the range. However, there is no positive evidence for such a fault.

DISCUSSION

Possible Areal Extent of the Sierra de Santa Rosa Formation

G. A. De Cserna (1961, p. 7) defined two major depositional realms for the Barranca Group. The Arrayanes, Santa Clara and Coyotes Formations were deposited in the fresh water and brackish marine "San Marcial Basin" in central Sonora, southeast of Hermosillo (Figure #2). The open marine shelf deposits of El Antimonio and the Sierra de Santa Rosa were deposited in the "Antimonio Paleobay" in the Caborca region of northwest Sonora. Figure 2 shows the known exposures of possible equivalents of the Sierra de Santa Rosa Formation also being deposited in the open marine shelf of the Antimonio Paleobay of northwestern and central Sonora. The descriptions of these locations are, for the most part, very brief, and thicknesses have only been measured at two locations outside the Sierra de Santa Rosa. Enumeration of these locations is not meant to propose that they be incorporated into the Sierra de Santa Rosa Formation, but simply as an aid to further field study and paleogeographic reconstruction of the marine Triassic-Jurassic System of Sonora.

At the mine at Las Animas (Figure 2) 35 km east of the Sierra de Santa Rosa, Flores (1929, p. 151) noticed a small outcrop of yellow-brown limestone with shales containing a fossil assemblage similar to that of the Sierra de Santa Rosa Formation. In the southern portion of the Sierra de Carachui 85 km east of the Sierra de Santa Rosa, Flores (1929, p. 150) found another assemblage of Lias fossils in a sequence of limestones, sandstones and argillites. At Cerro Pozo Serna 48 km southwest of the Sierra de Santa Rosa, R. J. Dowlen and T. E. Stump (Beauvais and Stump, 1976) described a 1,200 m thick sequence of laminated siltstones, biomicrites and litharenites containing Jurassic fossils. The sequence is in fault contact with all other rocks, but no low angle fault plane is exposed. During examination of the already mentioned sequence at El Antimonio 103 km west of the Sierra de Santa Rosa, White found Liassic fossils at the Sierra del Alamo Muerto just south of El Antimonio (Cooper and Arellano, 1946, p. 611).

The effect of a Coexistent Magmatic Island Arc-Trench System on the Deposition and Deformation of the Sierra de Santa Rosa and Cerro San Luis Formations

Oceanic magnetic anomaly evidence and plate tectonics theory indicate that western Sonora was bordered by a magmatic island arc (andesite-diorite belt) during Jurassic time (Gastil and Krummenacher, 1977). This north-south arc was bordered on the west by an oceanic trench. Stratigraphic evidence indicates that the region east of the arc, which is now west-central Sonora, including the depositional site of the Sierra de Santa Rosa and Cerro San Luis Formations, was covered by a shallow sea during Late Triassic and

Early Jurassic time (De Cserna, 1961). The relation of this sea to the arc-trench system was that of a "foreland marine basin in a back arc area" (Dickenson, 1971, p. 167). Stratigraphic evidence indicates that a similar, though larger, sea existed behind a northward extension of the arc-trench system in western Nevada and eastern Oregon during the Early Jurassic (Stanley and others, 1971). A sequence of Triassic and Jurassic "Western Belt Sediments" with lithofacies similar to those of the Sierra de Santa Rosa Formation were deposited east of the arc in the present-day northwestern Basin and Range province. These sediments are interpreted to have been derived from volcanic islands and rapidly emerging "tectonic lands" of the arc to the west. Clast mineralogies of the wackes and conglomerates of the Sierra de Santa Rosa Formation suggest that the source area was this andesite and diorite arc to the west.

The orogenic events during the Mesozoic evolution of the Sierra de Santa Rosa might also be related to the arc-trench system. Land thrust belts have been observed in areas immediately behind arcs (Dickenson, 1971, p. 170), and are probably the result of adjustments of the arc mass to adjacent subduction forces and isostatic adjustments. The orogenic event responsible for the abrupt introduction of the quartz arenite and dolomite conglomerate of the Cerro San Luis Formation into the depositional area could be related to these compressive forces between the arc and the site of deposition. Similar "orogenic" conglomerates occur in the Early Jurassic "Western Belt Sediments" of northwestern Nevada, notably the Boyer Ranch Formation (Stanley and others, 1971, p. 16). These orogenic conglomerates are commonly followed by andesitic volcanism and this is also the case in the Cerro San Luis Formation. The eastward emplacement of the entire Jurassic sequence forming the present-day Sierra de Santa Rosa was probably the result of gravity sliding from the west due to regional uplift of this back-arc area after thrusting.

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TERTIARY MARINE CONGLOMERATE SOUTHWESTERN TIBURON ISLAND

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ABSTRACT

An unexampled 988 meter thick section of fossiliferous volcanoclastic marine conglomerate outcrops in the southwest corner of Isla Tiburón. The section has been divided into three members, a lower fossiliferous member, middle nonfossiliferous, and an upper fossiliferous member. Microfossils extracted from the upper member give a tentative late Miocene or early Pliocene age. Deposition of the conglomerate took place in a subsiding, near shore basin, the origin of which is possibly the result of the initial dilation of the Gulf of California.

INTRODUCTION

Isla Tiburón, largest island in the Gulf of California, lies off the coast of Sonora Mexico at lat 29° N, long $112^{\circ} 30'$ W (Fig. 1). Almost nothing of the geology was known prior to the mapping by San Diego State University students in 1970 and 1971 (Gastil and Krummenacher, 1977). Their reconnaissance revealed a thick section of fossiliferous, marine conglomerate in the southwest corner of the island (Fig. 1).

GEOLOGIC SETTING

The geology of Isla Tiburón is a continuation of that of coastal Sonora. The rocks of Isla Tiburón have been divided into several time/lithologic units (Gastil and Krummenacher, 1974 and 1977). In the southwest corner of the island a major fault separates two distinct rock assemblages (Fig. 1). South of the fault the rock assemblage is tantamount to Gastil and Krummenacher's T1 time/lithologic series (greater than 22 m.y.). The assemblage here is comprised of rhyolite, andesite, basalt, ash flow tuff, and a small amount of lacustrine limestone.

North of the fault the rock assemblage is comprised of Gastil and Krummenacher's T2 and T3 time/lithologic units. Rocks in this area are andesites and rhyolites (18-22 m.y.) and volcanoclastic conglomerate (8-12 m.y.?) respectively. The volcanoclastic conglomerate has also been mapped along coastal Sonora and elsewhere on Isla Tiburón, but only in the southern corner of the island has it been found to be marine in origin.

DESCRIPTION OF THE MARINE CONGLOMERATE

Outcrops of the marine conglomerate are exposed in the bluffs of arroyos A and B (Fig. 1 and 2). Outcrops in arroyo A first appear a few hundred yards inland, forming the northwest wall of the arroyo. This exposure continues south-east for 2,430 meters. Beds of the conglomerate dip consistently $24^{\circ} \pm 5^{\circ}$ to the northwest, giving a true thickness of 988 meters. Top and bottom contacts of the conglomerate are covered. A few caliche filled joints cut

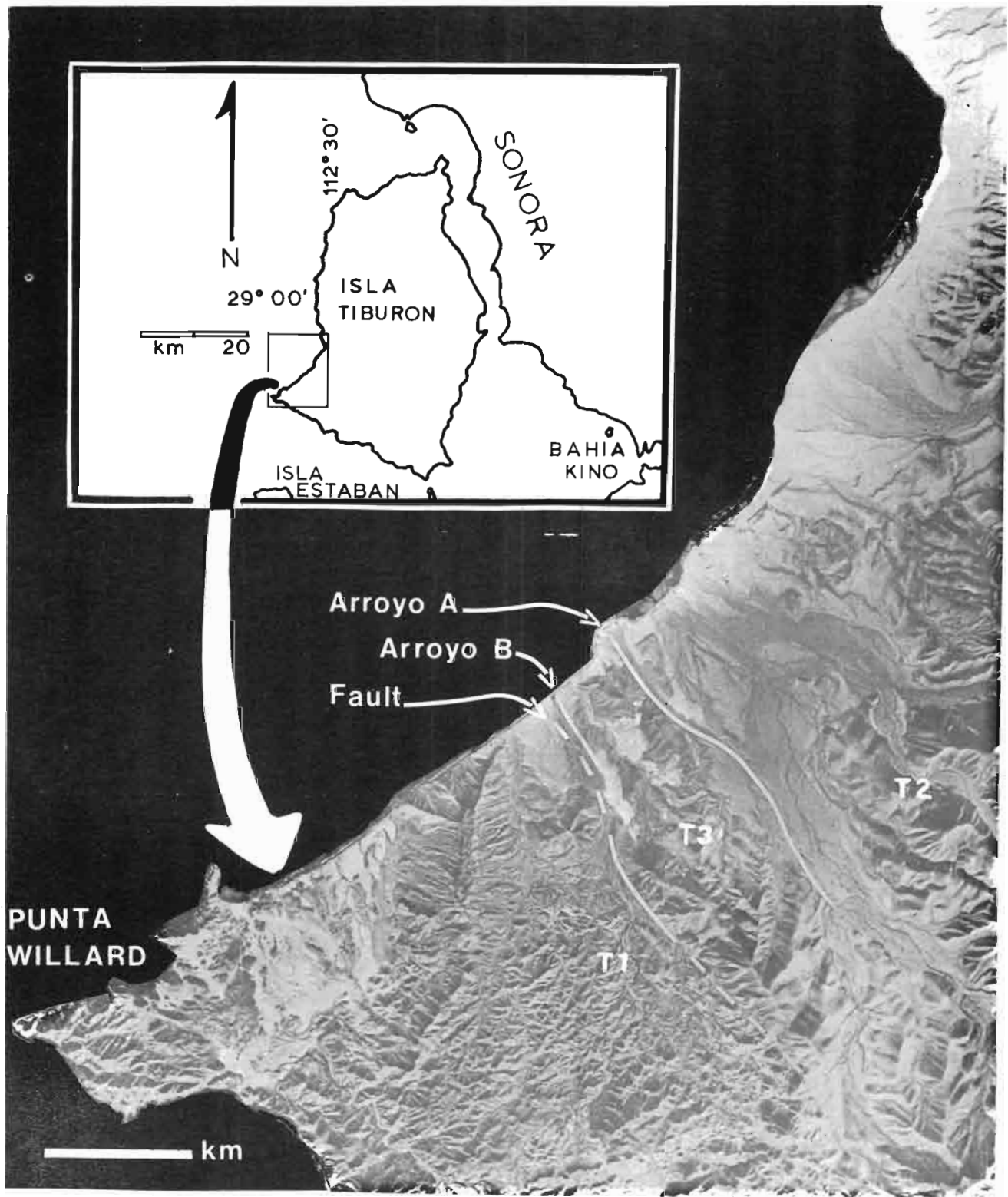


Figure 1. Index map to the southwestern corner of Isla Tiburón. Arroyos A and B are locations where the fossiliferous marine conglomerate outcrops.



A. View of conglomerate looking west in arroyo A.



B. Northwest dipping conglomerate overlain by flat lying quaternary deposits.



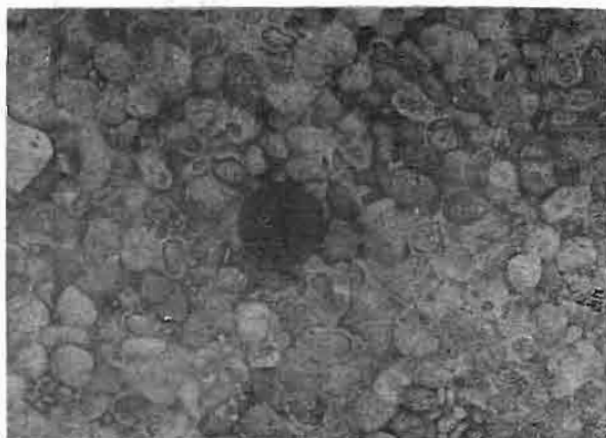
C. Pecten in arroyo A outcrop.



D. Oyster rich bed in arroyo A outcrop.



E. Typical individual sedimentation units.



F. Conglomerate bed with resistant carbonate cement.

Figure 2. Photographs showing features of the volcaniclastic marine conglomerate.

the conglomerate but no offset is apparent.

Three distinct divisions are recognized within the conglomerate outcrop, an upper fossiliferous member (mcg-1), a middle nonfossiliferous member (mcg-2) and a lower fossiliferous member (mcg-3) (Weaver, 1979). Member (mcg-1) forms the lower 400 meters of the exposed section. It is a bedded, poorly sorted, pebbly, light gray, fossiliferous conglomerate (Fig. 2). Clasts are well rounded rhyolite, andesite and olivine rich gabbro. Individual sedimentation units are 5 to 25 centimeters thick. Clast size ranges from grains to cobbles, with pebbles dominant. Individual beds tend to be clast supported. The matrix is a very immature, sandy, siltstone, in places grading to a clayey siltstone. Induration is moderate with increased cementation occurring in highly fossiliferous beds. Both biocoenose and thanatocoenose assemblages of oyster and pecten are present (Fig. 2). *Turritella* and rare coral have been reported (G. Gastil, oral commun., 1978).

Pebble-rich beds of a distinct texture occur in mcg-1 (Fig. 2). A resistant carbonate cement crust has formed on the pebbles and remains unweathered while the interior of the pebble has decomposed. The broken crust reveals the decomposed inner portion of the pebble resulting in the unusual texture of the outcrop.

Up section, the beds become unfossiliferous, member (mcg-2). Individual beds range in thickness from a few centimeters to over a meter. Sorting is very poor. Clast composition remains as that of mcg-1, but with an overall increase in clast size. Cross-bedding occurs in the more sandy facies.

Continuing up section, the return of pectins and oysters marks the beginning of member (mcg-3). The oysters are smaller than in mcg-1 and occur individually. Trace fossils are common and indicate the beds are right side up. Clast composition remains that of mcg-1 and mcg-2. Sorting is very poor with some clasts of boulder size.

Samples from a sandy facies of mcg-3, in arroyo B, contain abundant benthic and pelagic foraminifera along with lesser numbers of ostracods, fish teeth and echinoderm spines. The foraminifera *Lagena*, *Hanzawaia*, and *Elphidium* were identified with a tentative identification of *Globigerinoides obliquies*. The foraminiferal assemblage is representative of Neogene time and a Miocene to early Pliocene age has been suggested by A. D. Warren (oral commun., 1978).

CONCLUSION

An abundance of neritic mollusks, benthic foraminifera, and ostracods alternating with an apparently fluvial section indicates that the conglomerate was deposited in shallow water. The rate of deposition kept pace with and temporarily exceeded the rate of basin subsidence, with a total subsidence of at least 988 meters. Poor sorting and clast composition indicate that the conglomerate was locally derived. Whereas marine strata of Miocene age are reported at many localities beneath the alluviated basins of western Sonora (Lozano, 1975), this is the only known outcrop of marine strata interbedded with Miocene volcanic strata exposed on either side of the Gulf of California.

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THE MARINE MIOCENE FAUNAS OF TIBURON ISLAND, SONORA, MEXICO
AND THEIR ZOOGEOGRAPHIC IMPLICATIONS

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Abstract

The known late Middle Miocene marine fauna of Tiburon Island is listed. Many species proved to be representative of widespread early Pliocene taxa. Others are interpreted as being progenitors of key pectinids lineages which became widespread and underwent adaptive radiation during the Pliocene in the Gulf of California. Striking similarities between the Tiburón Island fauna and Neogene Caribbean taxa suggest active faunal interchange throughout the Neogene between Panamian and Caribbean waters.

Resumen

Se enlista la fauna marina conocida del Mioceno Medio Tardío en la Isla Tiburón. Muchas especies probaron ser representativas de taxa de amplia distribución del Plioceno temprano. Otras especies son interpretadas como progenitores de linajes de pectinidos claves, los cuales se esparcieron y se adaptaron en todo el Golfo de California durante el Plioceno. Similitudes notables entre la fauna de la Isla Tiburón y taxa neogenos del Caribe sugieren un intercambio faunístico activo durante todo el Neogene, entre las provincias Caribeñas y Panámicas.

During the reconnaissance mapping of western Sonora and Tiburon Island, previously unknown Miocene marine strata were discovered (see Gastil 1975 for locations of marine units). Here Gastil and his party collected a small but highly significant collection of marine invertebrates (Table 1). The Miocene section on Tiburon Island is composed of an extremely thick accumulation of volcanic pebble and cobble conglomerates interbedded with arkosic sandstone and volcanic flows. The marine intercalations on Tiburon Island remain of unknown thickness and aerial extent. Both the marine and non-marine beds were considered by Gastil as belonging to the Comondu Group. Fortunately a flow within the marine sequence was dated by Gastil and his associates as 14.0 ± 3.0 m.y.b.p. (Gastil, 1975).

The small collection of specimens represents the oldest known Neogene sequence yielding megafossils from the Gulf of California depression. These species have been described and the pectinids illustrated in an earlier work (Stump, 1979). Due to the vagueries of preservation predominantly calcite-shelled species are better represented.

Therefore, oysters and pectinids dominate the faunal list. Undoubtly additional collecting will yeild additional specimens of these and other taxa. Many of these species were precursors of widespread Pliocene taxa of lasting phylogenic importance. Others appear

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Localities

Bivalvia	S2R-19 (=S2R-18)	S2G-11	B2J-48
<u>Arca</u> (<u>A.</u>) cf. <u>A. pacifica</u>		x	
<u>Ostrea vespertina</u>		x	
<u>O.</u> cf. <u>O. columbiensis</u>		x	
<u>Anomia</u> cf. <u>A. peruviana</u>	x		
<u>Argopecten</u> n.sp.1	x	x	
<u>A.</u> n.sp. 2		x	
<u>Chlamys</u> n. sp. 1	x	x	
<u>C. (Chlamys)</u> n.sp. 4	x	x	
<u>Euvola keepi keepi</u>	x	x	
? <u>Lyropecten</u> sp.indet.	x		
<u>Oppenheimerpecten vodgesi</u>	x	x	
" <u>Pecten</u> " new genus & species			x
Gastropoda			
<u>Strombus</u> (? <u>Strombus</u>) sp.indet*		x	
<u>Turritella imperialis</u>	x	x	
Anthozoa			
<u>Solenastrea fairbanksi</u>		x	

Table 1. The known fauna of the Tiburon Island Miocene.

Locality S2R-19(=S2R-18) is the lowest fossiliferous horizon, S2G-11 is about 1,000 m above S2R-18 (personal communication R.G.Gastil, 1973). *Species preserved as internal molds.

to have become extinct at or about the close of the Miocene. Arca pacifica, Ostrea vespertina, O. columbiensis, Anomia peruviana and Turritella imperialis all have previously been reported from early Pliocene localities around the Gulf of California. These taxa, except for O. vespertina and Turritella imperialis, range into the Recent of Panamic-Pacific waters. Solenastrea fairbanksi, the only known anthozoan from the Tiburon deposits, has likewise been long known from early Pliocene deposits in the Gulf area. This species, plus several additional corals, were first described from the early Pliocene portion of the Imperial Formation, Imperial Valley. Certain members of this faunal association have subsequently been reported from other Gulf localities (see Stump, 1979). All known early Pliocene corals from the Gulf of California display remarkable similarities with several important taxa from the Caribbean Neogene. Likewise, to varying degrees, most or all of the early Neogene pectinids of the Gulf area also share affinities to the Neogene Caribbean. These similarities have convinced many workers of faunal interchange between the Neogene Caribbean and Panamic water masses. Affinities of Gulf Neogene mollusca stressing the pectinids have recently been restated in terms of transpecific evolution (Stump, 1979). The Tiburon Island Miocene marine faunas serve to reinforce and

add badly needed additional insight into the early faunal history of this region.

Tertiary migrations of several groups, notable among which are the Mollusca and Scleractinia, produced faunas of distinctive character along the Pacific basin. The resulting Pacific American faunas were assembled by immigration of northern taxa dispersing south and movement of other genera into the region from the Caribbean. Only a handful of genera appear to be Pacific New World Molluscan endemics, branching from either Caribbean or North Pacific stocks.

Many species of the large chlamytid Lyropecten Conrad (1862) have been described from early Miocene through late Pliocene deposits of western North America. The genus appears to be one of the very few tropical Pacific Neogene endemic pectinids. Species of this genus are common in deposits of southern California south to Baja California. But, Lyropecten is unknown for certain from the Miocene of the Gulf of California. Previous records of L. estrellanus Conrad from the early Pliocene portion of the Imperial Formation are undoubtedly based on adventitious specimens. The few fragments recovered of Lyropecten, a partial hinge plate, from the Tiburon have characters suggesting this genus (see discussion in Stump, 1979). If additional specimens of this taxon are recovered from Tiburon Island and substantiate my identification it will be the first known occurrence of

Lyropecten within the Gulf depression. As stated above Lyropecten is well known from the Pacific side of Baja California and its presence from the Gulf would shed invaluable light on migration paths of this important taxon.

Several workers have speculated on the origins of Lyropecten. Some suggest a Caribbean origin (Smith, 1919) or that its species are present in the Miocene of the Atlantic seaboard (Woodring, 1938). Many authors have also referred several Atlantic-Caribbean species to Lyropecten. It is true that certain of these taxa are superficially similar to Lyropecten s.s., though none are convincing. Even fewer are similar to the type of genus L. estrellanus Conrad (a west coast taxon). Olsson (1922) thought "Pecten (Lyropecten)" tucilla (= "P. (L.) hopkinsi Olsson, 1922 preoccupied) to be lyropectinid. The species was described from the late Miocene (?) Tumbez Formation, Peru. My restudy of the fragmentary types proved them to belong to Leopecten. Specimens indentified as "Pecten (Lyropecten) magnolia Conrad" from the Miocene of Carballo, Costa Rica (Haas, 1942) may also belong to Leopecten. Hertlein and Grant (1972:209) thought this occurrence to be a member of the "Pecten" jeffersonius Say group of species. This suggestion is highly improbable. Members of this group are unknown in the Caribbean at any time and are unreported in the east Pacific Neogene.

Species supposed to belong to the Pecten Jeffersonius group have from time to time been placed in Lyropecten. For example "Chlamys (Lyropecten)" sayanus Dall from the Miocene Oak Grove Formation of Florida is such an allocation (Gardner, 1926). Species also included in "Lyropecten" are taxa similar to the Miocene "Pecten" condylomatus Dall. Pectinids similar to this taxon are widely distributed in the Atlantic Coast Neogene and throughout the Caribbean. But, "P." condylomatus is a Nodipecten Dall (1898) while "C." sayanus has been placed in Chesapecten Ward and Blackwelder (1975).

Authors have compared west American and European pectinids of this general appearance and have suggested that east coast United States taxa are intermediate in this series (for example Grant and Gale, 1931; Clark and Durham, 1942). European workers, while recognizing similarities between east American and European species of this type, have not placed their taxa in Lyropecten (for example Roger, 1939; North in Hertlein and Grant, 1972).

Conrad, who was familiar with the known American species, thought Lyropecten to be restricted to West America. Recent workers tend to conform to his belief (Hertlein in Moore, 1969; Hertlein and Grant, 1972).

The "Pecten" madisonius Say species-group from the Miocene of the eastern United States is also similar

to Lyropecten. Its members are also found in the Miocene of Europe and American species have recently been placed in Chesapecten Ward and Blackwelder, 1975. Details of hinge morphology, patterns of sculpture and their known zoogeography suggest Lyropecten and Chesapecten are probably unrelated lineages.

Other authors (for example Hertlein and Grant, 1972:209) suggest a relationships between members of the "Pecten" magnolia group of West America and "P." jeffersonianus group of east America. The latter group bears considerable resemblance to species of Macrochlamys Sacco of Europe. A relationship between members of the "P." jeffersonianus group and European Neogene Macrochlamys species may exist (compare pl. 27, fig. 3 of Mongrin, 1959 with Ostrea latissima Brocchi, 1814 the type of Macrochlamys Sacco, 1897). But, there is little morphologic or zoogeographic evidence to suppose that members of the "P." jeffersonianus group are related to Lyropecten. Details of hinge morphology cited by Hertlein and Grant (1972) militate against such a relationship. Probably the "Pecten" jeffersonianus group and Macrochlamys spp. have a common ancestor. Lyropecten, however, may be a descendent of late Paleogene or early Neogene Vertipecten stocks which migrated to the west Pacific from northeast Asia.

At present it is impossible to say where the

generic affinities of Lyropecten lie with certainty. The fragmentary specimens from Tiburon Island questionably placed in this genus pushes the geologic range of this taxon into the Miocene of the Gulf. They may also support a Caribbean origin for the group through some unrecognized precursor instead of an origin from North Pacific stocks.

The earliest known species of Oppenheimopecten come from the early Miocene Burdigalian stage of Europe. In the Mediterranean alone there are over a dozen of its species described. The genus is also diverse in the Neogene Caribbean (Stump, 1979) where it appears during the early Miocene. Deposits of middle to late Pliocene age in the New World also have several species of Oppenheimopecten. Its species are also known from the late Cenozoic of the Indo-Pacific region extending north to Japan.

The genus first appears in West Mexico in the Isidro Formation and on Tiburon Island and ranges into the Recent of the region. During the Miocene, the genus was more widely distributed than at present. But only a handful of distinct East Pacific fossil species belong to this genus. Restudy of them, comparisons with the Tiburon species and with growth series of O. vodgesi suggest that O. hartmanni (Hertlein, 1925) and O. heimi (Hertlein, 1925) should be placed in its synonymy. Several valves of O. vodgesi are present in the late Middle Miocene of Tiburon Island thereby documenting its oldest known occurrence in the

Gulf depression.

The most widespread of the middle Miocene Caribbean Oppenheimiopecten was O. bowdenensis (Dall). This taxon ranges from at least the middle Miocene and last occurs in the Pliocene Bowden Formation of Jamaica (Stump, 1979). Several individuals of this species were studied in the Smithsonian collections. Many appear similar to variants of O. voderesi, and it is probable that O. bowdenensis was the root stock of the Panamic species (figure 1).

Species of Euvola Dall, 1898 are widely distributed in New World deposits of late Oligocene to late Pliocene age. The Tiburon Island specimens record their oldest occurrence in the Gulf of California. The two taxa referred to the genus in the Gulf are commonly encountered here in strata of early through late Pliocene age. At the close of the Pliocene the genus became extinct on the West Coast of the Americas. Only one species, E. ziczac (Linne), lives in the Caribbean.

Many species similar to the earliest New World Euvola occur in the late Oligocene to early Miocene strata of Tethyan Europe. The genus first occurs about this same time in the Caribbean and many of its species are found in early through middle Miocene deposits here. By the late middle Miocene a species of Euvola reached the Gulf of California region. This species, E. keepi (Arnold, 1906) and its subspecies E. k.

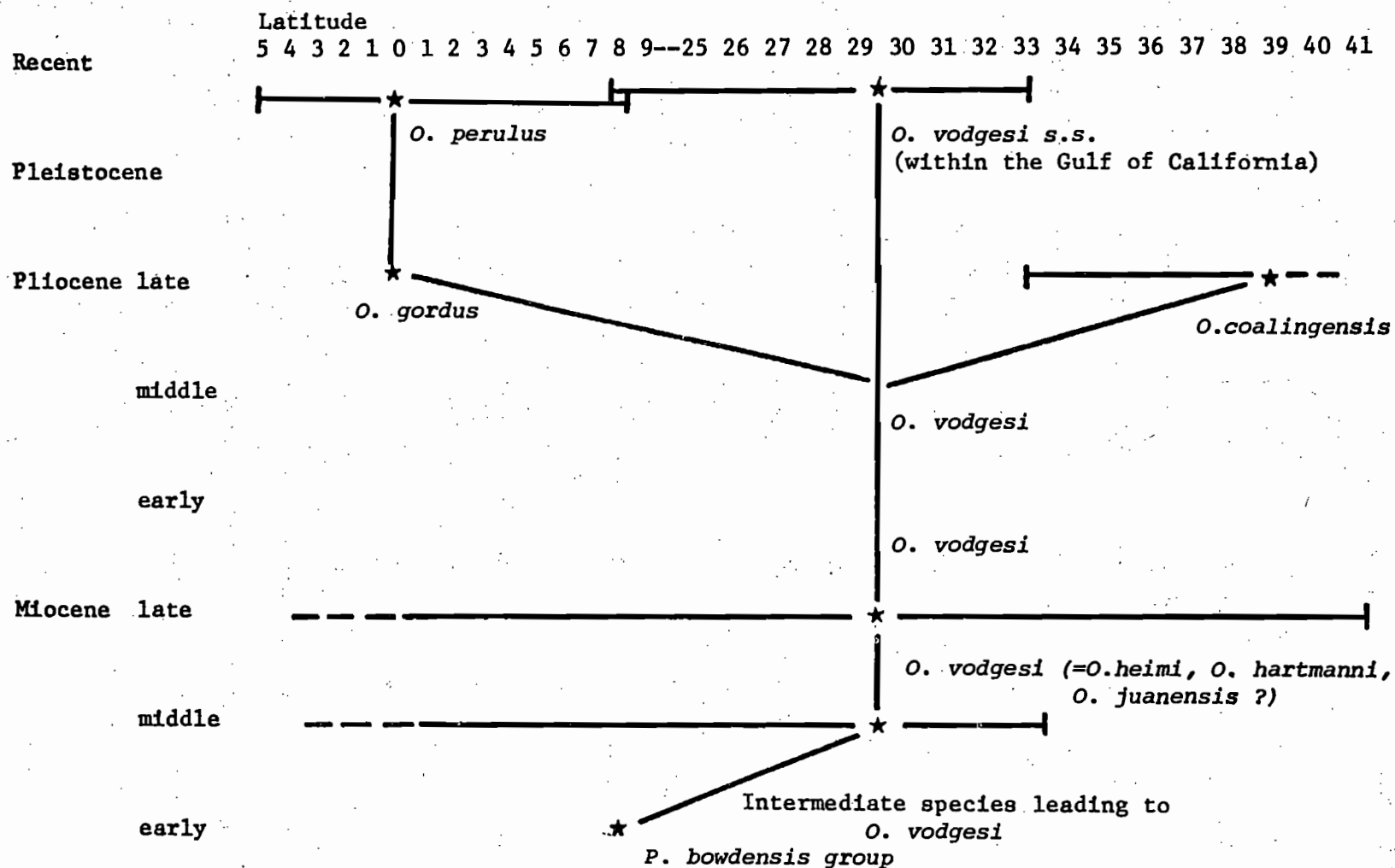


Figure 1 Biogeography and Phylogeny of New World Species of *Oppenheimopecten*.

refugioensis (Hertlein, 1925) flourished in this region from their first appearance until the late Pliocene. At this time both taxa and the genus became extinct in the Pacific Americas.

The oldest known Neogene Mexican Chlamys, C. n. sp. 1, appears in the Isidro Formation and is remarkably similar to the Miocene Caribbean C. buchivacoana F. and H. Hodson (1927). Species resembling the latter taxon first appear in the Oligocene of Venezuela (F. & H. Hodson, 1927).

Occuring with C. n. sp. 1 in the Isidro Formation and on Tiburon Island is C. n. sp. 4. The later species bears a remarkable resemblance to C. muscosa (Wood) from the Neogene and Recent Caribbean. Chlamys n. sp. 4, and other New World Neogene taxa, share characters which unite them with C. muscosa species-group. Likewise, its oldest members in the New World come from the early Miocene of the Caribbean.

Chlamys n. sp. 1, first known from the Isidro Formation, last occurs in the Tiburon Island Miocene. Its probable descendent, C. corteziana (Durham), however, is widespread in the Gulf of California Pliocene deposits (figure 2). The extenction of this species during the late Pliocene ended the long history of the C. muscosa like group of species in the eastern Pacific New World.

The first known species of Argopecten, A. n. sp. 1, in the Gulf of California dates from the late Middle

Miocene. This species and its descendent, A. n. sp. 2, are both found on Tiburon Island. Apparently they are absent in the older Miocene and Oligocene of peninsular Baja California. Argopecten n. sp. 2 occurs several hundred meters stratigraphically above its progenitor (Gastil personal communication, 1976) but both occur in the highest fossiliferous Miocene horizons in their type areas. Neither species is known from younger horizons in the Gulf of California region. The Miocene species and their descendent, A. mendenhalli (figure 3), were probably all broadly adapted species living in coastal waters and ranging into shallow depths. The associated faunas of the Miocene taxa are poorly known and their paleobathymetries are largely conjectural. But, A. mendenhalli is widespread throughout the Gulf in deposits of early to late Pliocene age. It is also associated with many other molluscan taxa most of which are still living. The habitats and bathymetry of the communities A. mendenhalli occurs with suggest such a broadly adapted species.

The genus Argopecten has a long and complex history in the tropical East Pacific. Waller (1969) has described the evolutionary history of the genus in West Atlantic and certain parallel themes in the histories of both exist. Waller showed the West Atlantic species followed three broad evolutionary trends: 1) speciation associated with newly evolved taxa entering

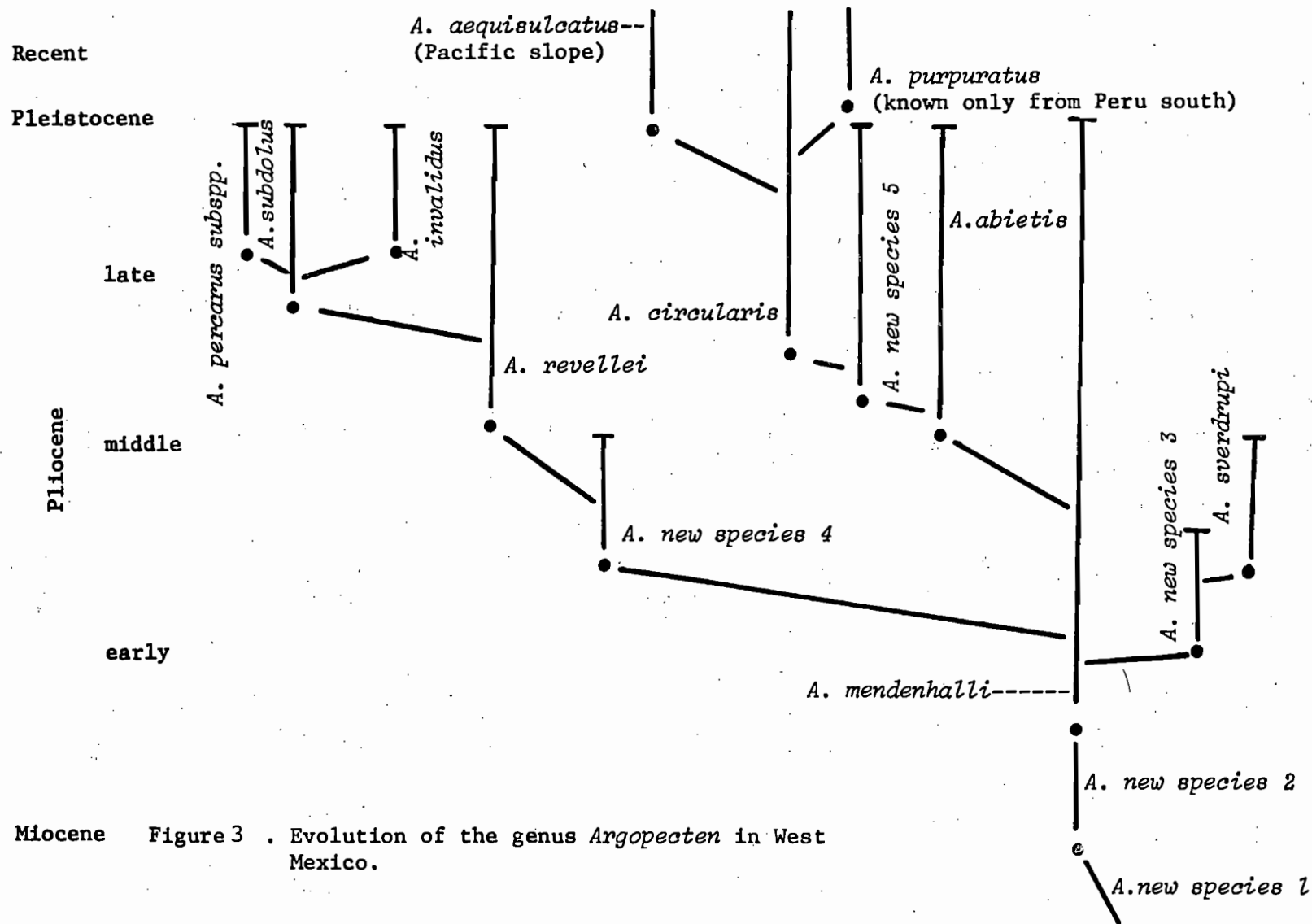


Figure 3 . Evolution of the genus *Argopecten* in West Mexico.

new adaptive zones; 2) evolutionary changes occurring within lineages leading to the "improvement" of the species (lineage anagenesis); 3) changes occurring within two or more lines of descent leading to the improvement of the species group (species-group anagenesis). These are common characteristics in the evolutionary histories of many groups. But, it is possible in the Gulf of California to relate species-successions of the East Pacific argopectinids with their entry into new adaptive zones and the production of allopatric species by the isolation of gene pools. During the Pliocene the Gulf of California argopectinids underwent rapid adaptive radiation resulting from taxa crossing adaptive thresholds and their moving into new habitats. This began not with the first appearance of Argopecten in the region (late middle Miocene) but much later (middle Pliocene). At this time three species of Argopecten first appear. The speciation accelerated through the middle Pliocene with the appearance of four additional taxa. The explosive radiation of the group ended progressively during the late Pliocene. Only one species, A. circularis, survived into the Recent of the tropical East Pacific.

The second and later theme in the diversification of the East Pacific argopectinids was allopatric speciation. Certain species of Gulf of California

Argopecten reached the outer coast of Baja California before the late Pliocene. During the late Pliocene they also underwent adaptive radiation giving rise to several additional taxa. Therefore, by the late Pliocene the outer and inner coasts of Baja California had similar suites of species adapted to similar habitats. Only two of the outer coast species, A. circularis and A. aequisulcatus, survived the Pliocene and are found there in the Recent.

Also present in the Tiburon Island Miocene was one poorly preserved pectinid that resembles no described taxon. Vaguely resembling a Nodipecten the one leached right valve probably is an undescribed genus.

The Tiburon Island Miocene deposits represent a highly significant discovery. When the section is more thoroughly recollected new and better insights will be gained on the degree and timing of trans-American faunal interchange and on the evolution of the Panamic faunas.

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SOME PLEISTOCENE FACIES AND FAUNAS OF COASTAL SONORA AND TIBURON ISLAND, MEXICO

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Abstract

Pleistocene marine faunas from the Puerto Peñasco, Puerto Libertad, El Infiernillo and Tiburon Island are listed.

It is tentatively concluded that the Puerto Libertad and El Infiernillo faunas are approximate time equivalent communities. The former assemblages lived in intertidal sandflats while the latter is a subtidal assemblage. The taxa from Puerto Peñasco may be slightly younger than the previous localities while specimens from Tiburon Island are probably very late Pleistocene.

Resumen

Se enlistan las faunas marinas del periodo Pleistoceno de Puerto Peñasco, Puerto Libertad, El Infiernillo y la Isla Tiburón. Se concluye tentativamente que las de Puerto Libertad y El Infiernillo son comunidades aproximadamente equivalente en tiempo. Las asociaciones mencionadas vivieron en llanuras arenosas de intermarea mientras que las otras corresponden a una asociación infralitoral. Los taxa de Puerto Peñasco pueden ser ligeramente mas juvenes que en las otras localidades, mientras que los especímenes de Isla Tiburón son probablemente del Pleistoceno.

During the National Science Foundation funded mapping of the State of Sonora several previously unknown megafossiliferous Pleistocene localities were discovered. One of these faunas has been described (Stump, 1975). This paper summarizes an additional prolific location and its relationships to previously known coastal Sonora Pleistocene faunas. New data from localities briefly mentioned by Durham (1950) on Tiburon Island are also summarized. Previously Hertlein and Emerson (1956) documented a Pleistocene fauna from the Puerto Penasco area.

Along coastal Sonora several midden localities were discovered that, while having archaeological significance, were not collected for this paper. The major localities summarized below (Table 1) are found in the northern Gulf of California near Puerto Penasco, along the central Sonoran coast at Puerto Libertad and in the El Infiernillo area adjacent to Tiburon Island. Additional localities are also included from Tiburon Island which were briefly discussed by Durham (1950).

No radiometric dates are available from any of these localities. But from stratigraphic considerations, the preservational and composition of their included fossils it is probable that the Puerto Libertad and El Infiernillo deposits are approximately the same age. Based on similar reasoning it is thought that the Tiburon faunas are younger than either the Puerto Libertad or El Infiernillo communities. The assemblage discussed by Hertlein and Emerson (1956) may also be slightly younger than the Puerto Libertad-El Infiernillo faunas.

The Pleistocene molluscs recovered from Tiburon Island were collected at several discrete localities along the southernmost portion of the island and its western shore. All of these localities are very near the present shoreline and on low terraces. Higher terraces proved to be mostly unfossiliferous. All species recovered from these deposits presently live in adjacent waters. Deposits from the Puerto Libertad and El Infiernillo areas yielded the extinct Encope grandis inezana Durham. This species was regarded by Durham (1950) as aguide to deposits of Pleistocene age. Also present in either the Puerto Libertad or El Infiernillo deposits were a handful of species presently living further south in the Gulf of California. This may point to the greater age of the Puerto Libertad and El Infiernillo deposits when compared to the Tiburon Island or Puerto Penasco Pleistocene.

The faunas from Puerto Penasco and Tiburon Island compose a mixed assemblage of intertidal and subtidal sandy-bottom and rock-dwelling organisms. The Puerto Libertad

assemblage was considered to have been preserved mostly in situ (Stump, 1975) and its fauna lived on or within intertidal sand flats. The previously unreported faunas from the El Infiernillo area also undoubtedly are mostly preserved in situ, but clearly are a subtidal assemblage. The El Infiernillo faunas represent the deepest water Pleistocene community documented to date from western Sonora.

The general stratigraphy of the Puerto Penasco deposits was described by Hertlein and Emerson (1956). Here conglomeratic sandstones rest on basaltic or granitic basement. The inclosed faunas were collected from the lowest terrace which was capped by a thin superficial Pleistocene cover. The stratigraphy of the Puerto Libertad area was treated earlier (Stump, 1975). These units are also relatively thin and are composed of approximately seven meters of interbedded silty sandstone with local lenses of conglomerate. The Puerto Libertad units rest on an unfossiliferous boulder to cobble-bearing, medium-grained sandstone. The most abundantly fossiliferous horizon in the Libertad area occurs in approximate midsection and is only about 60 cm thick.

The El Infiernillo deposits are found approximately 70 km north of Kino Bay and compose the west facing bluffs adjacent to Tiburon Island. The Pleistocene section here has an exposed thickness of about 12 meters. The youngest preserved portion of the deposit is composed of about five meters of tan silty sandstone which is capped by alluvium and midden material. Its base is gradational onto about a two meter-thick pebbly sandstone which contains scattered intertidal molluscs. Below this sequence rest approximately three meters of abundantly fossiliferous, well sorted, light brown sandstone. Local cobble lenses are present in this unit and its fauna is dominated by subtidal species. This horizon was collected and summarized on Table 1. The subtidal sequence rests on an exposed thickness of about two meters of brown, medium-grained, feldspathic sandstone. The base of the Pleistocene sequence in the Infiernillo area

is not exposed and its lowest exposed horizons were unfossiliferous.

Durham (1950) mentions some Pleistocene localities from the southern portion of Tiburon Island. During the mapping of this portion of the island his localities plus others were collected. The Pleistocene sediments here rest in the lowest terrace of the region. Fossils were recovered from tuffaceous and arkosic sandstone which rest on Miocene volcanics or on conglomerates erosionally derived from them. Both the flows and the volcanogenic sediments were considered to belong to the Comondu Group. Additional localities were discovered on the western portion of the island in terraces of about three meters elevation above the present shoreline. These fossils also were inclosed in similar sediments and rest on a terrace cut into the Comondu Group.

Further exploration of the western coast of the Gulf of California will undoubtedly yield additional Pleistocene and perhaps older faunas. Badly needed radiometric dates on these units would further help clarify facies relationships that can now only be guessed at.

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Table 1. The known Pleistocene faunas of coastal Sonora and Tiburon Island.

	Pleistocene				Recent Coastal Sonora
	Puerto Penasco	Puerto Libertad	El Infierno	Isla Tiburon	
<u>Bivalvia</u>					
Nucula declivis			x		x
N. exigua			x		x
Nuculana acrita			x		x
Arca pacifica		x		x	x
Barbatia gradata	x	x	x		x
B.reeveana		x			x
Anadara formosa		x			x
A. grandis				x	x
A. multicostrata		x		x	x
Arcopsis solida	x	x	x		x
Glycymeris gigantea		x	x	x	?
G. multicostrata	x	x			x
G. maculata	x	x			x
G.strigilla			x		Guaymas
Brachidontes adamsianus	x				x
B. semilaevis	x				x
Septifer sp.indet.		x			x
Crenela divaricata			x		x
Modiolus capax	x	x	x	x	x
Atrina tuberculosa		x			x
Isognomon janus			x	x	x
I. recognitus	x				x
Ostrea angelica	x	x	x		x
O. columbiensis		x			x
O. conchaphila				x	x
O. palmula	x	x			x
O. megodon		x			x
Pecten vodgesi	x	x	x	x	x
Argopecten circularis	x	x	x	x	x
Leptopecten tumbezensis		x			x
Nodipecten subnodosus		x		x	x

(Table cont'd)

	Pleistocene				Recent Coastal Sonora
	Puerto Penasco	Puerto Libertad	Ei Infierno	Isla Tiburón	
<i>Plicatula</i> sp.indet		x			x
<i>Spondylus calcifer</i>		x			x
<i>Lima orbigni</i>			x		x
<i>Anomia peruviana</i>	x	x	x	x	x
<i>Crassinella pacifica</i>			x		x
<i>Cardita affinis</i>	x	x	x	x	x
<i>C. laticostata</i>			x		x
<i>Lucina approximata</i>		x	x		x
<i>L. centrifuga</i>			x		x
<i>L. excavata</i>		x			x
<i>L. fenestrata</i>			x		x
<i>L. lampra</i>	x		x		x
<i>L. mazatlanica</i>		x			x
<i>Codakia distinguenda</i>	x	x		x	x
<i>Ctena mexicana</i>		x	x		x
<i>Divalinga eburnea</i>		x			?
<i>Diplodonta suprema</i>			x		Panama
<i>Felaniella sericata</i>		x			x
<i>Amerycina colpoica</i>		x			x
<i>A. cultrata</i>		x			?
<i>Aligena cokeri</i>			x		x
<i>Mysella compressa</i>		x			x
<i>Orobitella trigonalis</i>		x	x		?
<i>O. secura</i>			x		?
<i>Ensitellops hertleini</i>			x		x
<i>Basterotia hertleini</i>			x		x
<i>Sportella stearnsii</i>			x		x
<i>Chama echinata</i>		x			x
<i>C. frondosa</i>	x	x			x
<i>C. mexicana</i>		x			x
<i>Pseudochama corrugata</i>		x		x	x

(Table 1 cont'd)

	Puerto Penasco	Puerto Libertad	El Infiernillo	Isla Tiburón	Recent Coastal Sonora
<i>Trachycardium panamense</i>		x	x		x
<i>T. procerum</i>		x			Oaxaca
<i>T. senticosum</i>		x			x
<i>Trigoniocardia granifera</i>		x	x		x
<i>T. biangulata</i>		x			x
<i>T. guanacastensis</i>		x	x		Cabo San Lucas
<i>T. obovalis</i>		x			x
<i>Laevicardium elatum</i>	x	x	x	x	x
<i>L. elenense</i>		x	x		x
<i>Periglypta multicostata</i>				x	x
<i>Ventricolaria isocardia</i>		x		x	x
<i>Tivela byronensis</i>		x			x
<i>Pitar concinnus</i>			x		x
<i>P. pollicaris</i>		x			x
<i>P. paytensis</i>			x		x
<i>Megapitaria squalida</i>	x	x	x	x	x
<i>Dosinia ponderosa</i>	x	x	x	x	x
<i>Cyclinella saccata</i>		x	x		x
<i>Chione californiensis</i>	x	x	x		x
<i>C. fluctifraga</i>		x			x
<i>C. gnidia</i>	x	x			x
<i>C. tumens</i>		x			x
<i>Protothaca grata</i>	x	x	x		x
<i>Mactra angusta</i>			x		x
<i>M. californica</i>		x			x
<i>M. nasuta</i>		x			x
<i>Raeta undulata</i>		x	x		x
<i>Tellina carpenteri</i>		x	x		x
<i>T. meropsis</i>			x		x
<i>T. ochracea</i>			x		x
<i>T. reclusa</i>		x			x

(Table I cont'd)

	Pleistocene				Recent
	Puerto	Puerto	El	Isla	Coastal
	Penasco	Libertad	Infiernillo	Tiburón	Sonora
<i>Tellina simulans</i>		x	x		x
<i>Florimetus cognata</i>	x		x		x
<i>Psammotreta viridotincta</i>		x			x
<i>Strigilla dichotoma</i>			x		x
<i>Donax californicus</i>			x		x
<i>D. gracilis</i>	x		x		x
<i>Tellidora burneti</i>		x	x		x
<i>Tagelus californicus</i>	x	x	x		x
<i>Semele bicolor</i>		x			x
<i>S. flavescens</i>	x				?
<i>S. guaymasensis</i>	x		x		x
<i>S. junonia</i>			x		Guaymas
<i>Ensis californicus</i>			x		x
<i>Cryptomya californica</i>			x		x
<i>Corbula biradata</i>		x	x		Guaymas
<i>C. bicarinata</i>			x		x
<i>C. nasuta</i>	x				x
<i>Gastrochaena ovata</i>			x		x
<i>Hiatella arctica</i>		x			x
<i>Cytopleura subtruncata</i>		x			x
<i>Martesia</i> sp. indet.		x			x
<i>cyathodonta dubiosa</i>			x		x
<u>Gastropoda</u>					
<i>Diodora alta</i>	x		x		x
<i>D. inaequalis</i>			x		x
<i>Collisella dalliana</i>			x		x
<i>C. turveri</i>			x		x
<i>Scurria mesoleuca</i>		x			x
<i>Calliostoma bonita</i>			x		Mazatlan
<i>Tegula globulus</i>		x			Topolobampo

(Table 1 cont'd)

	Pleistocene				Recent Coastal
	Puerto Penasco	Puerto Libertad	El Infiernillo	Isla Tiburon	Sonora
<i>Tegula mariana</i>	x		x		x
<i>T. rubroflammulata</i>	x				x
<i>T. rugosa</i>	x	x	x		x
<i>Parviturbo</i> sp.indet.		x			x
<i>Arene fricki</i>			x		x
<i>Turbo fluctuosus</i>	x	x	x	x	x
<i>Nerita funiculata</i>		x			x
<i>N. scabricosta</i>	x	x		x	x
<i>Neritina latissima</i>		x			x
<i>Theodoxus luteofasciatus</i>		x	x		x
<i>Littorina aspera</i>		x			x
<i>L. varia</i>			x		?
<i>Turritella goniostoma</i>	x		x		x
<i>T. leucostoma</i>			x		x
<i>Tripsyche centriquadra</i>		x			x
<i>Cerithium stercusmuscarum</i>		x	x		x
<i>Liocerithium judithae</i>		x			x
<i>Selia assimolata</i>		x			x
<i>Triphora hannai</i>		x			x
<i>Cerithidea mazatlanica</i>		x	x		x
<i>Rhinocoryne humboldti</i>	x	x			x
<i>Strombus galeatus</i>	x			x	x
<i>S. gracilior</i>		x		x	x
<i>S. granulatus</i>		x		x	x
<i>Epitonium acapulcanum</i>			x		x
<i>E. sp.indet.</i>		x			?
<i>Balcis mexicana</i>		x			x
<i>Crepidula onyx</i>		x	x		x
<i>Crucibulum spinosum</i>		x	x		x
<i>C. scutellatum</i>		x		x	x
<i>Natica chemnitzii</i>		x			x

(Table 1 cont'd)

	Puerto Penasco	Puerto Libertad	El Infiernillo	Isla Tiburón	Recent Coastal Sonora
<i>Polinices bifasciatus</i>				x	x
<i>P. reclusianus</i>	x	x	x	x	x
<i>P. uber</i>	x	x			x
<i>Trivia solandri</i>				x	x
<i>Erato columbella</i>		x			x
<i>Cypraea annettae</i> <i>annettae</i>	x	x		x	x
<i>Malea ringens</i>		x			x
<i>Cassis coarctata</i>		x		x	x
<i>Ficus ventricosa</i>		x		x	x
<i>Murex elenensis</i>			x		x
<i>M. (M.) sp. indet.</i>		x			?
<i>Hexaplex erythrostomus</i>	x	x		x	x
<i>Murexiella vittata</i>			x		x
<i>Muricanthus nigrinus</i>	x	x	x	x	x
<i>Coralliophila parva</i>			x		x
<i>Pteropurpura erinaceoides</i>			x		x
<i>Thais biserialis</i>	x			x	x
<i>Acanthina angelica</i>	x	x	x		x
<i>A. tyrianthina</i>		x			x
<i>Neorapana tuberculata</i>	x	x	x	x	x
<i>N. cf. N. muricata</i>		x			Guaymas
<i>Solenostiera anomala</i>		x	x	x	x
<i>S. pallida</i>	x				x
<i>Strombina angularis</i>			x		x
<i>S. maculosa</i>		x			x
<i>S. solidula</i>			x		x
<i>Columbella fuscata</i>		x			x
<i>Melongena patula</i>		x			x
<i>Nassarius moestus</i>	x				x
<i>N. tiarula</i>		x	x		x
<i>Fusinus dupetithousarsi</i>		x		x	x

(Table 1 cont'd)

	Puerto Penasco	Puerto Libertad	El Infiernillo	Isla Tiburon	Recent Coastal Sonora
<i>Fusinus ambustus</i>			x		Mazatlan
<i>Harpa crenata</i>				x	South Gulf
<i>Oliva spicata</i>	x	x	x	x	x
<i>O. incrassata</i>	x	x		x	x
<i>Olivella dama</i>	x	x	x		x
<i>O. gracilis</i>			x		Guaymas
<i>Vasum caestus</i>		x			Guaymas
<i>Cancellaria cassidiformis</i>				x	x
<i>Conus brunneus</i>		x			x
<i>C. princeps</i>		x		x	x
<i>C. purpurascens</i>				x	x
<i>C. regularis</i>	x	x	x		x
<i>C. ximenes</i>		x		x	x
<i>Terebra formosa</i>		x			Mazanillo
<i>T. albocincta</i>		x			Mazatlan
<i>T. hindsii</i>			x		x
<i>T. variegata</i>		x	x	x	x
<i>Hormospira maculosa</i>			x		x
<i>Pyramidella mazatlanica</i>		x	x		x
<i>Bulla gouldiana</i>		x	x		x
<i>Acteocina angustior</i>			x		x
<i>A. inculta</i>		x			x
<i>Melampus olivacea</i>		x			x
<u>Polyplacophora</u>					
<i>Stenoplax magdalenensis</i>		x			x
<u>Scaphopoda</u>					
<i>Dentallium pretiosum berryi</i>			x		x
<i>Cadulus cf. C. fusiformis</i>		x			x

(Table 1, cont'd)

	Puerto Penasco	Puerto Libertad	El Infiernillo	Isla Tiburón	Recent Coastal Sonora
<u>Anthozoa</u>					
Astrangia haimei		x			x
Porites californica			x	x	x
<u>Bryozoa</u>					
Schizoporella unicornis		x			x
Hippopodinella adpressa		x			x
Antropora tinctoria		x			x
Hippuporida janthina		x			x
Conopeum commensale		x			x
<u>Cirripedia</u>					
Balanus sp. A		x			?
Balanus sp. B		x			?
Balanus sp. C		x			?
<u>Echinoidea</u>					
Encope grandis inezana		x	x		extinct
E. g. grandis			x		x

RECENT INVESTIGATIONS ON QUATERNARY GEOLOGY OF THE COAST OF CENTRAL SONORA, MEXICO

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INTRODUCTION

Between 1965 and 1980, Quaternary deposits of the coast of Sonora, between Puerto Lobos (30°15') and Guaymas (28°N) have been investigated from different points of view. This new interest in the recent evolution of this area, follows the important geologic, geophysical and oceanographic research accomplished in the Gulf of California around the mid-century.

The aim of this paper is to briefly present some aspects of these recent investigations and give a general idea of the state of knowledge in the field of Quaternary geology in the western Sonora coast. On- going research works will also be mentioned. With the exception of a few PhD dissertations, and other works, from U. S. scientists, a large part of the recent progresses were completed through a Franco-Mexican cooperative program (between O. R. S. T. O. M. and Instituto de Geología, U. N. A. M.) which began in 1974 (the publications related to this program, generally were written in Spanish, or in French, and not in English).

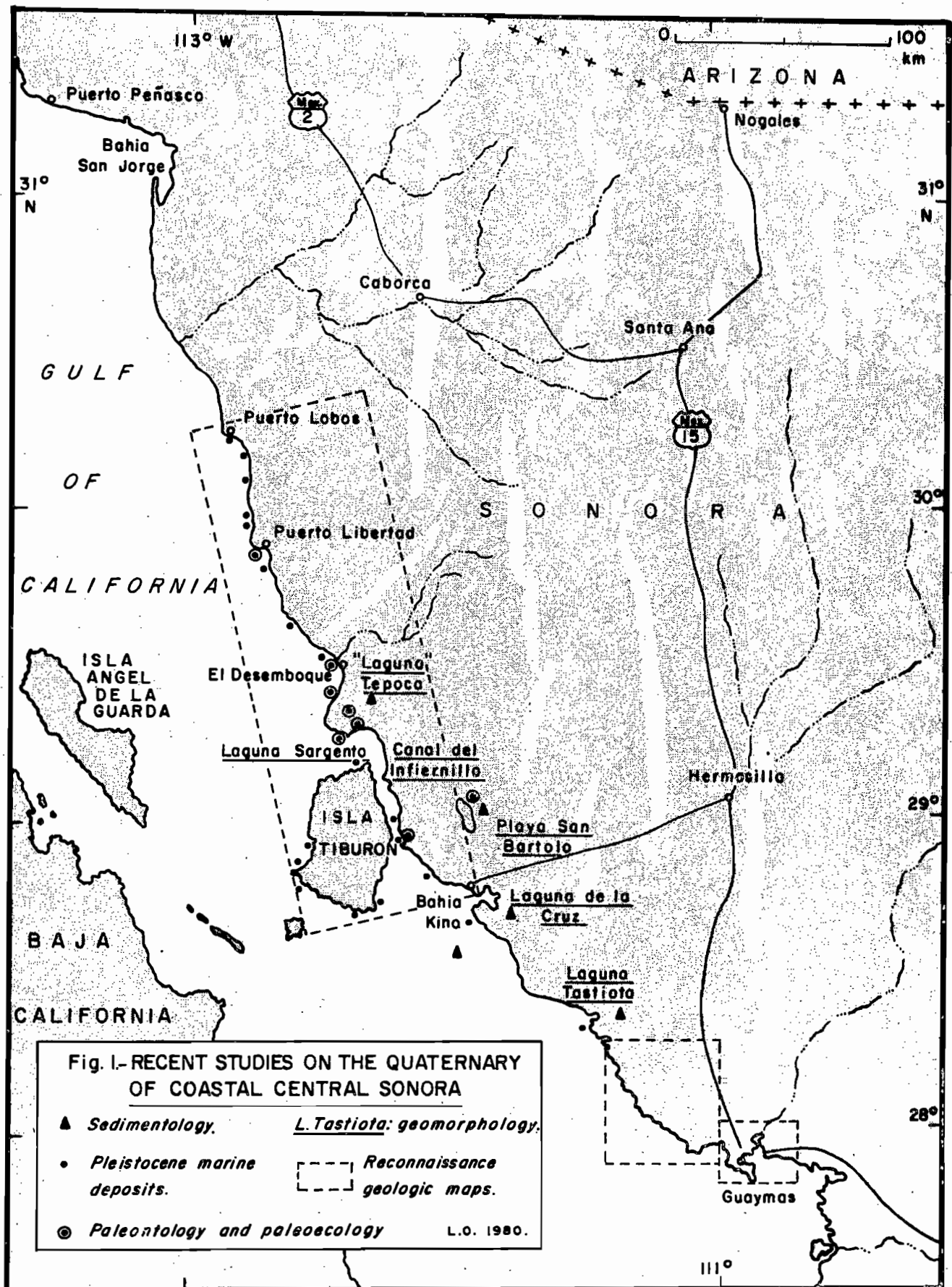
REGIONAL GEOLOGY AND RECONNAISSANCE MAPPING

In the early sixties, the geology of coastal Sonora was so poorly studied that Allison (1964) could write: "So little is known of rocks younger than Cretaceous that conclusions concerning the late geological history of land areas bordering the eastern side of the Gulf of California are hardly meaningful". At the time, for example, the only available geological map was at a scale of - 1: 2,000,000 (Mejorada, 1956).

In 1974 the first authoritative reconnaissance map of the portion of the coast between Puerto Lobos and Bahía Kino (Gastil and others) appeared (Fig.1). This map, at a scale of 1: 150,000, was based on original field mapping, photo-interpretation, and a series of K-Ar dates on Mesozoic and Tertiary plutonic and volcanic units. The southern part of the area treated in this report, around Guaymas (Fig.1), was mapped by Johnpeer (1977) and Wilson (1978).

In 1976, López Ramos compiled a geological map of Sonora which included, for the coastal area, the basic information of Gastil and others (1974). In a short time, Detenal is going to publish a set of geological maps, at a scale of - 1: 250,000, covering the whole State of Sonora. In addition, Roldán, at the Instituto de Geología U.N.A.M., is preparing another compilation of the geology of Sonora at a scale of 1: 500,000.

In the map of Gastil and others (1974), the Quaternary deposits are divided in: alluvium, sand, dissected fluvial deposits and marine. As is usual in this -



kind of reconnaissance map, no stratigraphic distinctions are made. In the explanatory note for their map, Gastil and Krummenacher (1977) emphasize the importance of the alluvial processes taking place in the wide coastal plains of central Sonora. The same authors report several localities where Quaternary alluvial, or fluvial, sediments are cut by faults. In coastal Sonora, the Quaternary faults are generally oriented NW-SE to N-S and belong to the Basin and Range system; in Isla Tiburón, a few Quaternary faults are oriented NE-SW, probably because they are related to the Gulf of California recent structural history. Up to now, no clear evidence of Quaternary strike-slip deformations have been found in this portion of Sonora (Gastil and Krummenacher, 1977; Colletta, oral communication).

The regional vertical comportment of the coastal region of Sonora during the Quaternary is studied through Pleistocene marine terraces data. In some previous works (Ives, 1951, 1963; Nichols, 1965; Richards, 1973), the coastal area of Sonora was considered tectonically active and recently uplifted. According to Gastil and Krummenacher (1977) only the coast north of Puerto Libertad appears emergent while south of Bahía Kino it appears submergent. Ortlieb (1978a, 1978b) considers that, on a regional scale, the coast of Sonora has been "stable", at least during Late Pleistocene and Holocene time; the only area which apparently suffered subsidence is the region of Guaymas. More detail on this question will be given below, in the chapter concerning the interpretation of the Quaternary marine deposits.

OCEANOGRAPHY, GEOMORPHOLOGY AND SEDIMENTOLOGY

Oceanography

The AAPG Memoir on "Marine geology of the Gulf of California", published in 1964 (Van Andel and Shor, eds), still constitutes a main source of information about the areas off the Sonoran coast. Of particular interest in this book, for Quaternarists, are the papers on recent marine sediments (Van Andel, 1964), on physical oceanography (Roden, 1964), on the late Quaternary history of the coast of Nayarit (Curry and Moore, 1964), on macroinvertebrates distribution (Parker, 1964) and on benthonic foraminiferas (Phleger, 1964).

Very few specific investigations were carried on in the nearshore area of the coast off central Sonora. Among them are to be mentioned the work of Curry (1961) on bathymetry and surface lithology of the continental shelf west of the Tastiota-Bahía Kino coast, and the study by Merifield and others (1970) of oceanographic processes in the Canal del Infiernillo, between Isla Tiburón and the mainland.

Geomorphology

In geomorphology two areas of the coastal central Sonora have been studied, with some detail, during recent years. The first one is Playa San Bartolo (sometimes also called "Playa, or Laguna, Noriega"), located east of the southern part of Sierra Seri, and the second one is the Canal del Infiernillo and its surrounding shores.

Playa San Bartolo is the largest dry lake of Sonora, about 13 Km long and 4 km wide; it is located 10 km from the seashore. As its elevation is only + 5 m above present mean sea level (MSL), it was sometimes interpreted as being of marine origin (Sherwin, 1971). Its western margin is limited by the piedmont (bajada) of the Sierra Seri, whereas the eastern shore of the playa is bordered by dune-like sandy bodies (probably of both eolian and fluvial origin). Geomorphic aspects of Playa San Bartolo were studied by Petit-Maire and Lancin (1976), Petit-Maire and Casta (1977) and Lecolle and others (1977).

The shores of the Canal del Infiernillo, and particularly the spits that recently grew on the Sonora and Tiburón side of the strait, have been investiga-

ted by Sherwin (1971) and Lancin (1979, in press). The geometry, the sedimentology and relative distribution of the 14 sandy spits were related to the hydrodynamic conditions of this area (relatively isolated from the Gulf of California) and were used to reconstruct the chronological evolution of these littoral features (Lancin, 1979, in press).

Another geomorphic peculiarity of western Sonora, specially where the coastline is backed by low lying coastal plains, is the extensive barrier dunes system. The morphology and the sedimentology of these Holocene coastal dunes are presently under study in the area comprised between Tastiota and Bahía San Jorge (Crespo, in preparation); only in a few localities are there older fixed ^{dunes} probably of early Late Pleistocene age.

Sedimentology

A wide variety of Quaternary deposits and environments of the sonoran coastal region have been, or are being, studied from a sedimentological point of view.

First investigated were the coastal lagoons, and particularly the Laguna (Estero) Tastiota (Nichols, 1961, 1962, 1965). Coastal lagoons, generally bordered by mangroves, are numerous in the area of the Canal del Infiernillo, but the widest ones are located at Bahía Kino and Tastiota. These two lagoons evolved from estuaries, at the beginning of the Holocene, to open and later restricted, marine lagoons, when the sea level reached its present elevation (around 5,000 years ago?) and as the coastal barriers were building up. During the last few thousand years the lagoonal area has diminished as a result of alluviation and sedimentary deposition.

The sedimentology of inland lacustrine deposits was described, in Playa San Bartolo, by Lecolle and Villaseñor (1976), Lecolle and others (1977), Petit-Maire and Casta (1977) and Lecolle (1979). The sediments of the playa, studied from the surface to a maximum depth of seven meters are predominantly clayey silts and silty clays and include an important fraction of montmorillonite and analcime, as well as some evaporitic minerals (halite and trona). This classical clay-pan system developed during a more humid period when it was provided with fine grained sediments, by sheet flood on the bajada of the Sierra Seri, west of the playa.

The littoral sandy formations and dunes throughout the area have been examined briefly, primarily to determine their chronologic relation to late Pleistocene and Holocene marine and alluvial deposits (Lecolle and Ortlieb, 1978; Lecolle and others, 1978; Ortlieb and Malpica, 1978). The sedimentology of the Holocene coastal dunes are presently being studied by Crespo (doctoral thesis at U.N.-A. M.).

Quaternary alluvial sediments, despite their wide distribution in western Sonora, were not specifically investigated. In the early sixties, Curray (1961) produced a preliminary map of the surface lithology of the coastal plain between Bahía Kino and Tastiota. Some data of grain-size composition of bajada deposits, from various sonoran localities, were given by Ortlieb and Malpica (1978).

Pleistocene marine deposits were extensively surveyed on the mainland coast and around Isla Tiburón. Sedimentologic descriptions of emerged terrace sediments were included in numerous papers and thesis (Stump, 1975; Celis, 1975; Chávez, 1975; Malpica and Ortlieb, 1976; Malpica and others, 1978; Ortlieb and Malpica, 1978; Celis and Malpica, in preparation; Luna, in preparation; González in preparation). Sedimentologic aspects of a Late Pleistocene marine embayment, at "Laguna" Tepoca, have been studied by Lecolle and Ortlieb (1978), Lecolle (1979, in press, in preparation).

PALEONTOLOGY

In the coastal region of Sonora, Quaternary marine deposits have yielded much more material to paleontologists than the terrestrial sediments. Among the relatively widespread outcrops of Quaternary marine deposits (Fig.2), a few fossiliferous localities have been investigated with some detail.

The first paleoecological study of Pleistocene marine mollusks, in this area, was provided by Stump (1975) near Puerto Libertad. In the same locality, Celis (1975) and Celis and Malpica (in press) examined the microfauna.

Another outcrop of Pleistocene infralittoral deposits, located between Punta Chueca and Punta Onah, yielded a large fossil assemblage, with most of the organisms in situ (Petit-Maire, 1976). The determination of these fossils, including the microfauna, was done by Chávez (1975) and Celis (1975, 1979). In this book, Stump reports a list of fossil material, collected a few years ago, in various Pleistocene outcrops in the Canal del Infiernillo, Isla Tiburón, Puerto Libertad and Puerto Peñasco.

Two thesis at U. N. A. M. devoted to the paleoecologic interpretation of four distinct Pleistocene littoral deposits, within the area of El Desemboque-Tepoca, are to be completed by early 1981 (Luna and González in press; González, in preparation; Luna, in preparation).

The microfauna of Holocene deposits, as well as its present distribution in Laguna Tastiota, was studied by Nichols (1965).

A general study of the fossil fauna collected by Ortlieb and collaborators in the localities of Pleistocene marine deposits mentioned in Fig. 2, is scheduled for 1981. Besides the paleontologic determinations and the paleoenvironmental interpretations, a series of isotopic measurements ($^{18}\text{O} / ^{16}\text{O}$ and $^{13}\text{C} / ^{12}\text{C}$) will be made on carbonates of fossil shells, in a attempt to reconstruct paleotemperatures of the past high stands of sea level registered on this coast.

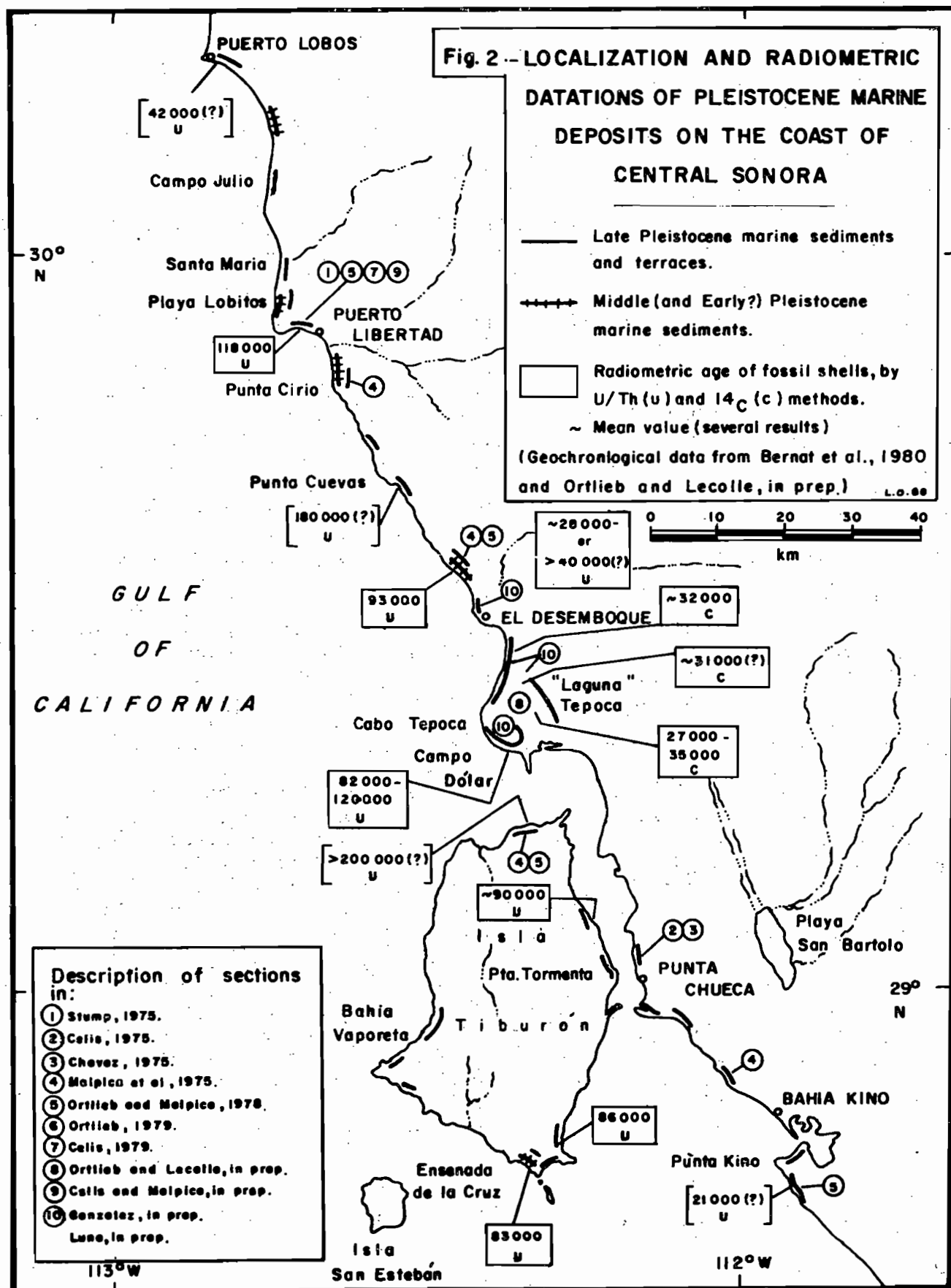
To our knowledge, the terrestrial fauna reported on the sonoran coast is essentially of Holocene age, and is generally associated with human occupation. No Early, or Middle, Pleistocene fossil vertebrates, like those found in eastern Sonora, have been described along the coast; this is due, in part, to the fact that the outcropping continental Quaternary sediments in western Sonora are mainly younger, and less dissected, than in the foothills of the Sierra Madre Occidental. Numerous archeological sites, between El Desemboque and Tastiota, (Bowen, 1976), provided an abundant fauna of Vertebrates which remains to be published (White, oral communication, and Braniff, oral communication). On the margins of Playa San Bartolo, were found fragmented bones and teeth of Elephas sp., Equus (?) sp., Bison sp., and of a Camelidae which was dated $7\ 630 \pm 460$ years B. P., by ^{14}C (Petit-Maire and Casta, 1977).

Very little is known of the paleobotany of this region. The present phytogeography was described by Shreve and Wiggins (1964), Felger (1966), Sherwin (1971), and Hastings and others (1972). Generally speaking, Holocene and late Quaternary hydrologic and sedimentologic conditions of coastal Sonora have not been suitable for the conservation of pollens. Only one section, in Laguna Sargento area, has been studied for its pollen content (L. González, in preparation).

SURVEY OF PLEISTOCENE MARINE TERRACES AND ASSOCIATED DEPOSITS

Reconnaissance study

The study of the deposits and coastal features related to Quaternary marine transgressions has been one of the principal themes of research during recent years in this region.



In 1973, the first ^{14}C dates of Late Pleistocene shells were published by Richards. These shells were collected in two localities, one north and the other south of El Desemboque. In 1975 Stump studied the fossiliferous intertidal deposits outcropping a few meters above MSL in the area of Puerto Libertad. He interpreted the age of the deposits as late Pleistocene.

In 1976-78 Malpica, Ortlieb, and collaborators published several papers and abstracts on their general reconnaissance of Pleistocene marine remnants, all along the coast north of Guaymas. The first two abstracts (Malpica and Ortlieb, 1976; Ortlieb and Malpica, 1977) emphasized that marine terraces of the sonoran coast were low-lying (elevations inferior to + 10 m) and probably corresponded, one to the Late Pleistocene (Sangamon interglacial), and the other one to an older Pleistocene (?) transgression. The maximum Holocene sea level was reported to have been, at most, only a few decimeters above present datum. The paper of Malpica and others (1978) gave a descriptive list of 23 distinct outcrops attributed to the Sangamon interglacial, in coastal Sonora and in Isla Tiburón. In an other paper, Ortlieb and Malpica (1978), provided more data of the same localities and discussed the question of the chronological framework of these deposits.

Ortlieb (1978a, 1979) compared these data from the sonoran coast with the terraces found on the Pacific and the Gulf coast of Baja California to draw conclusions on the distinct rates of recent vertical deformations in the latter area.

Geochronology

After unsuccessful results of dating by amino-acids geochronology, U/Th dates of fossil shells (*Dosinia ponderosa*), samples in different localities between Punta Kino and Puerto Lobos, were published by Bernat and others (1980). These radiometric ages generally confirm the attribution to the Sangamon interglacial (120 000 to 80 000 years B. P.) of the majority of the sampled outcrops (See Fig. 2). A few U/Th dates are interpreted as resulting from uncontrolled contaminations (dates of 42 000? and 21 000 ? years B. P.). Two more dates of 180 000 ? and 200 000 ? years B. P., argue for a but last interglacial age (pre- Sangamon), in two localities mentioned in Fig. 2, but field interpretation suggests rather a Sangamonian age.

In the area of Tepoca, where ^{14}C ages or around 30 000 years B. P. had been first reported by Richards (1973), further field work and sampling of marine fossil shells were done to check the hypothesis of a marine transgression during a late Wisconsin interstadial (Lecolle and Ortlieb, 1978; Lecolle and others, 1978). Twelve more dates from "Laguna" Tepoca provided apparent ^{14}C ages comprised between 27 000 and 35 000 years B. P. (Ortlieb and Lecolle, in preparation); precise isotopic and mineralogic analysis of the dated material are still necessary to validate these results (which might be due to only 3% contamination by modern carbon). The question of a "eustatic" high stand of sea level (near or above present MSL), at about 30 000 years B. P., has been largely debated by many investigators (see discussion on this controversy in Mörner, 1971 and Thom, 1974).

Tectonic significance

Field observations and presently available radiometric and paleontological data of almost all the outcrops of Quaternary marine deposits found on this coast, at a few meter elevation, suggest a Sangamonian age. The reconstructed paleo-sea level of the last interglacial in coastal Sonora is about + 5 m (above present MSL). As there are very few significant differences in the present elevations of the remains of the Sangamonian paleo-shoreline, it is inferred that the whole coastal area has not been affected by late Quaternary vertical movements; furthermore, as the altitude of the paleoshoreline correspond to the level presumably reached during the Sangamon (as determined worldwide) this coastal region of Sonora may be consi-

dered as tectonically stable, at least during the last 100 000 years. In the area of Guaymas the lack of marine terraces and the typically drowned valleys (peculiarly visible in the main bay) indicate that this region has been subsident in the Late Quaternary.

In general, the Middle Pleistocene interglacial high sea level are very poorly documented in Sonora (Ortlieb, in press). The pre-Sangamon marine deposits, observed in the areas of Puerto Libertad, Bahía Lobos and El Desemboque, do not crop out at elevations superior to + 10 m and this might be another indication of general stability of the coast in the Quaternary. In Playa Santa María (or Santa Margarita), 10 Km N of Puerto Libertad, are remnants of high stands of sea level, at + 22, at + 15 and + 10 m (unpublished data) these features are hypothetically attributed to surges (Hollin, 1972, 1980) of unknown age (pre- or post-Sangamon).

CONCLUSION: QUATERNARY EVOLUTION OF THE COAST OF CENTRAL SONORA

At the end of the Tertiary western Sonora had already acquired its characteristic structure of parallel NW oriented basins and ranges. Up to now, no marine transgression of Pliocene age has been identified in the coastal region of central Sonora. The sedimentation, in the last few million years, was essentially terrestrial: lacustrine, fluvial, alluvial and colluvial (bajada) deposits accumulated in the down-faulted valleys west of the Sierra Madre Occidental.

The Pleistocene (and even Holocene) climatic evolution in northwestern Mexico is still very poorly known. The results of the numerous investigations on Quaternary paleoclimates that have been carried out in southern Arizona and southern California cannot be extrapolated to Sonora. This last region is located in an intermediate position relatively to the truly tropical zone to the south, the mid-continent divide to the east, and the desert basins of the southwestern U. S. to the northwest. It may be assumed that the glacial periods of the Pleistocene were characterized by relatively high rainfalls, and possibly mild (or cool?) temperatures, while the interglacials presented a semi-arid to arid climate, similar to that of today. The Playa San Bartolo lacustrine deposits, as well as the thick bajada sediments which overlap the last interglacial (Sangamon) marine beds, at least prove that during (the major part of ?) the late Quaternary, the runoff has been greater than today.

In central Sonora, which is bordered by a relatively broad and shallow continental margin, there is no doubt that the Pleistocene coastline fluctuated between its present location and several kilometers (locally, some 20 or 30 Km) to the west. Thus, during Pleistocene glacial periods, when the sea level was about a hundred meters below the present datum, alluvial and deltaic sediments accumulated in what is now the nearshore area. During the interglacial periods, the maximum elevation reached by the sea was about the same than today MSL (or a few meters above, in the case of the Sangamon interglacial) and the coastline was probably similar in many aspects to what we can observe today. As a consequence of the relative tectonic stability of the coast and of the cyclic pattern of the glacioeustatic variations of sea level during the Middle and Late Pleistocene, only the last interglacial-glacial depositional sequence is conspicuous along the coast, the previous sequences being either eroded or concealed.

Almost all the characteristic features of the present coastline: lagoons and mangroves, barrier dunes, tombolos and spits, seacliffs and shore platforms (cut in late Pleistocene or older rocks) narrow cobble beaches and wide sandy beaches, etc. were formed in the last 5 or 6 000 years, during the last stage of the Holocene post-glacial rise of sea level.

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SEDIMENTOLOGY AND PALEOECOLOGY OF PUERTO PEÑASCO COAST,
NORTHWESTERN SONORA

(Roadlog not available at press time)

FIELD TRIP No. 3 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 21-23

TRIP LEADERS:

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REGIONAL GEOLOGY OF NORTHWESTERN SONORA

(Roadlog not available at press time)

FIELD TRIP No. 4 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 22-24

TRIP LEADERS:

J. F. LONGORIA,

(FORMERLY AT:) DEPARTAMENTO DE GEOLOGÍA,

M. A. GONZALEZ, FACULTAD DE INGENIERIA, UNIVERSIDAD DE SONORA,
HERMOSILLO, SONORA, MEXICO.

TECTONIC RELATIONS BETWEEN REGIONAL THRUST FAULTING
(CORDILLERAN OROGENY) AND THE RINCON MOUNTAINS
GNEISS-CORED DOME

FIELD TRIP No. 5 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 21-23

TRIP LEADERS:

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Roadlog For Trip No. 5

The Mesozoic-Tertiary Rincon gneiss-cored dome and the Mesozoic Cordilleran orogeny near Tucson, Arizona

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Introduction--The topics of the Cordilleran orogeny and of the gneiss-cored domes (commonly referred to as metamorphic core complexes) of southeastern Arizona are currently of particular interest. Each topic has generated its genetic models, some of which have already been set aside. It is our contention that the two types of deformation are genetically related and have, in part, developed together. The evidence is shown by field relations. Some key structural features will be examined during this 2-day excursion near Tucson. First, we will visit an area away from the Rincon dome where styles of Cordilleran deformation are illustrated without obvious influence of the dome. Then for comparison, we will visit the dome itself to observe both the Mesozoic Cordilleran style of compressive deformation and that of the Tertiary extensional gravity tectonics on the flanks of the dome. After comparing structural features across the dome, we believe that thrust faulting and initial doming have interacted. Tectonism and magmatism were long lived (or recurrent?) and overlapped in time and space.

As happens with most field excursions, this one has limitations of what can be shown, determined by the available time and the logistics of transportation and accessibility. It is easier, for instance, to reach thrust faults of the late stage of the Cordilleran orogeny, the Helvetian phase, than those of the earlier Piman phase. It is also easier to accept from a distance the basis for the generally accepted gravity sliding than the reason for thrust faulting in the Rincon Mountains, so we will concentrate on the latter.

We propose several walking tours of 1-5 km each. Thus, the participants must be suitably prepared and will carry their lunches and water one day. We hope that this extra effort will lead to the unfolding of an exciting tectonic story. In the course of the field trip, in addition to the tectonic story, we will see a variety of Precambrian, Paleozoic, and Mesozoic formations, plutons of many ages, and a variety of structural and fabric features. Among the sights will be Triassic volcanic rocks, tear faults merging with thrust faults, intrusive gypsum bodies, thrust faults that were reactivated as glide faults, and brecciated protomylonites.

Finally, the geology along the trip route has been studied recently, and so detailed geologic maps are available. Maps of the northern part of the Santa Rita Mountains have been prepared by several generations of workers (Schrader, 1915; Creasey and Quick, 1955; Drewes, 1971a and 1972a). Mesozoic rocks of that area were described by Drewes (1971b), as well as Cenozoic rocks (Drewes, 1972b) and plutonic rocks (Drewes, 1976). Aspects of the mineralization were described by Schrader (1915), Creasey and Quick (1955), and Drewes (1970 and 1972c); additional unpublished accounts are given by Lutton (1958), Heyman (1958), and Michel (1959). Geologic mapping in the Rincon Mountains was done by Drewes (1974 and 1977), Thorman and Drewes (1978 and 1981, in press), and Creasey and Theodore (1975). A study of the folded rocks around the dome was done by Davis (1975). The geologic maps will make the trip easier to follow, but they must be obtained independently for they are not included with the guidebook.

Day 1. Northern Santa Rita Mountains--Green Valley, Florida Canyon, Helvetia mining district, Sahuarita, and return.

Trip starting site is the junction of Interstate Highways 10 and 19, west from Park Avenue access on Interstate 10. In the roadlog directions from vehicle given by clock directions, with 12:00 oriented in direction of front of vehicle.

Mile

- 0.0 -- Junction Interstate Highways 19 and 10. Proceed south on 19 toward Nogales.
- 4.4 -- Pass beneath Valencia Rd. viaduct. Recheck odometer.
- 6.4 -- Del Bac Hills at 2:00. In middle distance, 1-3 km, is mid-Tertiary conglomerate capped by andesite dated as 24-27 m.y. (million years) old (Percious, 1968). Roadcut at 9:00 exposes tilted and faulted gravel. San Xavier Mission lies at small knoll at 9:00, 1 km from highway.
- 11.8 -- Overview: Tailings piles at 1:00-2:00 are near Pima-Mission porphyry copper pit. Knoll at 3:00 is underlain by andesite. White spot at 10:00 is a quarry and spoils pile in the Mississippian Escabrosa Limestone. High peak at 11:00 is Mount Wrightson, underlain by Triassic Mount Wrightson Formation, chiefly rhyolite.
- 16.9 -- Helmet Peak Road. Tailings piles of Twin Buttes porphyry copper pit at 1:00-5:00.
- 20.7 -- Beginning of Green Valley. Pass exit 65, Esperanza Blvd.
- 24.6 -- Leave Interstate 19 at exit for Continental and Madera Canyon and turn left.
- 25.2 -- Cross Santa Cruz River. Groves are pecan trees.
- 25.8 -- Turn right at Continental store.
- 25.9 -- Railroad crossing. Field trip route climbs terrace of mid-Pleistocene gravel.
- 32.7 -- Cross young fault, marked by a 2-m rise in gravel surface and capping red soil.
- 33.1 -- Turn left onto gravel road for Box Canyon
- 33.5 -- Bear right at road fork toward Florida Canyon USDA Range Experimental Station.
- 35.5 -- STOP 1: Traverse on foot (fig. 1) of several Mesozoic formations and the Sawmill Canyon fault zone, shown by Drewes (1971a).
 - a -- Knoll, underlain by rhyolite welded tuff and flows of the middle member of the Triassic Mount Wrightson Formation. Note two thin beds of eolian sandstone. Prospects contain copper minerals. [Drivers return to vehicles, drive to point f, or to point e if vehicles are suitable for rough track. Proceed on foot to b on traverse.]
 - b -- Cross narrow belt of Upper Cretaceous Fort Crittenden Formation, a brown conglomerate, sandstone, and siltstone, described by Drewes (1971b).
 - c -- Cross part of Lower Cretaceous Turney Ranch Formation, uppermost of the Bisbee Group, described by Finnell (1970) and Drewes (1971b).
 - d -- Sawmill Canyon fault zone and slices of Escabrosa and Permian Colina(?) Limestones. The fault has been repeatedly and diversely active; it is narrow here but flares out to the southeast.
 - e -- Pinal Schist of Precambrian age, here a phyllitic metasediment intruded nearby by the Continental Granodiorite, probably about 1600 m.y. old (Drewes, 1976). Fast walkers continue down track to vans at point f; others get to ride from the gate near the windmill.

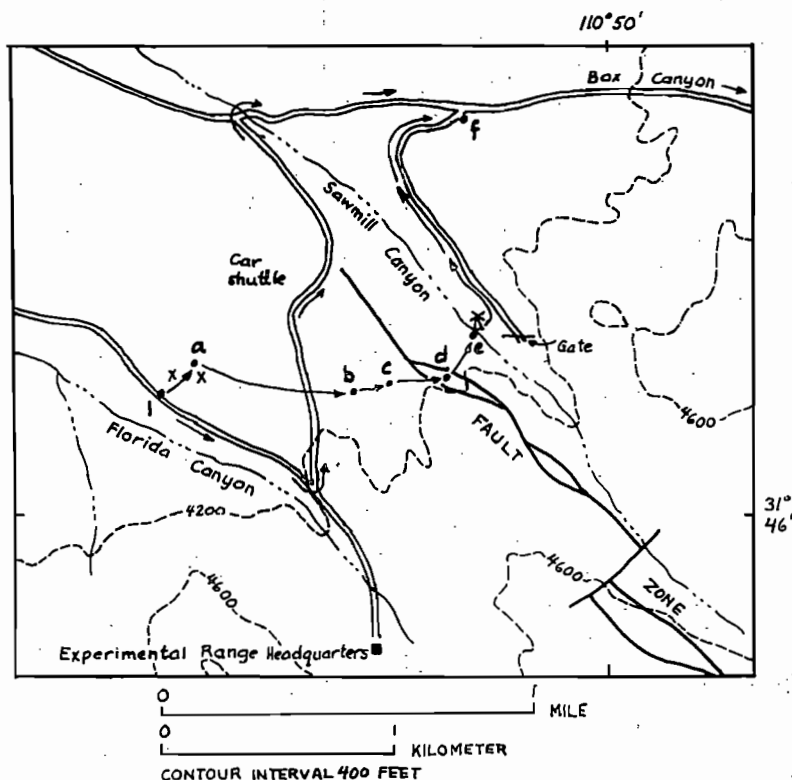


Figure 1. -- Sketch map of the Sawmill Canyon area, showing route of Stop 1 traverse on day 1. Base adapted from U.S. Geological Survey Sahuarita quadrangle, Arizona, 1:62,500, 1971.

Road log for drivers: Mile 36.0 - Turn sharp left onto Road 481. Mile 37.2 - Turn right onto Box Canyon Road. Mile 37.8 - Cross cattle guard and park vans at entrance to rough ranch road. Mile 38.4 - Gate above windmill. Mile 39.0 - Return to vans, point f. Continue northeast, toward Box Canyon.

Mile

- 40.0 -- Entrance to Santa Rita Ranch on right. Recalibrate odometer if necessary, but do not enter. Note Oligocene rhyolite dikes in hills beyond the ranch. One of them is dated at 26 m.y.
- 41.6 -- Turn left on road 485, to Helvetia. During the next several miles pass small outcrops of Continental Granodiorite, here recrystallized, with an age reset to 800 m.y.
- 44.8 -- Cross small saddle. Lowland at 12:00-2:00 is underlain by Paleocene granodiorite. In gully nearby it is dated at 54 m.y.; possibly its age is 1 or 2 m.y. older.
- 46.1 -- Turn sharp right, to Helvetia. Hill at 9:00-11:00 is underlain by Continental Granodiorite, cut by a mineralized quartz vein.
- 47.4 -- Hill at 3:00 marks west end of Helvetia Klippe. Bear right in next wash, keeping to main road.
- 48.7 -- East end of klippe, to be visited later.
- 49.5 -- STOP 2: Park in flats below shaft. Lunch at cars. Thereafter, begin traverse, using Figure 2 (and Plate 4 of Drewes, 1972a).

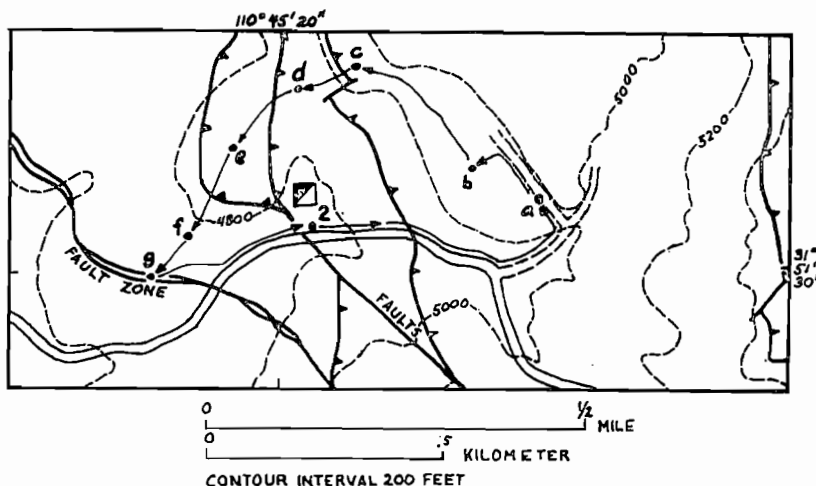


Figure 2. -- Sketch map of part of the Helvetia mining district, showing route of Stop 2 traverse on day 1. Base adapted from U.S. Geological Survey Sahuarita quadrangle, Arizona, 1:62,500, 1971.

- a -- Drill site at lower member of the Permian Epitaph Formation, mainly a marlstone or dolomitic siltstone, and some beds of argillite and limestone. Note pods of gypsum. Slope to east underlain by Permian Scherrer Formation and overlying Concha Limestone.
- b -- Hill, capped by upended Pennsylvanian Horquilla Limestone, provides overview of Paleozoic formation, thrust faults, and tear faults.
- c -- Thrust faulted slivers of Devonian Martin Formation and the Cambrian Abrigo Formation and Bolsa Quartzite.
- d -- Precambrian Continental Granodiorite thrust slice.
- e -- Second thrust plate of lower Paleozoic formations.
- f -- Another slice of Continental.
- g -- Recrystallized Bolsa Quartzite along Gunsight Notch fault zone. Return to cars and drive west, down the road.

- 50.3 -- STOP 3: East side of Helvetia klippe. Review of the local structure of the klippe cutting one stock and cut by another, of the klippe containing a segment of one tear fault and cutting across the Gunsight Notch tear fault, and of the localization of mineralization along the thrust fault. If time permits, visit 54-56 m.y. old "ore-porphyry" stock in gully north of road or part of the Willow Canyon Formation of the Bisbee Group and Concha Limestone on the hill south of the road.
- 50.6 -- STOP 4: Park at curve in road and walk down the road, southwest, to the thrust fault beneath the Helvetia klippe at mile 50.7, where the lower quartzite member of the Scherrer Formation lies upon the 54-56 m.y. old granodiorite (to quartz monzonite) stock. Ghost town of Helvetia lies nearby to the west.
- 51.7 -- STOP 5 (optional): View of west side of the klippe. Brown ledge to the northeast is the Bolsa Quartzite, the overlying soft slope is of Abrigo and Martin Formations, and the hill is capped by the Horquilla Limestone. In prospects all around the klippe the thrust fault is seen to dip 30°-45° inward.
- 52.8 -- Bear right at road junction.
- 52.9 -- Bear right again, on road for Sahuarita. -

- 54.3 -- Huerfano Butte at 3:00, underlain by Continental Granodiorite, and by an aplite body across the crest of the butte.
- 65.7 -- Turn left, on Sahuarita Road, driving through more pecan groves and recrossing the railroad and the Santa Cruz River.
- 66.4 -- Cross U.S. 89 at Sahuarita store; continue west.
- 67.3 -- Access to Interstate 19, turn right toward Tucson.
- 84.2 -- Junction of Interstate 19 and 10. End of trip. To return to motel turn east on 10 and exit on Old Benson Highway.

Day 2 - Rincon Mountains--Redington Pass and return; Saguaro National Monument Headquarters area and return.

Trip starting site is reached travelling north on Park Ave. to 22nd St., east on 22nd to Wilmot Blvd., north on Wilmot, which becomes Tanque Verde Rd. where it turns east, and to the U Tote M convenience store at the junction of Bear Canyon Road and Tanque Verde Road.

- 0.0 -- U Tote M store. Proceed east on Tanque Verde Road.
- 0.2 -- Junction Mt. Lemmon Road and Redington Rd.; bear right on Redington Road.
- 4.0 -- STOP 1: Park opposite billboard for 76^{er} Estates. Roadcut of mid-Tertiary Pantano Formation, a tilted and faulted pinkish-gray pebble conglomerate, described by Pashley (1966) and Finnell (1970). Clasts are subangular and comprise mid-Tertiary volcanics, Precambrian Rincon Granodiorite, and Paleozoic and Mesozoic sedimentary rocks, all of the allochthonous cover rocks. The beds are unconformably capped by a terrace deposit containing coarser detritus derived from the crystalline core of the Rincon dome. Review of geographic features of the Rincon Mountains and the Santa Rita, Sierrita, and Tucson Mountains.
Review of 3 main terranes of the Rincon Mountains, the crystalline core rocks, the cover rocks, and onlapping gravel deposits. Also note turtle-back-like anticline plunging gently southwest along Tanque Verde Ridge.
- 8.2 -- Junction of Wentworth road; proceed straight ahead.
- 10.1 -- First spur of the mountains at 9:00, of protomylonitic gneissic quartz monzonite. In gullies at 3:00, out of sight, is the Santa Catalina complex fault, dipping 20° south.
- 11.4 -- End of pavement; start of mountain road.
- 12.4 -- Roadcuts and hillsides nearby show lit-par-lit injection gneiss of Tertiary-Cretaceous light-gray quartz monzonite and Precambrian dark-gray granodiorite porphyry.
- 14.9 -- Slopes nearby at 10:00-11:00 show that the quartz monzonite is the dominate rock type. This is the Wrong Mountain Quartz Monzonite (originally mapped as Precambrian(?) by Drewes (1974) but now accepted as Cretaceous or Paleocene. The quartz monzonite of Sanmaniego Ridge of Creasey and Theodore (1975) is another subtly different phase of this rock.
- 19.0 -- Entrance to Bellota Ranch on left; keep to main road.
- 21.4 -- STOP 2: Park to right and walk up knoll to the southeast. Review geography and geology of the Italian Trap klippe. (Use map of Thorman and Drewes, 1978 if the bulletin on that area is still in press.)
- 23.6 -- Turn right onto unmarked road to Youtcy Pasture Tank.
- 24.0 -- STOP 3: Traverse and car shuttle. Pack water and lunch. En route to 3a view Continental Granodiorite and note foliation and weak lineation, here aligned north-south and not N. 70 E., as is typical. Tertiary rhyolite dikes occur west of jeep track. See figure 3 for route.

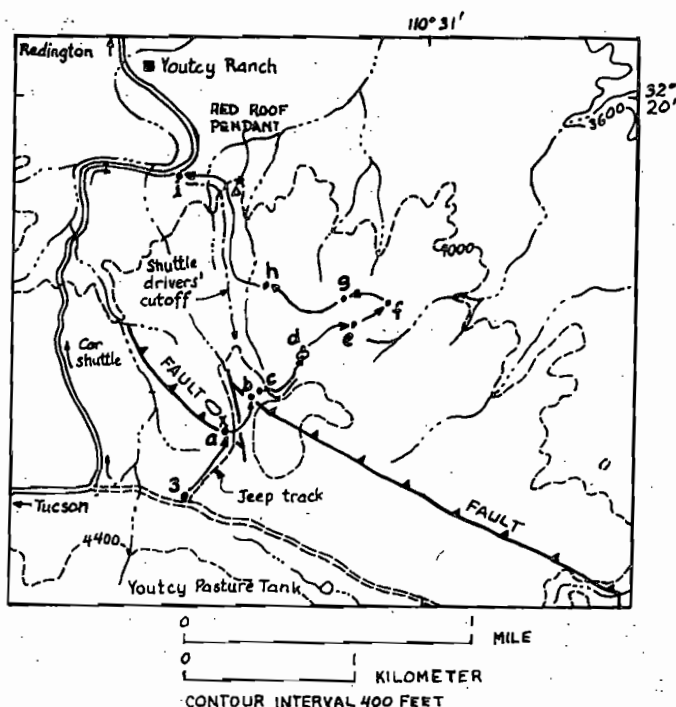


Figure 3. -- Sketch map of the Youtcy Ranch area, showing route of Stop 3 traverse on day 2. Base adapted from U.S. Geological Survey Bellota Ranch quadrangle, Arizona, 1:62,500, 1957.

Road log and instructions for drivers: Return to Redington Rd., 24.4, turn right on Redington Road, toward the northeast; 26.1, at sharp corner park at right side of road. Walk up the spur to the southeast by the red roof pendant, and up that spur to the jeep track. Follow the track uphill to the southeast and rejoin the group. Don't forget your water and lunch.

- a -- Saddle area, and spur to west. Horquilla Limestone is faulted upon the granodiorite. Fault in and near copper prospect dips about 60° northeast. Return to saddle and walk on the contour northeast, using cattle track.
- b -- Spur. Horquilla Limestone again faulted upon the granodiorite, offset from site a by a north-trending high-angle fault. On ridge to the west the brown rock overlying the Horquilla is the Bisbee Formation. Continue about 200 m to the jeep track in the small saddle.
- c -- Small draw. Thrust platelet of Continental Granodiorite in the larger plate, mainly of Paleozoic and Mesozoic sedimentary rocks. Note that the foliation and lineation in the platelet is oriented like that of the underlying crystalline core rocks. Continue on jeep track to second saddle; leave the track and climb the knoll to the north.
- d -- Knoll, underlain by metamorphosed quartzitic sandstone and siltstone of the Bisbee Formation. Review of geology of the Mineta Ridge area, to the east. Also indicate site of next stop.
- e -- Gully underlain by diabase (or metadiorite) associated with the Precambrian Apache Group; plate of Continental lies to the southwest, above the Bisbee;

and to the east and northeast are red argillite of the Pioneer Formation and quartzitic sandstone and round-clast conglomerate of the Barnes member of the Dripping Springs Formation, all of the Apache Group. Continue to knoll to the east.

- f -- Another quartzitic sheet has been mapped as more Dripping Springs (which, however, may be Bolsa Quartzite). The spur to the northeast has Martin Formation, Escabrosa Limestone, and Horquilla Limestone, which may be visited if time permits. Note that all plates seen thus far are part of the lower main thrust plate. To the east, beyond our traverse, is the second major thrust plate of Precambrian rock, here made up of the Oracle granodiorite (or granite?) (of Peterson, 1938). The upper major plate is present to the southeast. On Mineta Ridge is the glide plate of Tertiary Pantano Formation (locally and informally known as the Mineta beds) and the associated volcanic rocks; a very coarsely porphyritic andesite known as the Turkey-track porphyry of Cooper (1961), a non-porphyritic andesite, and a rhyolite welded tuff. Both the fossils in the sedimentary rocks and a radiometric on the porphyry, date by Damon's laboratory, place the rocks as probable earliest Miocene, about 26 m.y. old. Continue west to the bouldery outcrops.
 - g -- Bold outcrops of the Tertiary Youtcy Ranch stock of quartz monzonite. Lunch stop. Thereafter, continue west, about on a contour, along the margin of the stock.
 - h -- On spur. The Youtcy Ranch rock is subtly foliated here but is not lineated. A single sample of this quartz monzonite gives ages of 35-16 m.y., depending on mineral and method used. Perhaps these ages reflect a gradual (19 m.y.) cooling through various annealing temperatures, having a span of about 150°C. Foliation in this pluton must be younger than that in the core rocks and in the thrust plates. Continue down the spur and across the gully, past the edge of a reddish outcrop of a pendant or inclusion of the Bisbee Formation, to the cars.
 - i -- End of traverse. Resume driving, back to the west across Redington Pass and Tanque Verde Road.
- 43.6 -- Turn left at Wentworth Road, which is partly unpaved and crosses Tanque Verde Wash.
 - 44.7 -- Turn right on Speedway.
 - 45.6 -- Turn left on Freeman Road.
 - 46.6 -- Cross Broadway and continue on Freeman, partly unpaved.
 - 47.6 -- Twin Buttes about 1 km from the road at 3:00. They are underlain by Tertiary rhyolite and the Turkey-track porphyry of Cooper (1961).
 - 48.9 -- STOP 4 (optional): Second of 3 road cuts exposes a quartzitic phyllite of the Pinal Schist and, cutting the schist a fine-grained border phase of the Rincon Granodiorite(?). From nearby the rock is dated as 53.3-1390 m.y.; probably these are both partly reset from an original age of about 1560 m.y. This granodiorite is known only from the cover, not the core of the dome.
 - 49.6 -- Turn left on Old Spanish Trail.
 - 49.7 -- Turn left, entrance to Saguaro National Monument.
 - 49.8 -- Monument Headquarters buildings. Continue on Loop Drive. For the first few kilometers the hills are underlain by the quartzitic Pinal Schist.
 - 53.3 -- STOP 5 (optional): Turkey-track porphyry, exposed in rubble crop, to illustrate the large phenocrysts.

- 53.6 -- Outcrops along road are of crystalline core rocks. Santa Catalina complex fault is covered along the drive.
- 55.0 -- STOP 6: Traverse on foot; head northwest.

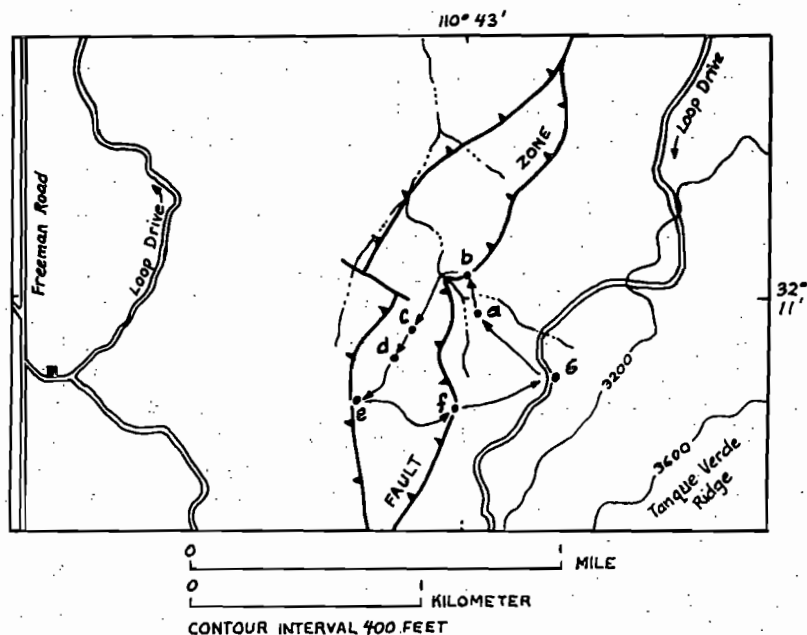


Figure 4. -- Sketch map of Saguaro National Monument Headquarters area, showing route of Stop 6 traverse on day 2. Base adapted from U.S. Geological Survey Rincon Valley quadrangle 1:62,500, 1957, revised 1968.

- a -- Overlook into gully. En route to a note the lit-par-lit gneiss, the foliation and lineation, and the protomylonitic fabric. On the opposite canyon wall the ledge, dipping 15° west, is the indurated brecciated protomylonite, separating the underlying crystalline core rocks from an overlying melange sheet.

Continue across the canyon to the stripped surface of the Santa Catalina complex fault.

- b -- Fault surface. The ledge is an indurated breccia of the protomylonitized core rock. The overlying rocks form a tectonic melange sheet, in which there are scattered blocks as much as 100 m across. The matrix is a red siltstone or marlstone of the Permian Earp Formation. The blocks are assorted Paleozoic and Mesozoic formations. Some of the limestone blocks are internally folded, and these have been systematically studied by George Davis (1975) and his students, who infer the melange to have moved northwest off the adjacent part of the Tanque Verde Ridge during or after Pantano time. It is a movement different from, and later than, that which formed the lineation.

Return across the canyon and climb the ridge to the south.

- c -- First knoll on the ridge. Nearby blocks in the melange are of red jasper pebble conglomerate (Earp), and of dark-gray dolomite (Permian Epitaph Dolomite). Outcrops on the hill below, to the west, are of quartzitic Pinal Schist, lying upon the melange sheet.
- Continue south along the ridge.

- d -- Second knoll on the ridge. Nearby blocks are of brownish-gray to purplish-gray quartzite (Bolsa), and of sandstone (Bisbee). In the next saddle to the south the matrix is exposed.
Cross the third knoll and descend to the southwest.
 - e -- At small gully west of the knoll. Stripped fault surface between the underlying melange sheet and overlying Pinal Schist.
Return to top of third knoll and head east along ridge. Bold outcrop on the northeast side of the knoll is of Glance Conglomerate, base of the Bisbee Formation. Descend to the east, toward the vehicles.
 - f -- Base of melange sheet, here poorly exposed at midslope.
Return to vehicle and continue around Loop Drive.
- 57.9 -- Monument Headquarters. End of trip. Return to motel via Old Spanish Trail, 22nd Street, and Park Avenue.
- End of Trip.

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OPOSURA MASSIVE SULFIDE, NEAR MOCTEZUMA, SONORA

FIELD TRIP No. 6 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 24

TRIP LEADERS:

C.K. MARRS, DEPARTMENT OF GEOSCIENCES, UNIVERSITY OF ARIZONA, TUCSON,
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D. GOMEZ H. SERVICIOS INDUSTRIALES PENOLES S.C. ROSALES No. 56 HERMO
SILLO, SONORA, MEXICO.

GEOLOGY AND DEPOSITIONAL ENVIRONMENT OF THE OPOSURA MASSIVE SULFIDE DEPOSIT,
SONORA, MEXICO

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ABSTRACT

The Oposura massive sulfide deposit is located in the Sierra La Huerta, approximately 150 km east of Hermosillo, Sonora. Stratabound and strataform lenses of massive and thinly bedded sulfides vary in thickness from several centimeters to 3.5 meters. These lenses and beds are interbedded in a sedimentary, volcani-clastic, and pyroclastic section of limestone, water-lain fine crystal tuff, and lithic tuff known as the Arenillas Formation.

Mapping has shown that localization of sulfides in this volcano-sedimentary sequence occurs spatially between two thicker, more extensive volcanic units, the footwall Revancha Rhyolite ignimbrite and a hanging wall sequence of dacitic tuff breccia, lithic tuff, and feldspathic tuff. Locally the rhyolite is intensely silicified and displays eutaxitic texture. The overlying dacitic sequence is characterized by chlorite-epidote-pyrite alteration. The age of the Oposura Volcanic Group is estimated at upper Cretaceous to Paleocene.

Field and laboratory examinations identified four types of mineralization. The most common type consists of beds of fine-grained sphalerite and galena hosted by fine crystal tuff. It is associated with chlorite, epidote, and important amounts of manganese, either as rhodonite or rhodochrosite. The mineralization is interpreted to have formed syngenetically with deposition of crystal tuffs and limestone in a shallow, sub-aqueous, volcano-sedimentary environment.

(This paper has been excerpted by J. M. Guilbert from C. K. Marrs' M. S. Thesis, Department of Geosciences, Univ. of Arizona, Tucson (1979).

INTRODUCTION

The Oposura deposit in east central Sonora, Mexico, consists of strataform, stratabound lenses of thick-bedded massive sulfides and thin-bedded sulfides interbedded in a complex sequence of rhyolitic to dacitic volcanic, volcanoclastic, and sedimentary rocks. This sequence of sedimentary rocks, about 30 meters thick, occurs spatially between two thicker and more extensive volcanic units, a footwall rhyolite ignimbrite or dome and a hanging wall rhyodacitic tuff breccia.

The Oposura deposit was chosen for study because an undeformed massive sulfide deposit could best demonstrate the relationship between volcanism, sedimentation, and mineralization. Oposura is a unique deposit because its undisturbed setting allows examination unhindered by metamorphic overprinting of primary volcanic and sedimentary features and their relation to sulfides.

Location

The Oposura deposit is located 15 kilometers west of the town of Moctezuma in the east central part of the State of Sonora, Mexico, 250 kilometers south of Douglas, Arizona (Fig. 1). The mine workings lie between 1200 and 1400 meters elevation on the east slope of the Sierra La Huerta, a north-northwest trending mountain range typical of the Basin and Range physiographic province. Access to the mine is by paved road northeast from Hermosillo on the Ures-Nacozari highway and then by rough ranch roads to Rancho Arenillas, turning south off the pavement at the crest of the Sierra La Huerta.

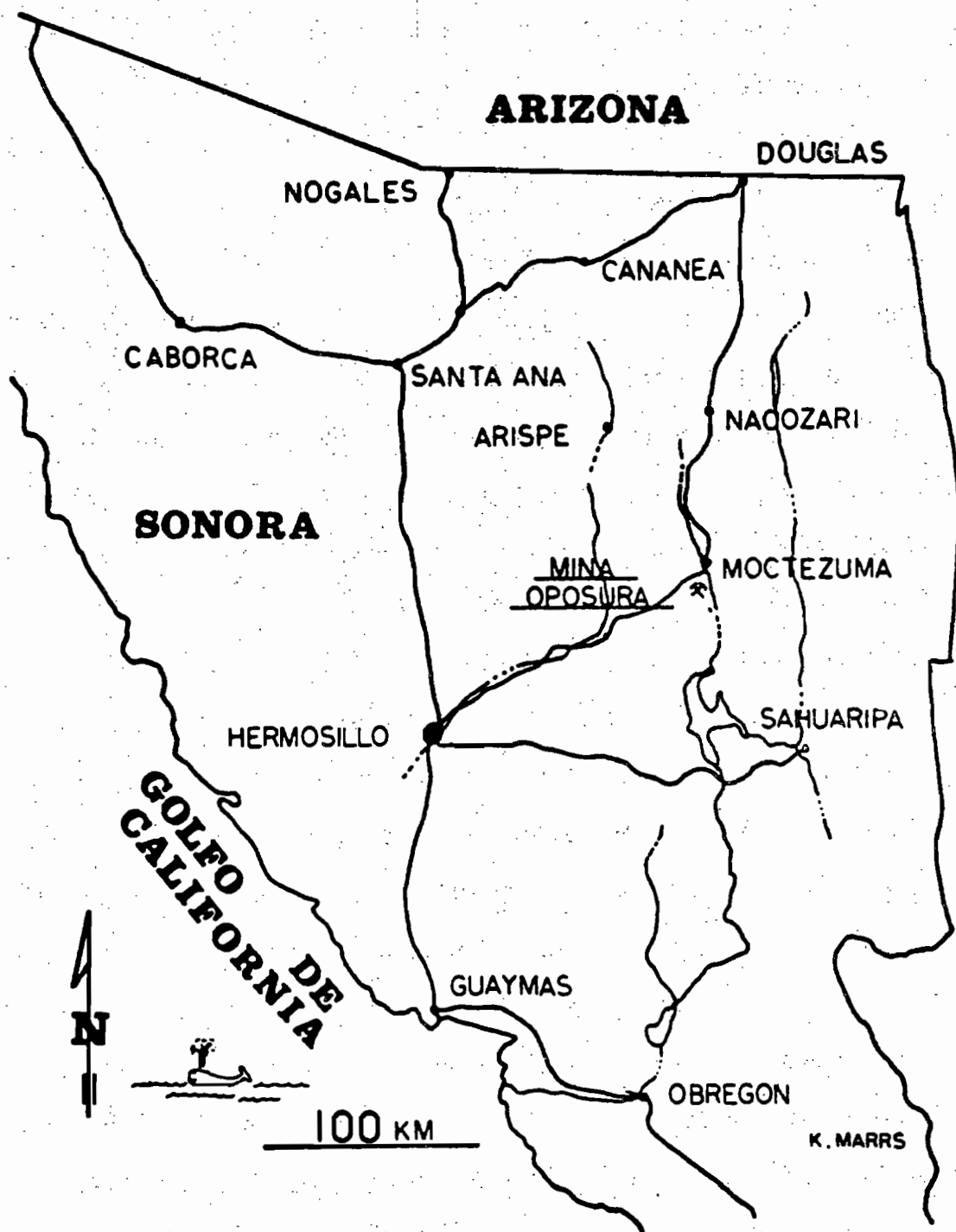


Figure 1. Location Map of the Oposura Mine, Sonora, Northern Mexico

Previous Work and History

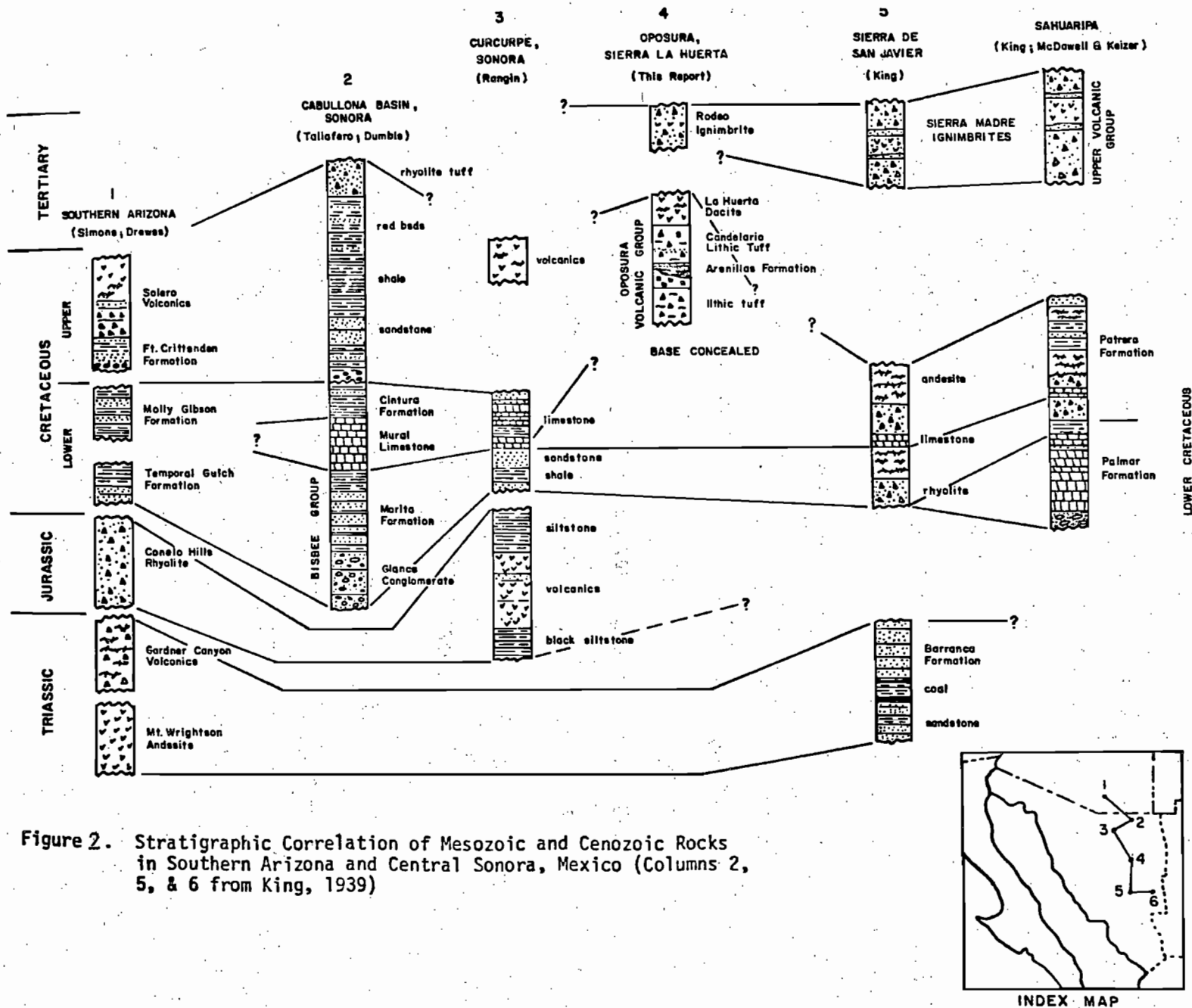
The Oposura mine, sometimes referred to as "Las Arenillas", has been mined sporadically on a small scale since the 1920's for copper, silver, lead, and zinc. This initial exploration was under the direction of Dr. L. D. Ricketts. In the 1930's, E. Wisser and R. N. Hunt of the Real del Monte y Compania de Pachuca made a brief geologic examination and property appraisal (Ayala, 1972). They described Oposura as a replacement deposit within limestone overlain by tuffaceous beds and underlain by quartz porphyry intrusive. The Anaconda Company acquired control of the property in the late 1940's and conducted an extensive underground exploration and development program headed by V. Perry and R. Mulchay. During this effort, approximately 1000 tons of ore produced at Oposura averaged 1.1% Cu, 150 gm/ton Ag, 9.9% Pb, and 29.5% Zn (Thoms, 1973) and were shipped to the Shattuck Denn smelter at Bisbee, Arizona. Exploration work continued intermittently until 1970 consisting of numerous shallow drill holes, surface trenches, twenty short adits, and one large development adit, "Tunnel D". In 1975, a joint venture was formed between Anaconda and Servicios Industriales Peñoles, S.A., to evaluate the potential of the Oposura deposit. A program of geologic mapping and diamond drilling was initiated in 1976. This study, supported by Servicios Peñoles and The Anaconda Company was conducted concurrently with, but separately from, the Peñoles program.

REGIONAL GEOLOGY

The regional geology of the state of Sonora is varied and complex, similar to that of Arizona to the north, but the detailed geologic unraveling of this large expanse is much less complete than is Arizona's. A regional compilation of Sonoran geology was recently updated by Lopez-Ramos (1974), but only scattered reports of more detailed field examinations are available. Earlier detailed geologic information of the immediate Oposura area and the surrounding portion of Sonora is totally lacking. No geologic maps of the Sierra La Huerta area exist other than the 1:500,000 geologic map of Sonora (Lopez-Ramos, 1974). The nearest area covered by published reconnaissance geologic mapping done by King (1934 and 1939) is 75 km south near Sahuaripa, Sonora (Fig. 1), which includes an excellent summary of regional geology and stratigraphy.

OPOSURA DISTRICT GEOLOGY

The Oposura District and the Sierra La Huerta are almost entirely underlain by volcanic rocks including flows, ignimbrites, and both lithic and crystal tuffs. Units ranging in age from probable late Cretaceous up to Pleistocene are depicted in the district stratigraphic column (Fig. 2). Prevolcanic Jurassic and Triassic sedimentary formations are not recognized in the map area but are included to complete the Mesozoic-Cenozoic stratigraphic section for east central Sonora.



Stratigraphy

In keeping with the work of Fredrikson (1974) and Henry (1975) in Sinaloa and Clark (1976) in Chihuahua, the volcanic rocks are divided into two major families, the Lower Volcanic Group (estimated at 100 to 45 m.y.) and the Upper Volcanic Group (32 to 22 m.y.). The ages are approximate and the limits vary according to locale. Rock units in the Oposura study area were designated on the basis of relative stratigraphic position, lithology, and texture. No pre-existing nomenclature was available for the Oposura District, nor or are there any published age dates.

The Oposura Volcanic Group is the proposed equivalent of the Lower Volcanic Group and includes the lower lithic tuff, Revancha Rhyolite, Arenillas Formation, Candelaria Lithic Tuff, and the La Huerta Dacite (Fig. 3). The Oposura Group is dominated by lithic lapilli tuffs, crystal tuffs, and flow-banded porphyritic dacite. The volcanic rocks are primarily intermediate in composition with locally important felsic rocks in the mine area.

The stratigraphically lowest member of the Oposura Group is the lower lithic tuff unit, a dark-colored dacitic lithic lapilli tuff containing small white feldspar crystals. The exposures of this unit are limited to the mine area. The base is not exposed, but the lithic tuff lies in conformable contact with the overlying Revancha Rhyolite.

The Revancha Rhyolite is a white to creamy tan, dense, siliceous rhyolite which locally contains white, flammé-like fragments. Spatially, it is one of the most important lithologic units in the Oposura Group because it always immediately underlies the massive sulfide

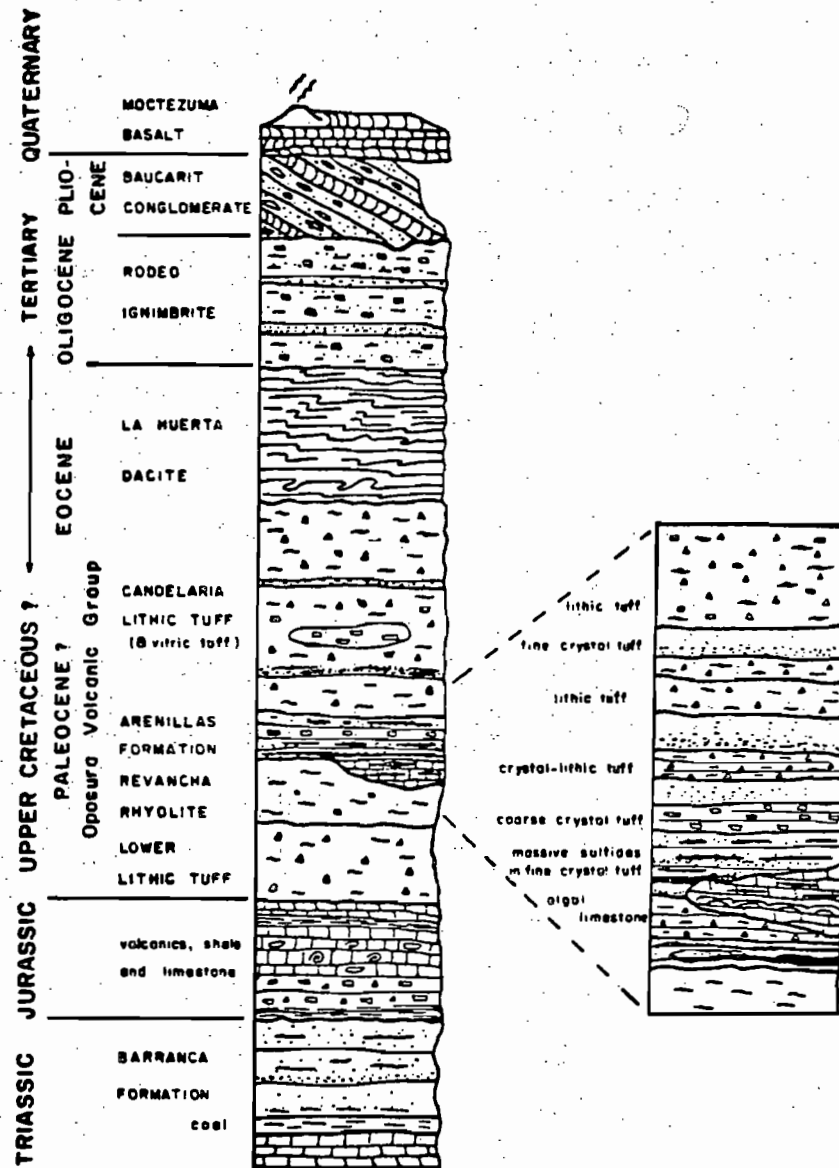


Figure 3 Stratigraphic Column of the Oposura District

The column on the right is an expansion of the Arenillas Formation.

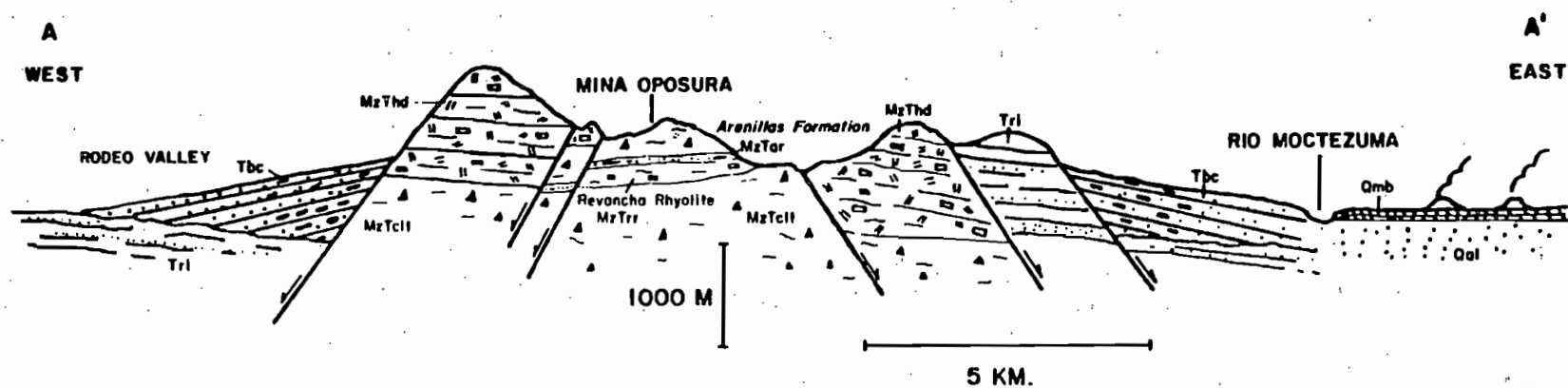


Figure 4 Generalized Geologic Section of the Northern Sierra La Huerta

mineralization. The upper contact of the Revancha Rhyolite is undulating and irregular and locally highly altered and discolored. This contact with the overlying Arenillas Formation is conformable for the most part but locally appears erosional.

The Arenillas Formation consists of thin-bedded crystal tuff and crystal lithic tuff displaying sedimentary textures and interbedded with thinly laminated limestone. All known massive sulfide mineralization occurs in the fine-grained bedded tuffs and limestone of the Arenillas Formation. The detailed stratigraphy of the Arenillas pyroclastics and volcanoclastics is extremely variable and complex when compared to the more massive and homogeneous rhyolite and lithic tuff units. This unique and distinctive unit always immediately overlies the Revancha rhyolite. No fossils occur in the limestone, although algal structures have been observed.

The Candelaria Lithic Tuff is the most extensive unit in the Oposura Group and occurs in conformable contact with the older Arenillas Formation. Thin green Arenillas type crystal tuff beds are occasionally interbedded in the overlying lithic tuff. The most common lithology in the Candelaria unit is dark-gray-colored lithic lapilli tuff, although a black feldspathic vitric tuff is also common.

The La Huerta Formation is the youngest unit of the Oposura Group and conformably overlies the Candelaria rocks. Porphyritic flow banded dacite is the most common rock type, but locally flow banded rhyolite and andesite are conspicuous. The light color and flow-banded texture contrasts greatly with the underlying Candelaria Lithic Tuff.

No diagnostic fossils were observed in the Oposura Group, nor are there any definitive contacts with known sedimentary rocks. A tenuous correlation can be made between the Arenillas Formation and the limestone sequence interbedded in Cretaceous volcanic rocks described by King (1939). One whole-rock potassium-argon age determination on a specimen from the Candelaria Lithic Tuff gave an approximate age of 40 ± 2 m.y. (P. Damon, oral commun.). However, this sample was of partially (50%) devitrified crystal-vitric tuff so its age should be regarded as minimum. The best age estimate for the Oposura group would be early Tertiary, pre-dating the 50-55 m.y. old Laramide intrusives of northern Sonora.

A thick and extensive rhyolitic ash flow unit, the Rodeo Ignimbrite (Fig. 3), is the proposed equivalent to the Upper Volcanic Group described in the Sierra Madre Occidental by McDowell and Keizer (1977). The Rodeo Ignimbrite includes all of the light pink, porous, and poorly indurated ash flows that lie in gentle angular unconformity upon the dark, siliceous, highly indurated rocks of the Oposura Group. Air fall tuff beds that contain abundant unflattened pumice fragments are commonly interbedded in the Rodeo Formation. Rodeo volcanic rocks fill the valley west of the Sierra La Huerta and overlie the La Huerta Dacite in the east. The ash flows have not been dated but are correlated with those of Oligocene age.

Coarse conglomerates occur along both the east and west margins of the Sierra La Huerta and are mapped as the Baucarit Formation, originally described by King (1939) in central Sonora. The Baucarit is typical of other Basin and Range type conglomerates, consisting of large

boulders and cobbles derived from the adjacent uplifted mountain range, filling in the newly formed valleys. The Baucarit is probably Pliocene in age (King, 1939).

Quaternary to Recent Moctezuma basalt flows and cinder cones occur in the Rio Moctezuma Valley and are parallel to the east flank of the Sierra La Huerta. The basalt flows cover a minimum of 50 square kilometers from several different outpourings. The presence of several uneroded cinder cones suggests an age similar to the 1-5 m.y. San Francisco volcanic field in northern Arizona.

This report will follow the nomenclature for volcanoclastic and pyroclastic rocks proposed by Fisher (1961, 1966). The general description of volcanic rocks will utilize the work of Williams, Turner, and Gilbert (1954). Confusion abounds over the use and meaning of such terms as ignimbrite, ash flow, pyroclastic flow, and volcanoclastic rocks in general. The reader is urged to consult these references and the brief summary in the appendix in order to sort out complex volcanic rock terminology.

Revancha Rhyolite

The Revancha Rhyolite is an extremely dense, siliceous rock and is the most felsic member of the Oposura Group. It is easily recognized by the creamy tan to white color, which is in strong contrast to the much darker lithic tuffs. When fresh and unaltered, the aphanitic groundmass is pinkish tan, although weathered samples are commonly red and altered samples may be bleached to a sugary white. Flattened and elongated light tan fiamme (2-10 mm) are present in varying amounts (0-30%), and form a faint, relic eutaxitic texture in some varieties of rhyolite. The rhyolite contains up to 10% minute plagioclase phenocrysts (1 mm), but they are often inconspicuous and difficult to recognize. Disseminated pyrite, ranging from 1 to 5%, is ubiquitous, and disseminated galena may also be present. Three separate zones of anomalous amounts of disseminated galena and sphalerite in the rhyolite were observed in drill core (BD-1) with values up to 2800 ppm lead and 1000 ppm zinc.

The Revancha Rhyolite crops out along a lateral distance of 3 kilometers and forms a crescent-shaped topographic pattern stretching from the West End workings through the main camp area to the North End area. The general features of the rhyolite will be discussed here, but some details will be left to the section on mine geology. The total thickness may be as much as 150 meters but measurements are complicated by faulting. It is apparently thickest beneath the main camp or Tunnel D area. The Revancha Rhyolite is cut off on the west by the Arroyo de Campo Fault, covered by the Candelaria Lithic Tuff to the

northwest, eroded on the south, and either pinches out or is obliterated by the quartz-pyrite vein swarm on the northeast. It appears that the rhyolite is of limited lateral extent, perhaps due more to faulting than to actual thinning.

The contact with the underlying lower lithic tuff is transitional and the lower limit of the rhyolite is placed where lithic fragments dominate over feldspars or fiamme. No consistent size grading of fiamme or feldspar phenocrysts was defined although the rhyolite does show numerous vertical and lateral variations in color, abundance of phenocrysts, and prominence of fiamme.

The presence of elongate and flattened fiamme indicates that the Revancha Rhyolite was emplaced as hot ash flow or welded ignimbrite. Its minor areal extent indicates a local type of eruption characterized by the mingling of more explosive fiamme-rich tuff with medium porphyritic rhyolite.

The most common varieties of rhyolite are 'massive medium-porphyritic' and 'ignimbritic'. The porphyritic type, most common near the upper contact with the Arenillas Formation, has a milky aphanitic groundmass that contains relatively abundant feldspar crystals. Its lack of internal structure may be due to alteration near the contact. The ignimbritic rhyolite displays a faint to well-developed eutaxitic texture with feldspar crystals subordinate to the fiamme particles. Disseminated pyrite cubes are common and comprise up to 5% of the matrix. Oxidized pyrite imparts a distinctive red color to weathered samples.

Petrographic Description. All varieties of Revancha Rhyolite are dense, leucocratic, siliceous rocks. Euhedral to subhedral lath-shaped plagioclase crystals occur in a cryptocrystalline quartz-rich groundmass. Most of the plagioclase is albite or oligoclase, twinning is common, and crystals rarely exceed 1 millimeter. Carbonate alteration occurs in the plagioclase grains, and blebs of carbonate can be found locally in rhyolite.

The groundmass consists of microgranular or cryptocrystalline quartz with occasional blebs of pistachio-green epidote and leucoxene. Cut and polished surfaces have a well-defined eutaxitic texture composed of wispy, elongate, cream-colored fragments. No porous pumice has been identified but the habit is very similar to that of flattened pumice. Plane polarized light reveals an excellent fluidal structure not evident under crossed nicols. This texture probably indicates a wholly devitrified glassy matrix. In some instances, epidote occurs in the centers of altered plagioclase grains.

Arenillas Formation

The Arenillas Formation is a complex interbedded sequence of limestone, water-lain crystal tuff beds, and crystal-lithic pyroclastic beds. The limestone forms the lower part of the formation, but is not everywhere present. The predominantly green color and granular texture of tuff beds associated with light gray, thin-bedded limestone make the Arenillas a distinctive, easily recognized formation. All the mineralization at Oposura is contained in the Arenillas, making it economically the most important formation.

The granular tuffs and interbedded limestone were first recognized by Hunt and Wisser (1934) at the old Arenillas workings, now included in the Oposura mine property, and the name is taken from this old prospect. Previous workers all noted the close association of bedded tuffs with the sulfides, but they were mainly concerned with the limestone and thought it to host the majority of mineralization. The type area for the Arenillas Formation is the Tunnel "D" area. The best exposures and examples of the tuff beds are in the underground workings and drill core.

Distribution. The Arenillas Formation crops out in a narrow, nearly 3 kilometer belt along the upper contact of the Revancha Rhyolite.

Its near-horizontal attitude is reflected in outcrop pattern which closely follows contour lines, generally dipping gently to the northwest, but rarely to the southeast. Thickness of the Arenillas Formation varies from 20 to 60 meters, depending primarily on the thickness of the limestone which ranges from 20 meters thick to entirely absent.

The base of the Arenillas is defined as the top of massive Revancha Rhyolite but in some instances a zone of apparent mixing between the two units forms a "dirty", discolored hybrid rock or transition zone. This hybrid zone shows more affinities to the underlying rhyolite than to the bedded tuffs.

The Arenillas is conformably overlain by the Candelaria Lithic Tuff. The contact is distinct because of the sharp contrast between the fine granular tuffs and the overlying coarse lithic lapilli tuff.

However, thin fine-grained tuff beds do occur in the Candelaria, demonstrating sporadic recurrences of the earlier Arenillas-type volcanic environment.

Stratigraphy. Although the Arenillas Formation consists of numerous alternating beds of similar texture and mineralogy, a gross pattern may be discerned. Where it occurs, laminated limestone is at the base of the sequence, usually directly on the Revancha Rhyolite or separated from it by a thin crystal tuff bed. In areas where the limestone is absent, a dark green granular crystal tuff containing abundant epidote and chlorite lies in contact with the footwall rhyolite.

The lowest portion of the Arenillas Formation is host to the majority of bedded to massive sulfide mineralization, with other mineralized horizons in granular crystal tuff higher in the section.

Overlying the limestone is a middle section dominated by fine to medium-grained granular crystal tuffs 5 to 20 meters thick. This section is often weakly mineralized with thinly bedded, fine-grained galena and sphalerite. Interbeds of milky green crystal-lithic tuff may occur. These tuffs have much smaller percentages of crystals or lithic fragments than the granular tuffs and are never mineralized.

The upper part of the Arenillas also contains thin granular crystal tuff beds, but interbeds of coarser lithic lapilli tuff become increasingly common. This lithic tuff is similar in mineralogy and texture to the overlying Candelaria rocks. The contact with the Candelaria is therefore interbedded and the actual boundary is arbitrary.

Lithologic Description. The limestone unit contains three different varieties; massive or structureless, thin laminated, and thin

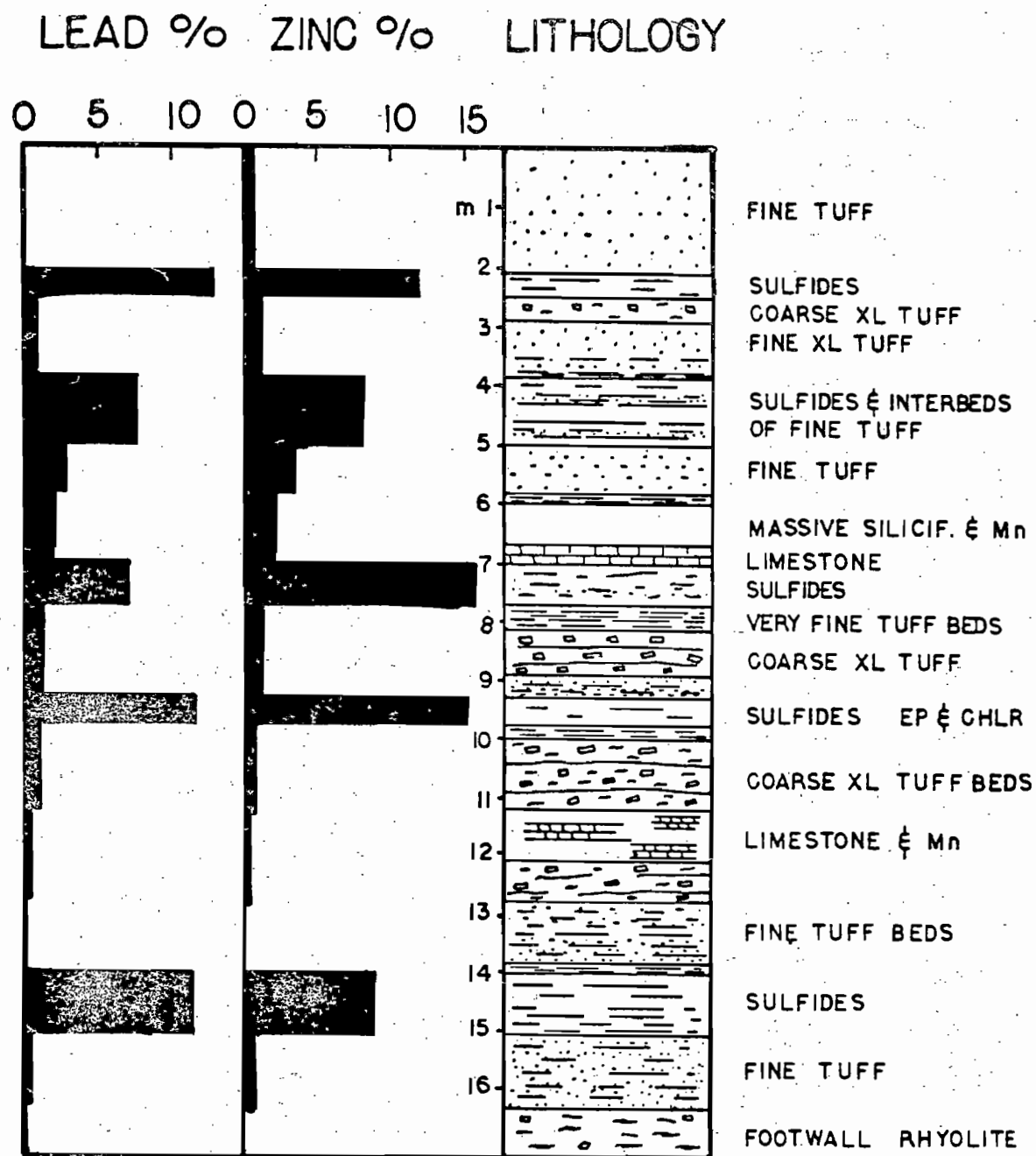


Figure 5 Stratigraphic Section of Mineralized Arenillas Formation, Tunnel 5

Detailed stratigraphy in the Tunnel 5 winze is correlated with lead-zinc assay intervals. Note the multiple sulfide horizons interbedded with crystal tuff and limestone.

laminated with algal growth structures. Thinly laminated limestone locally grades upward into massive partially recrystallized limestone. The thin laminated limestone is composed of 5-10 mm-thick beds of light gray carbonate alternating with light green clay beds (1 to 5 mm) or occasionally thin crystal tuff beds. The limestone is usually undeformed, but may be folded and contorted on a small scale and cut by small low angle faults. These structures are local over distances of several meters, are overlain by undeformed beds, and probably represent a syn-deposition disruption on a locally unstable slope. This irregular topography is reflected by rapid changes in the thickness of limestone outcrops and numerous lateral pinchouts of limestone into adjacent tuff beds.

No fossils were observed in any limestone beds, but distinctive algal structures are not uncommon. The laminated algal limestone

is composed of thin beds of clean, gray carbonate alternating with extremely thin layers of sandy or silty dolomite. The thin dolomitic layers form distinctive stromatolitic growth structures. Algal structures are indicative of a shallow, near surface, or intra-tidal environment.

The volcanoclastic and pyroclastic tuff beds of the Arenillas Formation are an important key to understanding both the depositional environment of the Oposura Group rocks and the mineralization which they host. Petrographic description, although not always in complete agreement with field terms, was useful in describing and differentiating various tuff beds.

The most common variety of volcaniclastic rocks are the dark green, granular tuffs of the middle unit. Different beds are easily recognized by variations in grain size, but beds are often irregular and can grade into one another. The equigranular texture, graded bedding, and penecontemporaneous faults led to the use of "volcanic sandstone", "siltstone", and "claystone" as field terms, but the lack of true clastic clay or micaceous particles and the abundance of minute crystals in the groundmass led to the use of "fine to medium grained crystal tuff" as a petrographic description. Discerning the difference between slightly reworked or locally transported tuffs from those of totally primary origin is difficult. This uncertainty is due to the general lack of published literature that discusses the origin of water-lain pyroclastic rocks.

The Arenillas tuffs are dark green in color and the matrix is fine grained to aphanitic. They contain abundant white feldspar crystals and quartz grains which comprise 60-80 percent of the rock. The grain size is never more than 1 millimeter and is usually 0.1 to 0.5 millimeters. A distinctive feature is thin (1-10 cm) pyroclastic interbeds composed of white feldspar crystals, rounded quartz grains, and epidote. The abundant epidote gives the bands a distinctive bright, pistachio green color. The beds locally display lenticular sedimentary boudinage-like structures which have a lensoid or oblate shape when viewed on a broken surface. These are probably loading or compaction structures. The granular tuff beds often show a size grading from fine-to medium-grained crystals. Feldspar grains are randomly replaced by epidote and the groundmass is often chloritic. Occasionally

load or piercement structures between two beds of different grain size or density are observed in drill core. Fragments of massive sulfide have been observed in the granular tuffs which indicates that some reworking has occurred on a local scale.

A second lithology is milky green crystal-lithic tuff.

This is a dense aphanitic rock with only sparsely disseminated feldspar crystals and lithic fragments which amount to only 10 to 20 percent. The dominant light green milky groundmass and lack of pyroclastic grains or fragments are distinctive characteristics of this lithology. Contacts between granular tuff and crystal-lithic tuff beds are usually sharp.

A third distinctive member of the Arenillas is the "salt and pepper" tuff or welded ignimbrite, best exposed in the Tunnel D workings. The "salt and pepper" tuff consists of layers of coarse-grained (1 to 3 mm) feldspars alternating with thinner, light gray, aphanitic ash beds. The feldspar crystals have diffuse boundaries, "melting" into the groundmass and creating a distinctive cloudy texture. Epidote grains are common and epidote replacing feldspars results in a white and yellow "salt and pepper" appearance. A faint size grading is sometimes apparent, and thin flattened chlorite fiamme are sprinkled through the matrix. The "salt and pepper" tuff may directly overlie massive sulfide mineralization, but the unit itself is never mineralized.

Scattered outcrops of bedded chert and massive magnetite beds, both up to 1 meter thick, were also observed in the Arenillas Formation. The chert often contains thin layers of fine grained pyrite.

Petrographic Description. The green granular Arenillas tuffs

are composed of 60-70% equigranular, well-sorted quartz and feldspar crystals that average 0.05-0.5 mm. The grains are suspended in a chloritic cryptocrystalline groundmass of quartz and feldspar. The groundmass may be up to 50 percent chlorite. Most crystals are rounded to sub-angular although some bladed crystals are present. Feldspars are often altered to epidote, or less commonly, carbonate.

The greenish-white lenticular zones are 80 percent epidote and 20 percent carbonate. Some relict feldspar crystals now completely altered to epidote are recognizable. The matrix resembles a microcrystalline ash where unaltered and embayed crystals are common.

The green crystal-lithic tuff contains 10 percent sub-rounded fragments of aphanitic volcanic rock and 15 to 25 percent minute (0.1 mm) quartz and feldspar crystals and glass shards. The crystals are poorly sorted, angular, and have sharp, distinct outlines. The devitrified, cryptocrystalline matrix sometimes exhibits a eutaxitic texture, bending around the feldspar crystals.

The distinctive, banded, coarse-crystal "salt and pepper" tuff has a matrix of extremely fine-grained interlocking quartz and feldspar crystals. The coarse-grained crystal-rich layers are dominantly feldspar (70%) and quartz (20%) with minor glass and chlorite. The gray ash between the coarse-crystal layers contains only 5 percent quartz and feldspar crystals, but has a matrix similar to the coarser layers. Albite-oligoclase is the most abundant feldspar and is commonly altered to epidote. Grain boundaries are diffuse into the groundmass

and ragged or broken crystals are common. Many of the broken crystals have adjoining rotated fragments which indicates that little movement has taken place since fracturing. Feldspar and quartz crystals often exhibit selvages of different extinction around the rims of the crystals. Some quartz grains have thin altered selvages of minute bubbles around the rims. Both features tend to round the crystals and produce diffuse boundaries.

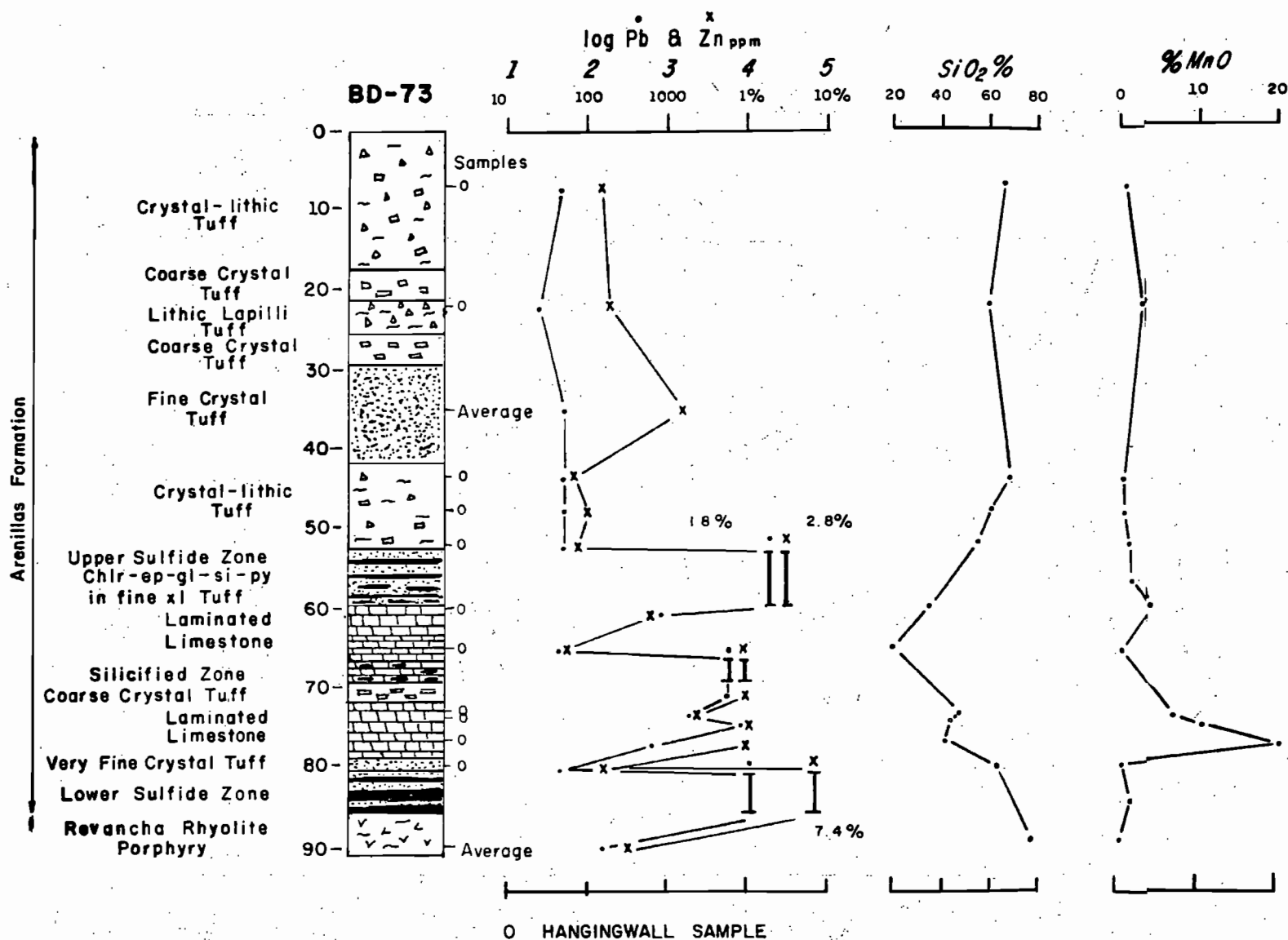
The broken crystals and devitrified glass indicate that the tuff is a primary ignimbrite with no reworking. The altered selvages indicate a rapid chilling during deposition.

MINERALIZATION

The copper-lead-zinc-silver mineralization at Oposura includes four separate major types of sulfide texture and mineralogy and several minor variations. Major types of mineralization are: 1) massive sulfides composed of galena-sphalerite-chalcopyrite-pyrite, 2) bedded semi-massive sphalerite-galena-pyrite interbedded with chlorite-epidote layers, 3) galena-sphalerite interbedded or replacing massive, thinly laminated or brecciated limestone, and 4) thin beds of very-fine-grained sulfides interlaminated with fine crystal tuff beds. Important amounts of manganese, either as rhodonite or rhodochrosite, are intimately associated with the mineralization.

Massive sulfide mineralization along the base of the Arenillas Formation is concentrated in two areas, one at the West End or old camp area and one at the Tunnel D area 1000 meters to the east. The West End has two or three distinct sulfide horizons exposed along a strike distance of 500 meters. The Tunnel D sulfide horizon is exposed for almost 600 meters of strike distance. The intervening distance contains thin, continuous mineralization with some locally important sulfide accumulations.

Massive sulfide bodies are stratiform and lenticular to podiform in shape, and they grade laterally to semi-massive sulfides and finally to thinly bedded sulfides on the fringes. All mineralization is stratabound, hosted by the Arenillas Formation. In the Tunnel D area, the

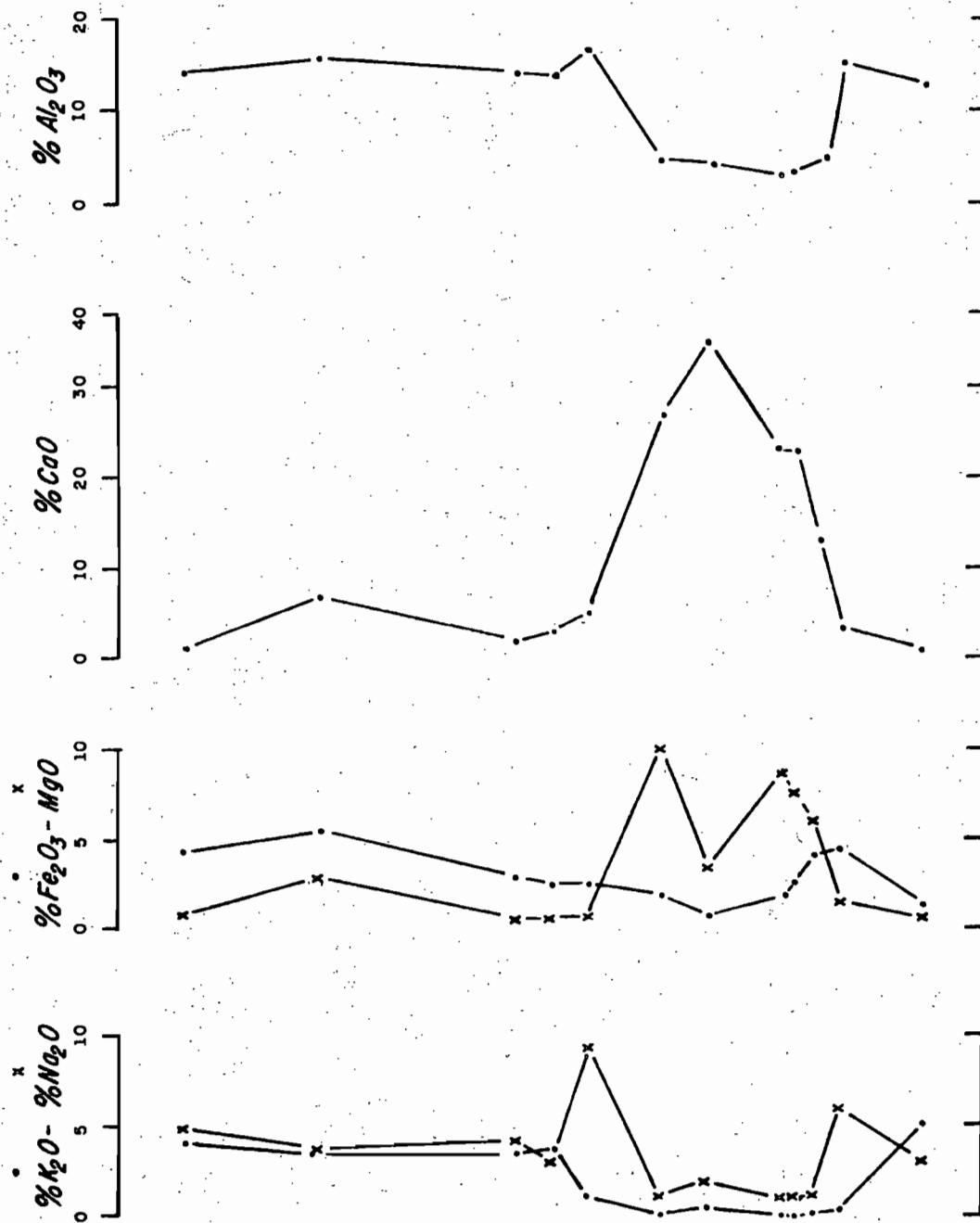


(Continued next page)

Major Element Comparison for Drill Hole BD-73

Major element values of selected core samples are compared to a geologic section of drill hole BD-73. Note enrichment of manganese associated with the mineralized Arenillas Formation.

Figure 6



(Continued from preceding page)

largest individual body, the massive sulfide horizon, is a minimum of 100 meters wide, 250 meters long and ranges in thickness from 3.5 to 0.5 meters. One, two, or sometimes three thinner and less extensive mineralized zones may occur stratigraphically above the main horizon.

Strataform Mineralization

Massive Sulfide

Massive sulfide mineralization may contain up to 90 percent sulfide minerals, dominantly sphalerite, galena, pyrite, and chalcopyrite, with trace amounts of argentian tetrahedrite-tenantite. It is generally granular, fine-to medium-grained (0.1 to 1 mm), and intergrown with chlorite and epidote with lesser amounts of magnetite, quartz, and carbonate gangue. True massive sulfide mineralization occurs principally in the Sonora Salon of Tunnel D in beds up to 3 meters thick, but only accounts for approximately 20 percent of the orebody. Examples of compositional banding in massive sulfide ore are minor. Where present, the bands are coarse-to-medium grained and have a higher-than-average copper content. The galena and sphalerite are fine-grained, but the pyrite usually occurs as coarse disseminated euhedral crystals.

Bedded Semi-massive Sulfide

This most common type of mineralization consists of 2 to 20 millimeter beds of massive, fine-grained sphalerite and galena alternating with layers of fine-grained chlorite, epidote, and K-feldspar.

This layering may give a banded appearance to the ore. Semi-massive bedded sulfides comprise up to 60 percent of the mineralization and have a larger areal extent than massive sulfide. Medium-grained blebs of pyrite (0.5 to 2 mm) are concentrated along chlorite layers or scattered throughout the matrix. In the West End, thin (1 to 4 mm) interbeds of magnetite are common. Occasionally these beds thicken to 0.5 meters. They are anomalous in gold, running up to 1.5 ppm Au. The style of bedding varies from thin, regular beds to thicker, wavy, and irregular beds. Sulfides usually make up to 20 to 50 percent of the rock; grade varies accordingly.

Some spectacular examples of semi-massive sulfide consist of beds of sphalerite-galena alternating with layers of dark green massive to very fine-grained chlorite and light pistachio-green epidote.

The chloritic layers are composed of small feldspar crystals altered to light green epidote and disseminated pyrite in a very fine-grained groundmass of almost black chlorite. The epidote-rich layers consist of feldspar grains altered to epidote in a matrix of epidote with minor chlorite and disseminated pyrite. Contacts of the beds range from diffuse to sharp and they may show size grading. Small scale slump folds, faults, and slip structures are common. Piercement and loading structures occur between beds of different densities, as for example between chlorite and sulfide beds.

Banding is often less well developed and consists of wavy, undulating layers of sphalerite and galena intermixed with layers of light green fine-grained chlorite and epidote. Relic feldspar crystals

replaced by epidote are common and may give the specimen a spotted texture.

A lower-grade variety of the semi-massive bedded mineralization is found in the central area between the two main centers of mineralization. These thin, weakly mineralized, green, bedded tuffs are lateral equivalents to the semi-massive type but have a smaller proportion of sulfide beds to unmineralized tuff. Multiple sulfide horizons are prevalent in the central mine area and the corresponding stratigraphy has thinner but more numerous granular green tuff beds which alternate with lithic or crystal-lithic tuff. The grain size of the sphalerite and galena can vary from very fine (0.1 mm) to coarse (1 to 2 mm), and is always hosted by chloritized and epidotized stratified volcanoclastic tuff beds or calcareous beds. Accessory minerals are 1 to 5% pyrite, rhodonite, hematite, and calcite. Rhodonite can occur as irregular splotches or laminae. The mineralized tuff beds vary in thickness from 10 centimeters to 3 meters.

Petrographic Description. The sulfide layers are composed of medium-grained galena and sphalerite (0.2-0.5 mm) in a matrix of chlorite and epidote. The epidote and chlorite content varies from 5 to 25 percent. Boundaries between the sulfides are usually straight, although galena has been observed invading sphalerite. Sphalerite may have minute solid solution blebs of chalcopyrite, usually aligned into parallel trains. No sulfosalts were positively identified from the polished surfaces. Pyrite is more abundant in the chloritic tuff layers than in the sphalerite-galena-chalcopyrite zones.

A layer of massive epidote-chlorite underlying the massive sulfides contains relict subangular feldspar crystals. The feldspars are replaced by blebs of epidote, but in plane polarized light the feldspar outlines suggest a pyroclastic texture. The groundmass is totally altered to an unoriented, felty mass of interlocking dark blue chlorite, probably penninite or prochlorite. Pyrite grains (23%) have corroded boundaries due to replacement by epidote and chlorite. The pyrite is disseminated throughout the chlorite matrix.

Carbonate Mineralization

The limestone beds of the Arenillas Formation form a very distinctive lithologic unit in a stratigraphic section dominated by volcanic rocks. The altered and mineralized portions of the limestone are even more conspicuous and have received most of the attention devoted to the Oposura mineralization by previous workers. However, this type of mineralization is volumetrically subordinate to volcanoclastic-tuff-hosted mineralization, comprising 10-20 percent of the total.

The limestone is either massive, thinly bedded, or brecciated. Brecciated limestone consists of tabular fragments of limestone beds in a carbonate matrix. All three types may be mineralized and altered or in other areas nearly fresh and only moderately recrystallized. Massive limestone, when mineralized, consists of irregular zones of coarse grained galena, sphalerite, chalcopyrite, and pyrite. It is usually silicified to some degree and contains abundant manganese as rhodonite or rhodochrosite, in epidote, or as minor bustamite. Bustamite is a pale pink, flesh-colored calcium manganese pyroxenoid silicate

composed of radiating crystals and was identified by X-ray (H. Salek, oral comm.) . Thin veinlets of purple fluorite may cut irregular blebs of epidote. Black manganese oxide, probably pyrolusite, is ubiquitous on weathered surfaces of mineralized limestone.

Thin-bedded limestone is sometimes replaced by rhodonite and the matrix replaced by galena, sphalerite, pyrite, and chlorite. This type of mineralization forms a distinctive pink and gray "zebra" textured rock and is referred to as "bedded manganese silicate mineralization". It is much more common than the massive type, and is generally interbedded with chlorite-epidote mineralization.

The rarest but most spectacular carbonate mineralization is the tabular breccia type. Thin, tabular breccia fragments of limestone are replaced by bright pink rhodonite or occasionally by galena and sphalerite. The matrix is black chlorite, light green epidote, and disseminated pyrite. This type is commonly found in the Tunnel D workings. Both the laminated limestone and tabular limestone breccia may be interbedded with green granular crystal tuff, the sequence grading upward from medium to fine grained crystal tuff and thin ash beds to mineralized limestone. Mineralized breccia zones usually occur immediately above semi-massive chlorite-epidote sulfide layers.

Sulfides in Fine Crystal Tuff

Thinly laminated crystal tuff may contain interlaminae of fine-grained galena and sphalerite. This type of mineralization occurs in the Candelaria Lithic Tuff as a distinct upper sulfide horizon below survey station V-5 on the upper slopes of Cerro Oposura.

Other exposures of this type are located in the central zone near Tunnel #3. The galena, sphalerite, and pyrite are found in an extremely fine-grained to aphanitic, thinly laminated, green crystal tuff.

A weathered sample resembles a thin-bedded argillite and some exposures were mapped as such. The rock is well-indurated and silicified, breaking in a conchoidal fracture.

Thin layers of green crystal tuff, 1 to 10 millimeters thick, are composed of minute feldspar crystals in a light green aphanitic matrix and are interbedded with slightly thinner 1 to 2 mm laminae of pink rhodonite. Galena and sphalerite grains are associated with the weakly calcareous manganese-rich layers. The individual layers are very regular with no apparent size grading, but are deformed by frequent syn-sedimentary slumps and faults. Manganese oxide is abundant along the layers on weathered surfaces. The lead-zinc values of this type of mineralization are relatively low, ranging from 0.2 to 1% Pb and 0.2 to 3% Zn. Silver is anomalous but copper is significantly low (2 to 8 ppm).

Several thin but higher grade chlorite-epidote sulfide zones were mapped stratigraphically just above the thin laminated tuffs, but they are not very extensive. This upper sulfide horizon does not appear to be directly related to the Revancha Rhyolite or the Arenillas Formation, but is a separate and higher-level mineralized horizon. The individual lithologies are very similar to Arenillas-type volcanoclastic tuff beds a short distance down section.

Petrographic Description. In thin section, the feldspar-rich layers were found to contain tiny, 0.04 to 0.15 mm, irregular albite

grains altered to carbonate. The surrounding matrix is almost micro-crystalline and also contains very fine grained carbonate. Under high power (500X), the matrix proved to be a mosaic of interlocking grains of feldspar, sericite (?), and illite (?). The sericite-illite grains are probably secondary. In addition, blebs of opaque leucoxene are common.

The manganese layers are composed of a randomly oriented, interlocking aggregates of equigranular prismatic or bladed crystals approximately 0.05 millimeters in diameter. These euhedral crystals have high relief and a high extinction angle ($\sim 45^\circ$) and are similar to much coarser-grained crystals in other samples identified as bustamite [$\text{CaMn}(\text{SiO}_3)_2$]. Blebs of carbonate (0.5 mm) are also common and may replace the prismatic manganese minerals.

Vein Mineralization

Several different types and ages of vein mineralization were mapped in the Oposura district and vicinity. These have entirely different mineralogy, texture, and rock association than the Oposura strata-form massive sulfide deposit. All are structurally controlled by fault or fracture zones and have a definite vein morphology.

Quartz-Pyrite Veins

By far the most extensive vein mineralization is the quartz-pyrite system associated with the large northwest-trending faults. Closely spaced, parallel quartz-pyrite veins were mapped at several localities along the mine access road associated with the intense fracturing of the Rancho Arenillas fault. A typical vein is 1 to 5 millimeters wide with a dark gray quartz-sericite alteration envelope 5 to

10 millimeters wide. Vein swarms up to 20 meters wide were observed, usually completely obliterating the host rock texture. Unweathered samples display pervasive bleaching and surface specimens show strong staining by hematite and jarosite.

Numerous prospect pits are located along the Rancho Arenillas fault trace, both in fault gouge and intensely veined rock. The largest prospect is the Los Defuntos adit on the mine road just north of Rancho Arenillas. Little effort was expended studying these occurrences, but one random grab sample of fault gouge from the Los Defuntos adit was analyzed. It contained 0.04 oz/ton gold, 1.0 oz/ton silver, 0.1% Pb, and 0.2% Zn.

A second major zone of quartz-pyrite veins is associated with the Arroyo de Campo fault where veins are numerous even 100 meters from the fault zone itself. Vein swarms cutting through intensely fractured rock are similar to those several kilometers east along the Rancho Arenillas fault, though not as extensive. Drill core specimens from DDH OP-9 located 150 m east of the fault contained abundant quartz-pyrite veins and occasional specks of molybdenite. All exposures of quartz-pyrite veining occur in rocks of the Oposura Volcanic Group.

Fluorite Veins

Minor fluorite mineralization was mapped along a northwest trending fault approximately 1 kilometer west of the mine road in the northern part of the district. Veins containing purple and green fluorite were observed, all less than 10 centimeters in width.

Projections both north and south are lost under talus. The fluorite veins occur in Candelaria Lithic Tuff.

Quartz-Silver-Lead Veins

The La Blanca mine is situated at the base of the western flank of the Sierra La Huerta, nearly 5 kilometers due west of the Oposura camp. The La Blanca vein is localized in La Huerta Dacite along the range front fault and has been mined on a small scale to depths of over 50 meters. The vein is composed of white bull quartz and contains coarse-grained sphalerite, galena, and unidentified silver-bearing sulfosalts over widths of several meters. Bright blue covellite partially replaces sphalerite. Samples of randomly selected vein material assayed 4.9 oz/ton silver, 0.2 oz/ton gold, 0.9% lead, and 0.2% zinc.

Gold-Silver-Telluride Veins

The Bomboya mine, also known as the Moctezuma mine, is 8 kilometers south of the Oposura camp along the general trend of the Rancho Arenillas fault system. It is described by Gaines (1965) to be a tellurium-gold deposit, localized by quartz veins in highly altered bedded tuff. The host tuff unit is folded and faulted and described as probably Late Cretaceous in age. Younger andesitic flows were also noted by Gaines. Mineralization includes native tellurium, native gold, pyrite, barite, cinnabar and several exotic newly described minerals including moctezumite, a lead uranyl tellurite.

Clark (1976) noted that all the better known gold-silver districts of the Sierra Madre (Ocampo, Batopilas, Lluvia de Oro) are found in the Lower Volcanic Group.

Uranium Veins

A prospect trench is located on a small west northwest-trending fault-vein structure in the Rodeo Ignimbrite approximately 1 kilometer east of the Los Defuntos adit. Along the vein the Rodeo tuff is bleached and the pumice fragments are altered to a light green clay. Uranium mineralization, possibly torbernite or uraninite, has been reported in the prospect trench (D. Gomez, oral commun.). Fissure-vein uranium deposits in Chihuahua occur in rhyolitic tuffs of Eocene age (Clark, 1976).

Arenillas Formation

Deposition of the Arenillas Formation marks an abrupt break from rhyolitic volcanism to shallow water sedimentation. However, bursts of volcanism continued, producing the fine crystal tuffs and banded pyroclastics. Sedimentary structures, well-bedded texture, and petrographic evidence all indicate that the Arenillas tuffs were laid down in a shallow water environment. Tuffs may be interbedded with the limestone.

Because the limestone immediately overlies the rhyolite in a depositional contact, it could be argued that the Revancha Rhyolite was also deposited in a subaqueous environment. The amount of time between extrusion of the rhyolite and deposition of the limestone and bedded tuffs is probably short, but the timing can only be speculated upon.

Variations in the limestone lithology indicate changes from an algal-rich intertidal zone to a shallow sub-tidal zone represented by carbonate laminae alternating with silt and finally to slightly deeper-water, more massive limestones. These variations occur over lateral distances of several hundred meters and reflect an irregular depositional surface. The effect of this irregular topography is also demonstrated by the lensy nature and lateral variability of the bedded volcanoclastics. Correlation of individual units across the deposit is difficult because of continual thinning, thickening, or lateral pinch outs.

During the latter part of Arenillas deposition, explosive and fragmental volcanism resumed, represented by lithic lapilli tuff. Although fragmental Candelaria type volcanism becomes dominant, thin volcanoclastic beds of fine granular tuff still persist up into the lower parts of the Candelaria Lithic Tuff. This transitional contact represents a gradual change without break from Arenillas-style volcanism to Candelaria-style volcanism. The thin-bedded tuffs of the V-5 mineralized area indicate a local resumption of Arenillas-type volcanic activity and sedimentation.

GENETIC MODEL OF MINERALIZATION AND INTERPRETATION

The Oposura deposit is a member of the large class of strataform base metal deposits that formed syngenetically with their enclosing host rocks. It is a product of the intermingling of volcanic and sedimentary processes in an anomalously metal-rich environment.

Interpretation of Oposura stratigraphy gives rise to the following geologic setting. After eruption of the Revancha Rhyolite as a hot, pumice-rich, felsic ash flow, a system of fumaroles and hot springs developed at its surface. Most of the rhyolite was submerged in shallow water but some parts may have been exposed. While the rhyolite was cooling and the fumarolic activity continuing, limestone was deposited in a shallow-water environment. Deposition of limestone ranged from that of an intertidal zone supporting algal growths to a slightly deeper zone receiving a minimum of clastic input. Occasional small eruptions produced crystal tuffs that were deposited in the shallow water as bedded tuff deposits over the irregular topography of the rhyolite. Slumping of the volcanoclastics locally created syn-depositional folds or faults.

Hydrothermal fluids generated in the hot springs and related fumaroles formed pools in some areas and generally percolated through the porous tuff beds at the surface. Hot spring emanations from similar environments at Katmai, in the Valley of 10,000 Smokes (Zies, 1929), and

Matupi Harbor, New Britain (Ferguson and Lambert, 1972) have been shown to carry anomalous amounts of magnesium, calcium, potassium, iron, manganese, copper, lead, and zinc. The fumaroles also contain H_2S , CO_2 and HCl . Perhaps due to modification of Eh and pH of the waters at the surface, the contained metals were precipitated. The actual products of precipitation are of course unknown but they may have been amorphous iron sulfides enriched in manganese, zinc, and lead.

Mineralization at Oposura

The proximity of the limestone created an ideal location for replacement of recently deposited carbonate by lead, zinc, and especially manganese, and probably the direct precipitation of base metals. Depending on the location of a fumarole, hot spring, or local topography, either limestone or tuffs or both might be altered and mineralized. Sufficient temperatures could easily be generated by the fumaroles for alteration of the limestone. The abundance of lead and zinc sulfides in the tuffs probably reflects their porous, semi-consolidated nature as inviting hosts for the mineralizing fluids. Intense hydrothermal activity immediately preceding deposition of the metals altered the thin crystal tuff beds to massive chlorite and the feldspar to epidote. A pulse of mineralization followed, creating a lead-zinc-rich layer overlying the chlorite bed. In the Tunnel D area and elsewhere, a brief pyroclastic eruption then laid down a coarse "salt and pepper" crystal tuff, creating the stratigraphic sequence so common at Oposura. In areas adjacent to volcanic vents, tuffs being deposited at

the surface and the underlying rhyolite were intensely altered and silicified. The massive silica in the cross cut north of the Sonora Salon and the bedded cherts observed in OP-1 are products of this alteration. In marginal areas, massive magnetite sinter deposits were formed, such as the gold-bearing magnetite beds of the West End.

The stratigraphy suggests a complex volcano-sedimentary environment of several adjacent but possibly interconnected centers of mineralization (Fig. 7). Although the same processes were operating in many areas at the same time, the depositional sequence varied from place to place. The thin, multiple sulfide horizons probably formed slightly distally to the mineralizing centers, or perhaps in areas where volcanoclastic and pyroclastic input was more rapid. This pyroclastic input would dilute the metal-rich layers, creating thinner interbedded sulfides and tuff instead of thick semi-massive sulfides.

Erosion of pre-existing mineralized layers occurred locally, demonstrated by the small manganese-lead-zinc rich fragments contained in overlying unaltered volcanoclastic beds. This process probably occurred over short distances and involved no outside clastic input. The resulting rock would be very difficult to recognize from unworked tuffs of the same composition, except for the presence of sulfide fragments.

The formation of metal-rich accumulations in modern day volcano-sedimentary systems is recognized to be more a function of the right depositional environment combined with metal-rich sources than just the

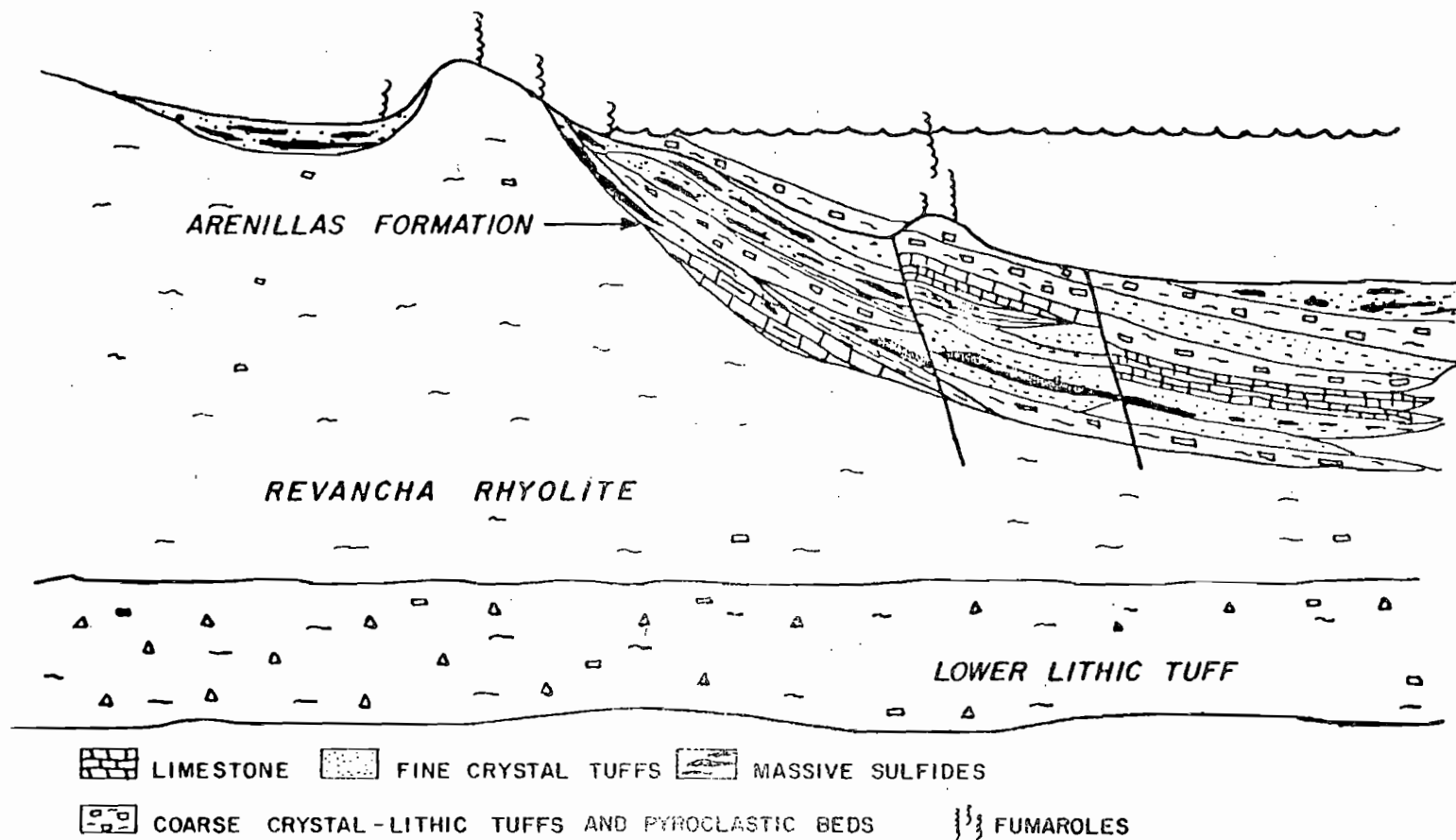
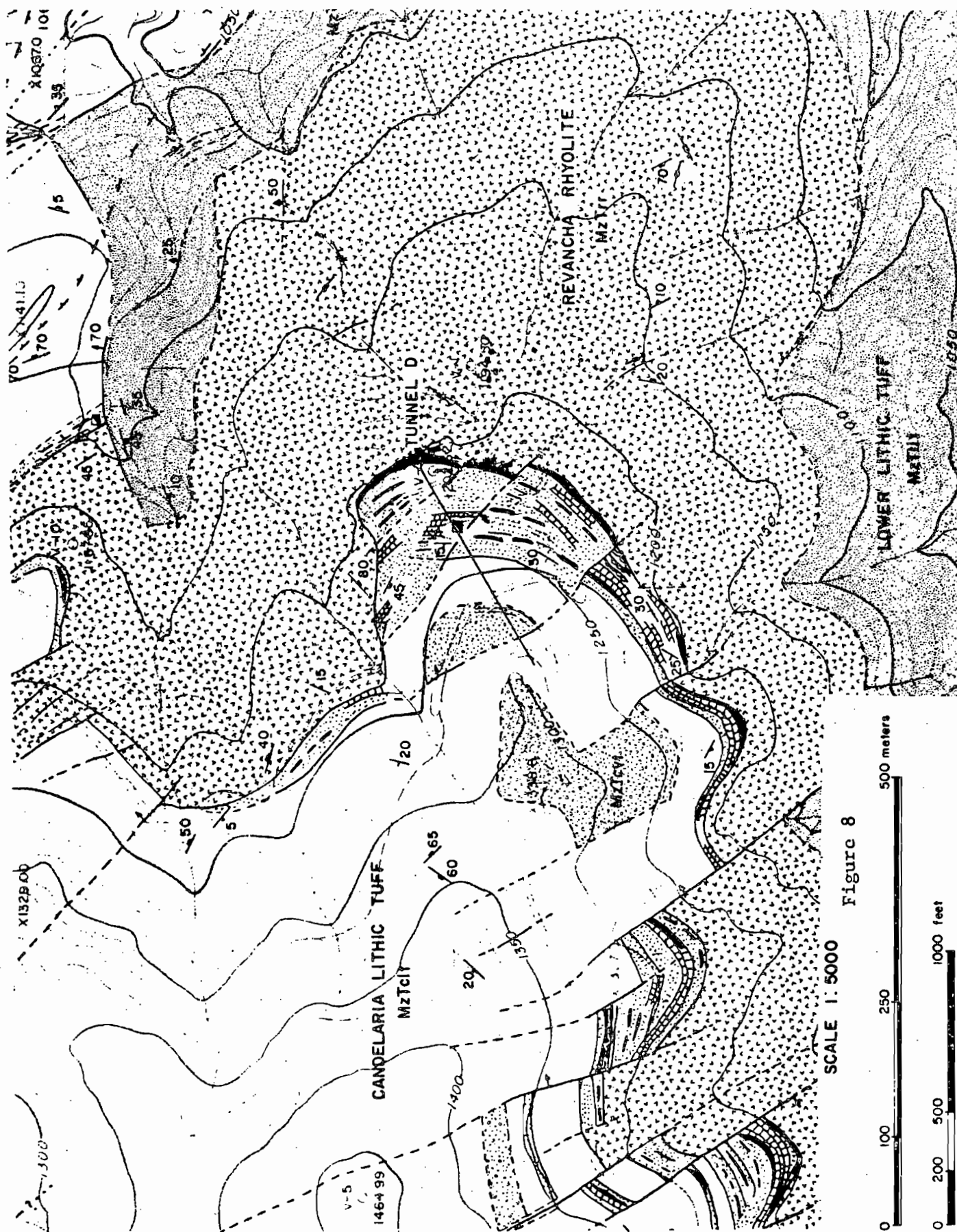


Figure 7 Oposura Genetic Model

Genetic model for deposition of the Oposura massive sulfides and the enclosing volcanoclastic Arenillas Formation.



presence of mineralizing fluids alone. Metaliferous deposits are only formed when the influx of clastic or volcanic material is low in relation to the input of metal rich exhalations.

At Oposura, the mineralizing episode was terminated by the eruption of large volumes of dacitic lithic lapilli tuff. After several small spurts of lithic tuff, which became interbedded with the upper part of the Arenillas crystal tuffs, the Candelaria Lithic Tuff event overwhelmed the existing Arenillas depositional environment and ended formation of metal-rich deposits.

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PROYECTO OPOSURA

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Hermosillo, Son., Méx.

R E S U M E N

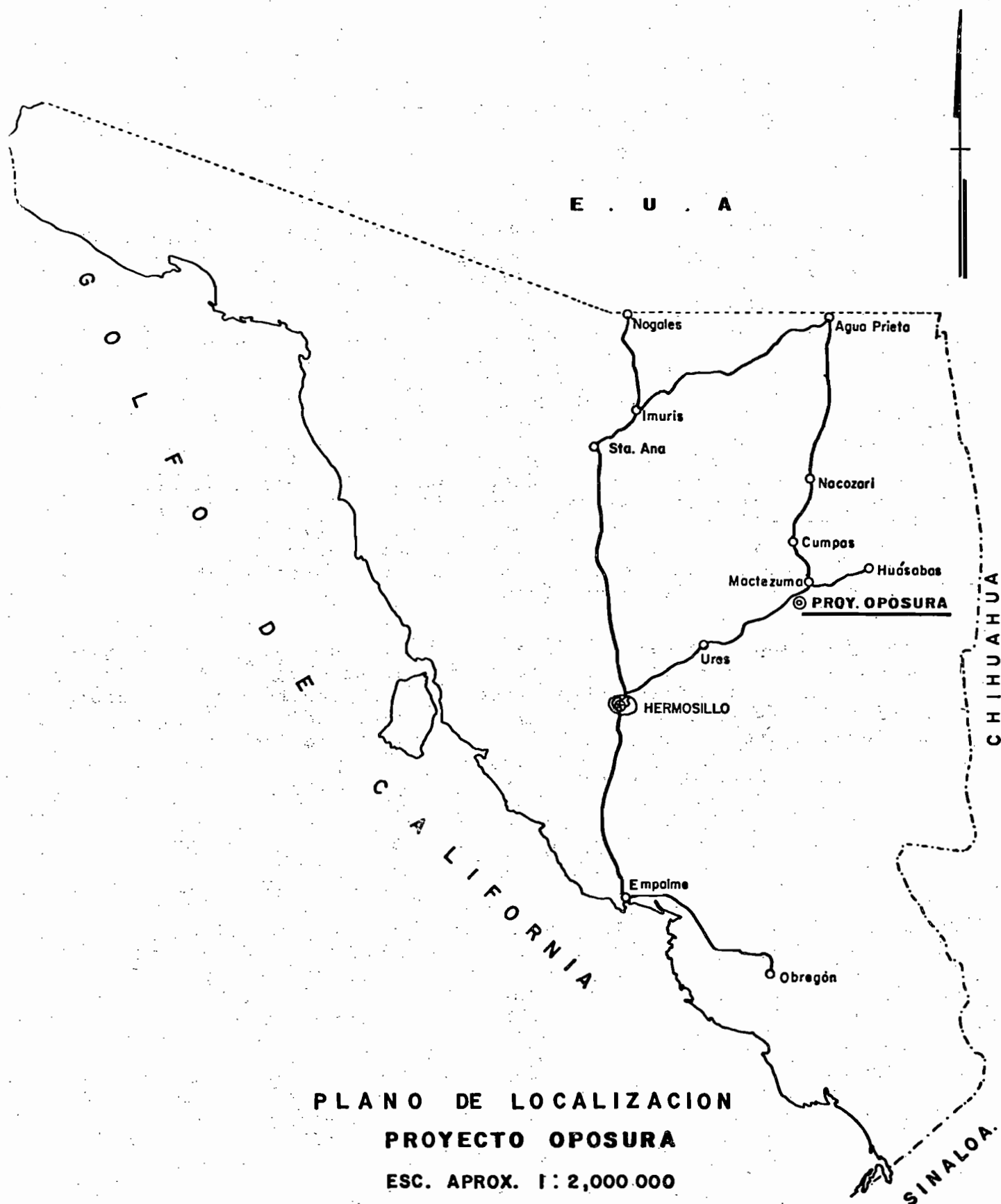
Oposura, antiguamente conocida como mina de Las Arenillas, es un yacimiento con sulfuros masivos, semi-masivos, diseminados y retrabajados dispuestos en forma de capas o mantos, habiendo también algunas vetas angostas. Galena, esfalerita y pirita son los sulfuros principales de la mena, con calcopirita subordinada. Contiene en promedio bajas cantidades de plata.

Varios geólogos consideran a éste un yacimiento tipo skarn, mientras que otros lo consideran volcanogénico. En opinión del autor, las dos formas de origen están presentes pero fuertemente relacionadas al origen volcánico félsico y sedimentario de la secuencia presumiblemente Cretácica que contiene la mineralización.

S U M M A R Y

Oposura, named before as Las Arenillas mine, is a stratiform deposit of massive, semi-massive, disseminated and reworked sulfides. Some barren veinlets with sulfides are present. Galena, sphalerite and pyrite are the principal components of the ore, with subordinated chalcopryrite. The average contents of silver is low.

Some geologists think that Oposura is a skarn tipe deposit, and many others think that it is a volcanogenic one. In the author's opinion both forms of origin are present, but strongly related to a felsic volcanic and sedimentary events in a possible Cretacic sequence that hold the mineralization.



LOCALIZACION Y ACCESO

Oposura se localiza a 163 km al Nor-este de Hermosillo, capital del estado de Sonora, México, y a 22 km al Suroeste de la población de Moctezuma.

El acceso se hace a partir del kilómetro 157 de la carretera Hermosillo-Moctezuma, donde se toma a la derecha un camino de terracería transitable todo el año, con una extensión de 6 km.

ANTECEDENTES

Este proyecto fue trabajado de 1920 a 1930. En 1934 lo exploró la - Compañía Real del Monte y Pachuca. De 1947 a 1952 la Cananea Consolidated Copper Company efectuó trabajos de exploración subterránea y barrenación.

Hasta 1970 se contaba con 51 barrenos de superficie y 1085m de cuele subterráneo.

En tres etapas de exploración comprendidos de septiembre de 1975 a junio de 1980, la sociedad Peñoles-Anaconda ha efectuado una exploración integral del Proyecto habiéndose efectuado 22 barrenos que suman 2482m perforados.

EXTENSION DEL AREA MINERALIZADA

La mineralización económica está circunscrita a una superficie de 1400m de largo por 400m de ancho, elongada al NE, donde la mineralización está mayormente concentrada en los extremos a los que se conoce como Oposura Oeste y Oposura Este.

RESERVAS

Las reservas cubicadas actualmente son del orden de 800,000 toneladas probadas con un espesor de 3.60m y leyes de 19 gr/Ton. de Ag, - 3.5% de Pb y 6.04% de Zn.

ESTRUCTURA

Este proyecto se localiza dentro de la Subprovincia Fisiográfica de

Sierras y Valles Paralelos y se encuentra enmarcado por un sistema de fallas regionales de rumbo N 20° W que delimitan la Sierra de la Huerta, paralela al río Moctezuma.

Esta sierra constituye un horst formado por una potente serie de rocas volcánicas y piroclásticas que incluyen algunos horizontes de rocas sedimentarias tales como grauvacas, calizas y pedernales. La base de esta secuencia incluye tobas líticas dacíticas, riolitas blancas y aglomerados tobáceos de posible edad Cretácica, mientras que la parte superior contiene andesitas y riolitas Terciarias.

Localmente el sistema de fallas NW es predominante, pero está cruzado por otros sistemas de rumbos NE y WNW que complican fuertemente la estructura, desplazando a muy diferentes niveles a los mantos de mineral que son sensiblemente horizontales con pliegues ligeros y echados fuertes locales.

La similitud entre unidades de diferente nivel estratigráfico y el intrincado sistema de fallas ha impedido hacer una buena correlación entre ellas quedando todavía mucho trabajo que hacer a este respecto, sin embargo, las dos unidades que subyacen a la mineralizada están reconocidas como las mas antiguas que afloran, limitadas por bloques de rocas mas jóvenes que han caído por efecto de fallas normales conocidas como: falla del Aguaje (al Oeste), falla del Rancho Arenillas (al Este) y falla del Campo (al Sur).

GEOLOGIA LOCAL

Estratigrafía.

Como se dijo antes, las rocas mas antiguas son unas tobas líticas y feldespáticas que forman la base del cerro del yacimiento.

Sobreyace a esta unidad un estrato de unos 90 metros de espesor - compuesto por riolitas porfídicas blancas, ignimbritas y tobas líticas blanqueadas por una fuerte silicificación. Su contenido de pirita diseminada es de 1 a 3%.

Descansando sobre la unidad anterior se encuentra la Formación Oposura que contiene a la mineralización. Está formada por grauvacas, calizas y pedernales lenticulares, y tobas líticas y feldespáticas intercaladas de composición dacítica.

Los mantos de sulfuros se encuentran entre las grauvacas que infrayacen o suprayacen a las calizas o sustituyendo total o parcialmente a estas últimas. También se observan en fragmentos de esas rocas erosionados y redepositados entre las tobas líticas posterior-

res (Candelaria Inferior).

Sobre el paquete de tobas y aglomerados de Candelaria Inferior descansan tobas feldespáticas o de cristales soldadas denominadas Candelaria Superior.

A raíz de observaciones someras recientes se ha pensado que puede ser un domo riolítico el gran afloramiento de riolitas porfídicas blancas que aflora hacia el Poniente del área mineralizada, el cual pudiera tener relación con el origen del yacimiento. También cabe la posibilidad de que se trate de un potente derrame riolítico como se ha interpretado anteriormente.

GEOLOGIA DEL YACIMIENTO

Rocas Encajonantes.

La mineralización de sulfuros masivos estratiformes de Oposura está distribuida hacia la base de una secuencia piroclástica de composición dacítica que contiene grauvacas, calizas, areniscas, pedernales y tobas líticas, todas de carácter lenticular, la cual descansa sobre una toba riolítica a cuarzo latítica casi estéril, de actitud general horizontal.

Los mantos y lentes de sulfuros están distribuidos verticalmente dentro de un espesor de unos 35 metros, con bruscos cambios laterales en ley y espesor, existiendo partes donde éstos cuerpos se acúñan - hasta desaparecer o persistiendo sólo uno de ellos.

Hacia los extremos Este y Oeste de Oposura los mantos son más potentes y de mas alta ley que en la porción central donde los sulfuros están dispersos en un espesor mayor de sedimentos que diluyen su ley.

Mineralogía.

Los minerales presentes en la mena son galena, esfalerita y calcopirita, pirita, magnetita, especularita y rodonita o tulita principalmente. Como minerales de ganga se tiene cuarzo, clorita, epidota, calcita y fluorita. En un solo barreno se identificó wolastonita y en exámenes petrográficos se ha reportado la presencia de granates microscópicos en muestras de Oposura Este colectadas en el socavón "D".

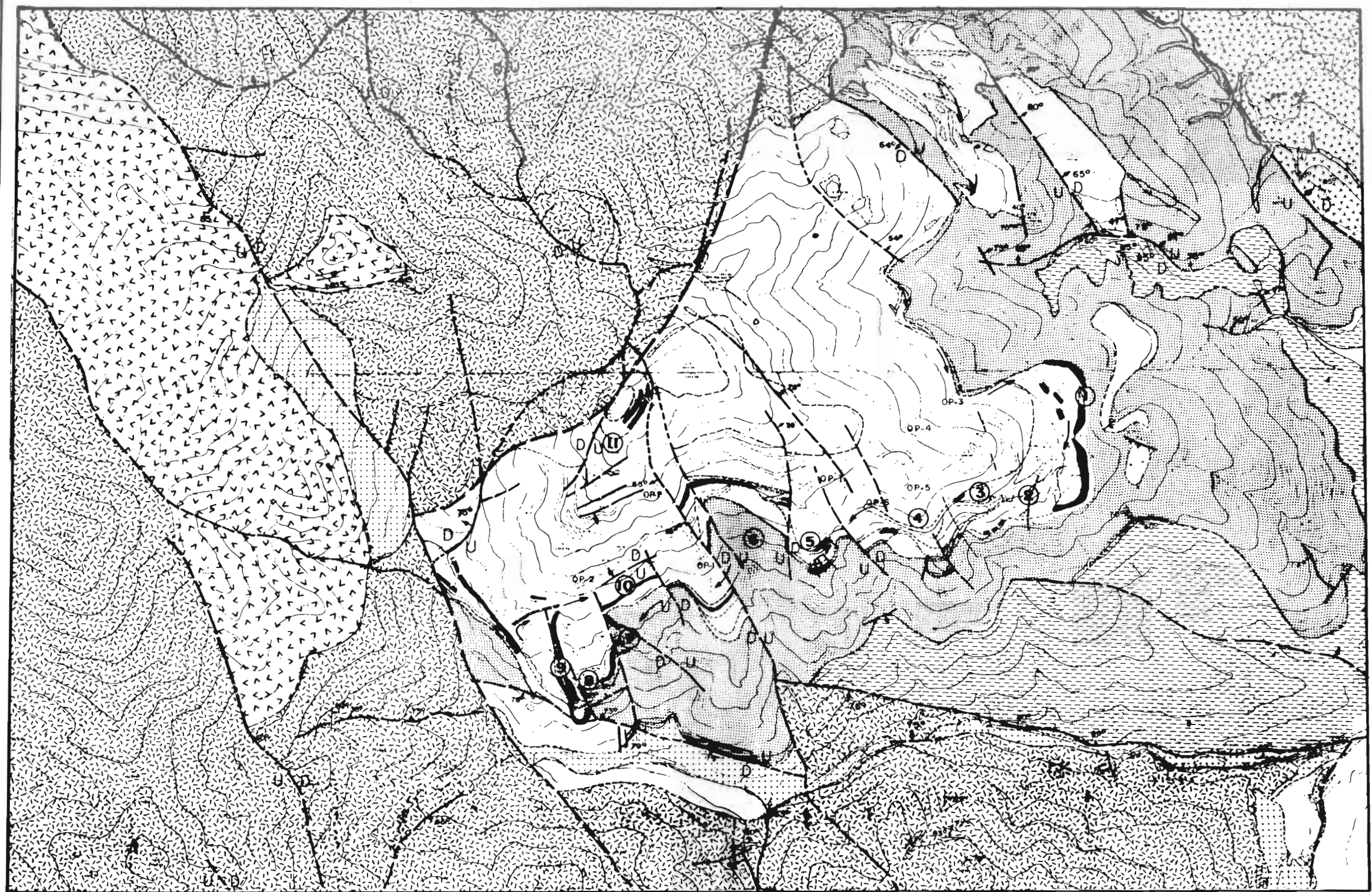
Controles de la mineralización.

Es incuestionable la existencia de sulfuros en bandas intercaladas

entre capas de grauvacas y calizas. En estas últimas se aprecian - bandas de mineral y de caliza deformadas y brechadas por deslizamiento durante la etapa plástica de los sedimentos; sin embarbo, - es comun encontrar a los sulfuros encajonados en grauvacas que yacen inmediatas abajo o arriba de lentes de calizas estériles.

La amplia distribución horizontal de los mantos de sulfuros y el - estar frecuentemente ligados a silicatos de manganeso son otros de los factores que determinan el origen singenético exhalativo de - los mismos.

Por otro lado, la presencia de granate microscópico y muy rara de wolastonita indica que hubo también sustitución de rocas calcáreas posiblemente cerca de las fuentes mineralizadoras, las cuales no - son conocidas.



E X P L I C A C I O N

 Aluvión y Talud.	 Toba riolítica (Ignimbrita)	 Secuencia Piroclástica Sedimentaria Oposura.	 Tobas líticas y feldespáticas
 Derrames de And. porfídica	 Tobas líticas y feldespáticas jóvenes.	 Toba Riolítica	 Afloramientos de mineral.

SERVICIOS INDUSTRIALES PEÑALES S.C.

PROYECTO OPOSURA

PLANO GEOLOGICO SIMPLIFICADO

② PUNTOS DE INTERES

DICIEMBRE DE 1977 D.G.H.

PUNTOS DE INTERES SEÑALADOS EN EL PLANO

- 1.- Portal del socavón "D" y afloramiento del manto en contacto con la toba riolítica.
- 2.- Manto de sulfuros asociado a calizas y silicatos de Mn.
- 3.- Punto en que se acuña una gruesa lente de calizas laminares plegadas, intercalada entre grauvacas.
- 4.- Lente de tobas feldespáticas soldadas (cristal tuff).
- 5.- Obra minera en bloque fallado de mineral de alta ley asociado a magnetita.
- 6.- Toba riolítica (Foot wall por phry).
- 7.- Bloques fallados de mineral con óxidos de manganeso.
- 8.- Mantos de sulfuros, calizas plegadas y brechadas durante su etapa plástica, tobas líticas gruesas y pequeñas obras mineras.
- 9.- Sulfuros interestratificados en las obras 10 y 11.
- 10.- Manto de sulfuros en grauvacas y socavón No. 5 con un pozo en su tope de 17m de profundidad, el cual cortó 5 mantos de alta ley con espesores de 0.40m a 1.20m.
- 11.- Mantos de pedernal bandeado con trazos de sulfuros diseminados, descansando sobre un conglomerado basal. Representa una etapa tardía de sedimentación y mineralización.

OPODEPE MOLYBDENUM-COPPER PORPHYRY DEPOSIT

FIELD TRIP No. 7 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 24

TRIP LEADERS:

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GEOLOGY OF THE CRESTON MOLYBDENUM-COPPER DEPOSIT

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AMAX Exploration, Inc.
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The Creston molybdenum-copper deposit is about 70 kilometers northeast of Hermosillo near the town of Opodepe in the State of Sonora, Mexico. It is within a roof pendant complex resting on locally differentiated batholithic rocks of Tertiary age. Molybdenum mineralization occurs dominantly as quartz-molybdenite stockwork set in pre-batholithic rocks which show hydrothermal sericitic alteration and silicification. The economic copper mineralization occurs as irregular zones of supergene enrichment which are spatially above and straddle the molybdenum ore zone.

The project area has a complex postmineral structural history. The known Creston deposit is interpreted as being only a portion of a large mineralized system which was sliced by postmineral, generally low-angle faults.

RESUMEN

El depósito de cobre y molibdeno denominado el Crestón está ubicado a mas ó menos 70 kilómetros al nordeste de Hermosillo, capital del estado de Sonora, Mexico. El depósito esta localizado en un complejo de rocas metamórficas que ocurren como un techo colgante que descansa sobre rocas batolíticas diferenciadas de edad Terciaria. La mineralización de molibdeno ocurre generalmente en forma de vetillas de cuarzo y molibdenita en forma de "stockwork" emplazadas en rocas pre-batolíticas las cuales exhiben alteración hidrotermal del tipo filico y silicico. La mineralización de cobre de grado económico ocurre en zonas irregulares de enriquecimiento supergénico las cuales están sobrepuestas en la parte de arriba y separadas del cuerpo de molibdeno.

El area del proyecto tiene una historia estructural post-mineral muy compleja. El deposito del Crestón hasta ahora conocido es interpretado como una fracción de un sistema mineralizado de proporciones grandes el cual fué segmentado por fallas post-minerales de bajo ángulo.

INTRODUCTION

The Creston molybdenum-copper prospect is in northwestern Mexico about 70 kilometers northeast of Hermosillo, the capital of the State of Sonora. It is readily accessible from the village of Opodepe, which is connected with Carbo to the west by an all-weather gravel road (Figure 1). The best way to get to the prospect from Hermosillo is through Carbo; this involves a distance of approximately 130 kilometers, or a driving time of about 2.5 hours. A gravel airstrip for relatively small aircraft was constructed near the town of Opodepe during the exploration activities at the property.

The property is controlled by Minera Opodepe, S. de R. L. de C. V., a Mexicanized company in which Industrias Peñoles, S. A. holds a 60% interest and AMAX, Inc. controls the remaining 40%. At the present time, the property is being administered by Compañía Minera Fresnillo, S. A., which is a subsidiary of Peñoles (60%) and Rosario Resources (40%). Rosario Resources is fully owned by AMAX.

The authors wish to express their appreciation to AMAX, Fresnillo, Peñoles, and Rosario Resources for granting permission to publish this paper. Javier Contreras of the AMAX Tucson office prepared the illustrations.

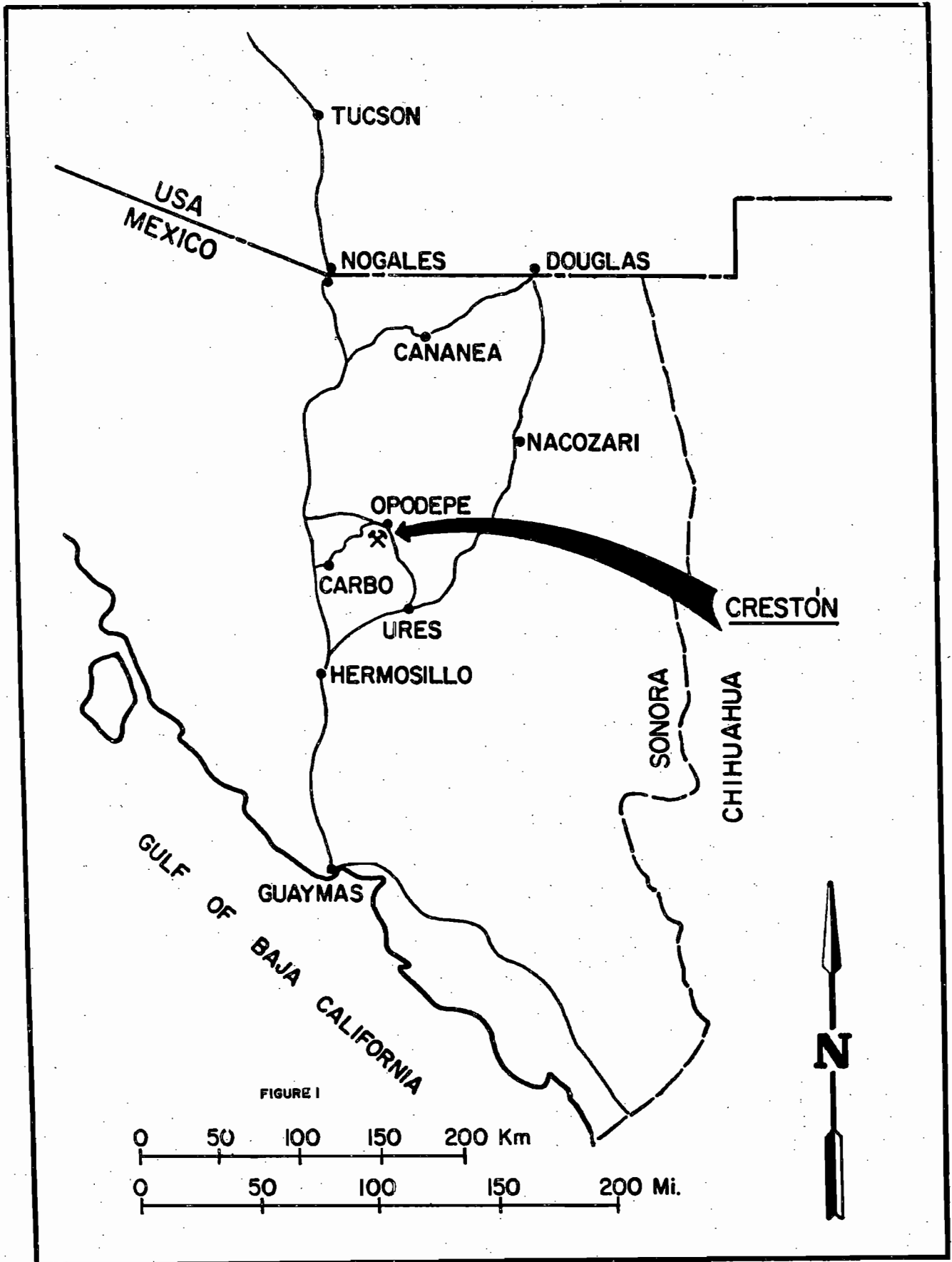
REGIONAL GEOLOGY SETTING

Regionally, the Creston molybdenum-copper deposit lies within a large roof pendant resting on the Sonoran batholith. The roof pendant complex is dominantly composed of pre-Cambrian to Paleozoic (?) metamorphic rocks. The underlying batholithic rocks are of Tertiary age and correspond to an extensive igneous unit which ranges from granodiorite to quartz monzonite in composition; absolute age determinations of this unit in the Creston region yield an average age of 55 m.y. by the K/Ar method. Emplacement of this Tertiary batholith caused extensive fracturing of the roof pendant units; this event is interpreted as being responsible for the rock preparation mechanism that preceded mineralization in the Creston area. Although conclusive evidence of a younger intrusive responsible for the molybdenum mineralization at Creston has not been encountered, it is highly conceivable that such an intrusive does exist in the prospect area.

Structurally, this region is within the extension of the Basin and Range Province of the Southwestern United States. This structural province has been recognized to prevail in most of the State of Sonora.

GEOLOGY OF THE CRESTON DEPOSIT

The geologic features of the Creston deposit are complex. The rock types exposed on the surface are affected by the strong hydrothermal alteration and postmineral structural events, and supergene



effects which made some of the rock types virtually unrecognizable. This problem is especially critical in the main mineralized zone where lithologic contact relationships are difficult to map with certainty.

Lithology

Basically, the rock types at Creston can be divided into Precambrian and Paleozoic (?) metamorphics; and Paleozoic (?) and Tertiary intrusives, dikes and breccias (Figure 2). Recent conglomerates, talus and landslides are common in the area.

Metamorphic rocks

The oldest rocks in the area are the phyllites and metavolcanics, which are treated as one unit (pma); gneisses of various textural variations (gn); and the Creston granite (Cg) which intruded these rocks prior to metamorphism.

Of these metamorphic rocks, the Creston granite (a gneissic granite) is important because of its great affinity to alteration and mineralization; it hosts most of the molybdenum mineralization in the Creston deposit as presently known. The other metamorphic rocks, especially the phyllites and meta-andesites, exhibit strong resistance to pervasive alteration and mineralization, these effects being present exclusively in fractures.

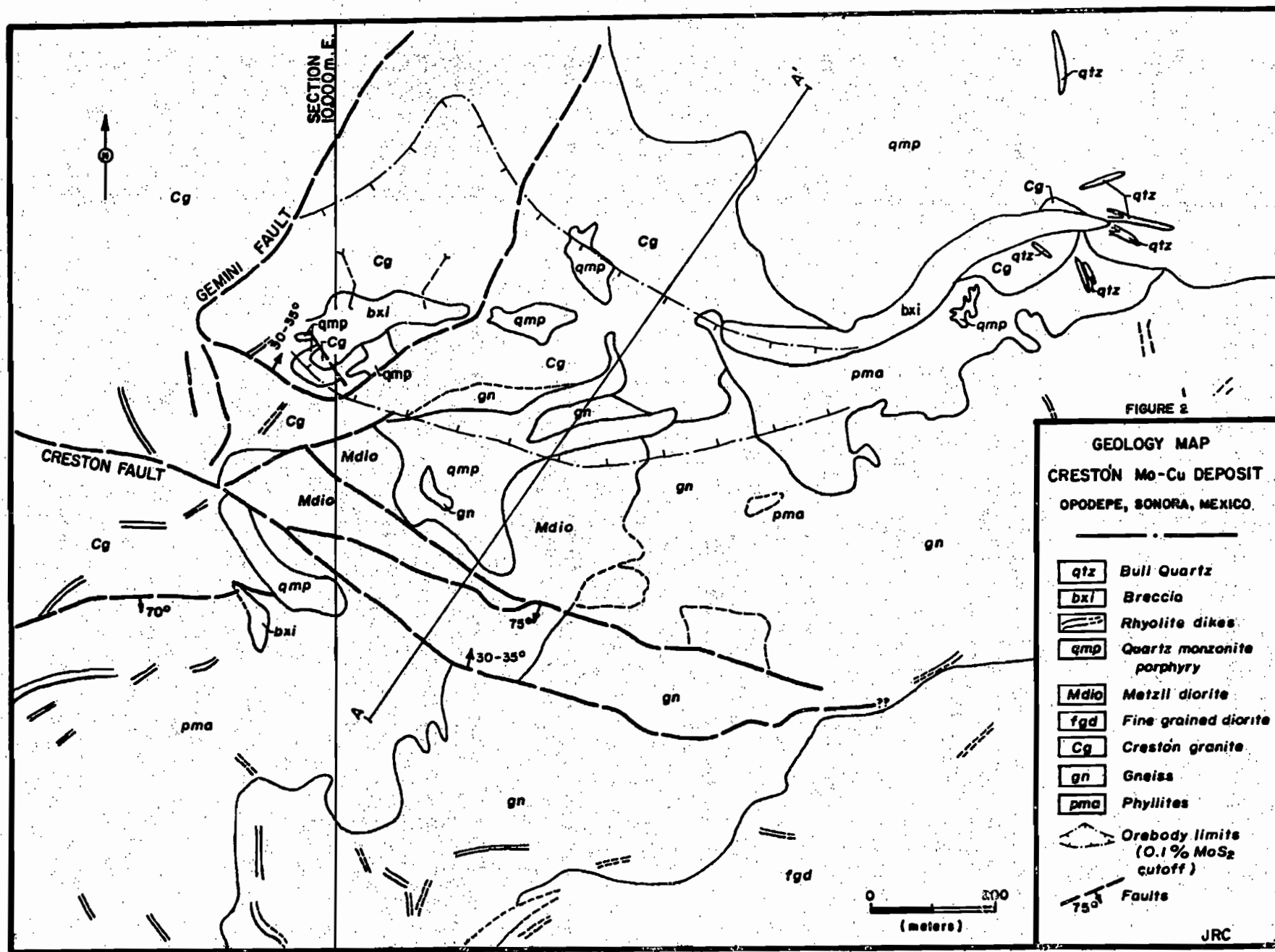
Intrusive rocks

Three post-metamorphism intrusive rocks have been mapped in the prospect area. These are the Meztli diorite, and a fine-grained diorite of Paleozoic (?) age, and the porphyritic quartz monzonite of Tertiary age. A number of acidic to intermediate dikes cross-cut the prospect area.

The Meztli diorite (Mdio) is an elongated, tabular body occurring exclusively along the southern flanks of the Creston hill. Megascopic observations show it to be a medium-grained, porphyritic diorite. Economically, this unit is adversely important in that it does not contain significant molybdenum mineralization; however, it hosts a good percentage of the secondary copper mineralization at Creston.

The fine-grained diorite (fgd) occurs over a large portion of the southeast of the main Creston area and as dikes at depth in the central area. It is a dark greenish-white, fine- to locally medium-grained equigranular rock with a dioritic composition. Except for a few locally oxidized shear zones, this unit is barren.

The porphyritic quartz monzonite (qmp) is exposed immediately northeast and about four kilometers west-southwest of the Creston area. It is also found at depth below the mineralized zone and as smaller apophysis around the Creston hill. The porphyritic quartz



monzonite was dated at 55 m.y. by the K/Ar method.

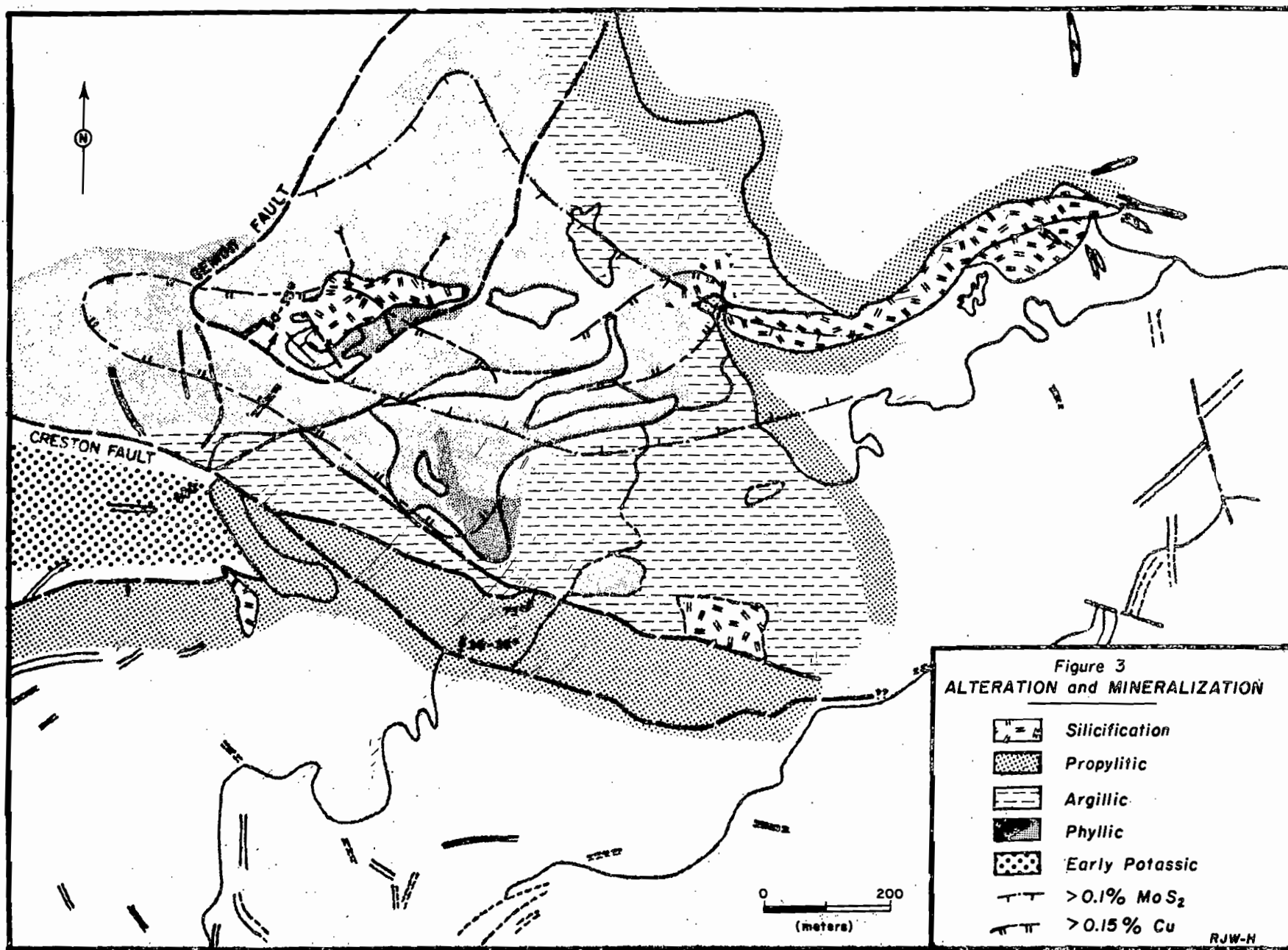
The emplacement of the porphyritic quartz monzonite was very closely followed by the intrusion of a variety of acid to intermediate dikes. These dikes are nearly barren.

Breccias

A number of bodies of breccia occur in the Creston area. Classification of these different types of breccia, without touching upon genetic implications, is rather difficult. Thus, for purposes of this discussion, they have been divided into types "c" and "i". These breccia types are described below.

The "c" breccias (bxc) are characterized by the presence of tabular, angular fragments of phyllites, meta-andesites, and, occasionally, Creston granite, exhibiting no rotation, and set in a quartz-rich matrix in amounts exceeding 10% by volume. The matrix-volume factor was a criterion to discriminate between a "c" breccia as defined here and those areas with strong brecciation but no quartz-rich matrix. Surface mapping and drilling results show that this type of "dry" brecciation is indeed very extensive south of the main Creston deposit. This is not surprising, however, since the brecciation, strong fracturing and local mild dislocation were caused by the emplacement of the porphyritic quartz monzonite. A very large area, therefore, has been structurally affected by this intrusive event, creating an area of intense "dry" brecciation (i.e., no matrix). Later, silica-rich solutions were selectively introduced, and it is in those zones where the "c" breccias can be readily recognized. The amount of matrix material in the "c" breccias ranges from 3% to 40%, but the areas classified as "c" breccias per se have an excess of 10%, with local areas with 20% to 30% matrix material. At depth in the southwestern portions of the area, the "c" breccia matrix consists of quartz-epidote-chlorite-pyrite and local molybdenite. In the central area, at depth in the ore zone, the "c" breccia matrix is made up of quartz-sericite-molybdenite-pyrite (Figure 5).

The type "i" breccias (bxi) occur in two localities in the Creston area: the main Creston breccia which is a large elongated body to the east, and a small occurrence to the southwest. These breccias are characterized by rotated, subangular to rounded fragments of Creston granite, porphyritic quartz monzonite, and rhyolites set in a vuggy quartz or quartz-sericite matrix with minor rock flour. Unlike the "c" type breccias, the characteristics of the "i" breccias suggest a dynamic origin in which late-stage silica-rich solutions probably travelled along large fissures and chimneys (?), and in some cases probably vented through the surface. Volume of matrix material in these breccias are in the 5% to 90% range, but average about 30%. The "i" breccias are either intramineral, or slightly postmineral in age.



Bull quartz veins

These massive veins of white, barren quartz are common in the area. They represent the latest event in the mineralizing history of the area.

Recent conglomerates and talus

Scattered occurrences of conglomerates cemented by iron and manganese oxides were mapped in the area. These are generally two or three meters thick. They are poorly sorted and contain angular to sub-rounded fragments of most of the rock types present in the area.

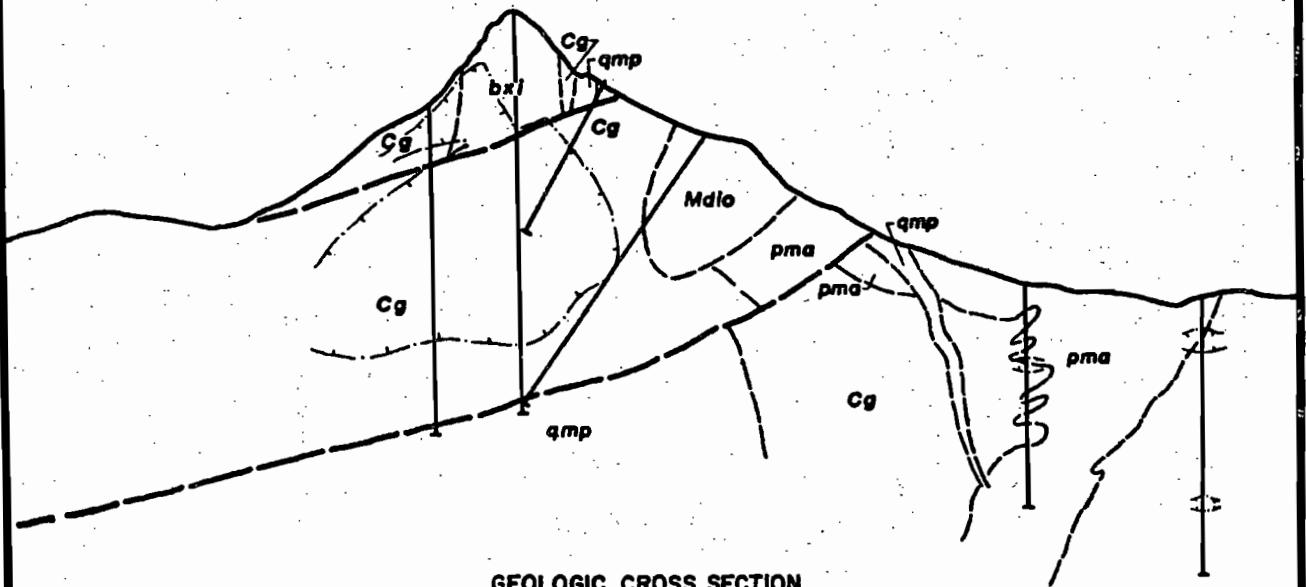
Several bodies of recent talus occur on the property. A large mass in the central area is composed mainly of fragments of porphyritic quartz monzonite and Creston granite ranging from pebbles to large boulders in size in a loose matrix of crushed rock.

Structure

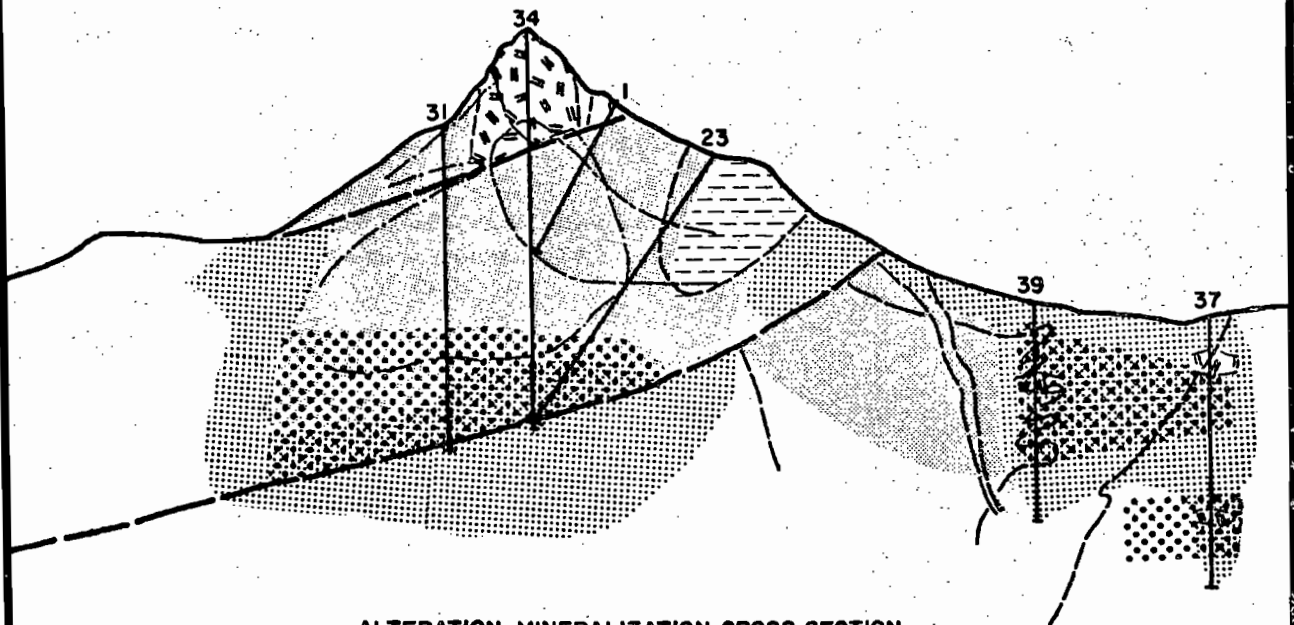
Undoubtedly the Creston area has undergone intense intermittent tectonic activity since the Precambrian, with the most important and relevant episodes having taken place from early Tertiary to Recent times. These latter events include intrusion of the porphyritic quartz monzonite, which caused intense fracturing and faulting in the intruded rocks, followed by the hydrothermal alteration and mineralization stages and the postmineral faulting.

The most important structures which affect the economic aspects of the deposit are a series of low-angle, postmineral normal faults of which the Creston fault is the most significant one (Figures 2, 4, 5). Dipping from 30° to 35° to the northeast, the Creston fault sharply separates intensely altered and mineralized units in its hanging-wall from weakly altered to fresh, barren units in its foot-wall. In drill holes, the Creston fault zone consists of a thick sequence exhibiting strong shearing, brecciation, and deformation; it ranges in thickness from 5 to 70 meters. Normally, these structural effects are present in the hanging wall, but sometimes they are evident in the underlying porphyritic quartz monzonite. The fault breccias consist of small, rounded to subrounded pebbles of different rock types in a matrix of crushed rock and gouge. These features suggest that movement along the Creston fault has been of considerable proportions. In fact, based on the approximate displacement of the steep contact of the porphyritic quartz monzonite in the upper block with a similar contact projected in the lower block (Figure 5), the amount of dip-slip displacement can be as much as 450 meters. On the surface, the Creston fault zone is expressed as wide zones of strong shattering, changes in rock type and alteration, fault breccias, and sudden changes in topographic slopes. The abundance of colluvial material and vegetation makes it difficult to physically pinpoint its trace on the surface.

SECTION 10,000 E. LOOKING EAST



GEOLOGIC CROSS SECTION



ALTERATION MINERALIZATION CROSS SECTION

FIGURE 4

0 200
(meters)

SECTION A-A' LOOKING NORTHWEST

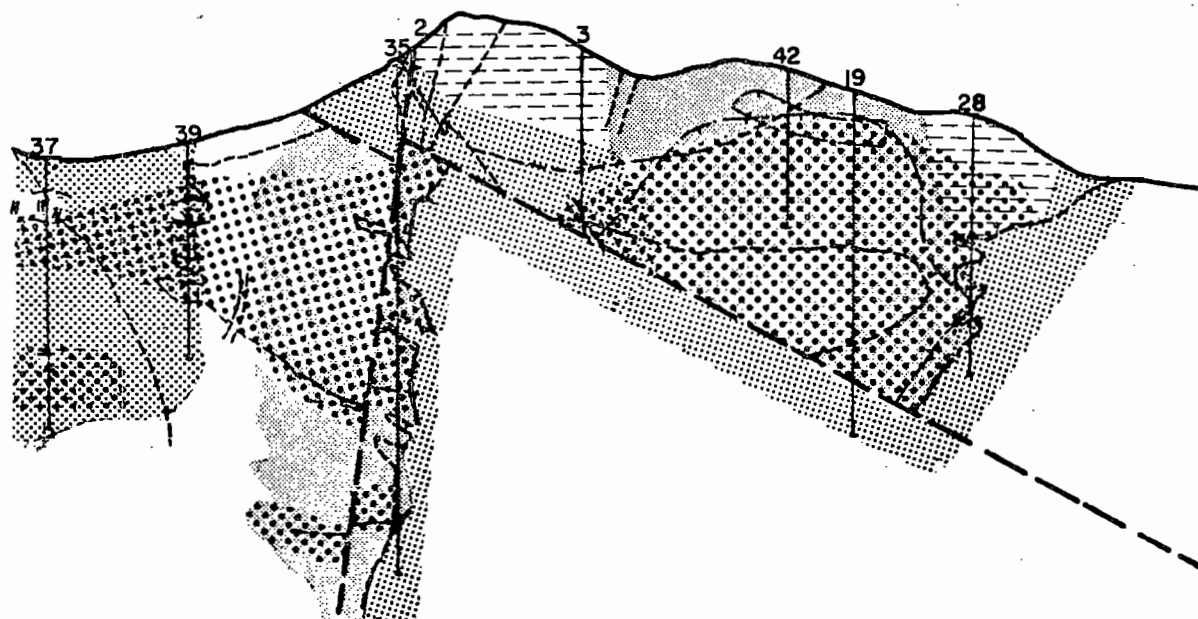
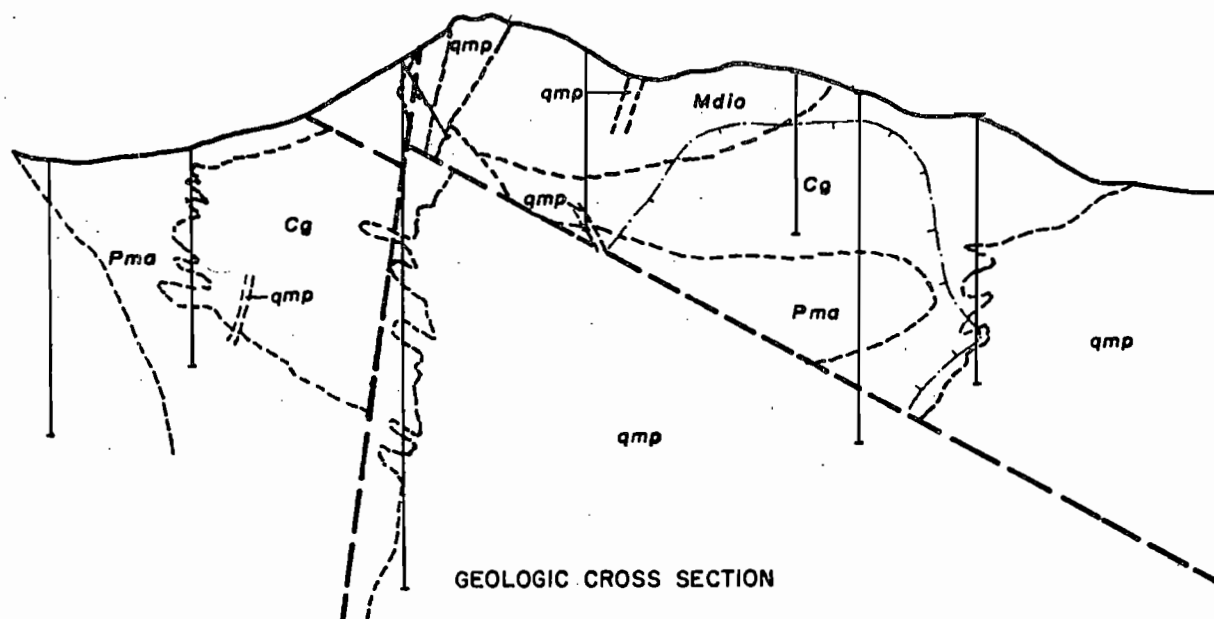


FIG. 5

0 200
(meters)

Parallel to the Creston fault and located in its hanging wall, the Gemini fault is important because its upper block contains the uppermost portion of the Creston breccia pipe (Figure 4). This fault has been intercepted in six drill holes; it strikes about N60W and dips about 30° , slightly flatter than the Creston fault. Dip-slip movement along the Gemini fault is estimated to be at least 200 meters. The main reason for this estimate is the absence of the Creston "i" breccia in the foot-wall of the Gemini fault as checked by several drill holes. The surface trace of the Gemini fault has been plotted from cross-sections (Figure 2). It coincides fairly well with a NE-trending gulley to the north and behaves rather erratically to the east of the Creston hill.

Alteration

Excluding the supergene effects, at least two distinct alteration events are evident at Creston: an early potassium-rich event, and a later, silica-rich, molybdenum-bearing hydrothermal event. The nature of the distribution of the alteration mineral assemblages logged in drill holes is erratic in detail; they do not appear to be distributed in distinct zonal arrangements. The post-mineral structural history of the area makes interpretation of the alteration patterns even more difficult. The important fact is that molybdenum mineralization is associated with the phyllic alteration assemblage often accompanied by various degrees of silicification. Refer to Figure 3.

Potassic alteration

This alteration event is characterized by strong potassium metasomatism which was probably brought about by a late magmatic event associated with the emplacement of the porphyritic quartz monzonite. Its characteristics are strong, pervasive K-spar flooding and veining, and formation of secondary fine-grained biotite either from pre-existing biotite or from recrystallized orthoclase feldspars. Spatially, this type of alteration is widespread; it occurs at depth in the central mineralized area near the contact with the porphyritic quartz monzonite, and on the surface, in the area immediately north of the "c" type breccia to the southwest.

The mineral assemblages of this type of alteration have been affected by the later hydrothermal events where the secondary biotite is commonly either chloritized or sericitized. No significant molybdenum mineralization is directly associated with this potassic event.

Hydrothermal alteration

This later event is closely associated with the molybdenum mineralization at Creston. The mineral assemblages associated with this alteration phase may be divided into propylitic, argillic, and phyllic, with zones of silicification being most common in the phyllic zone. Their areal distribution suggests that some of these

alteration assemblages are closely related to rock types (Figure 3). The Creston granite is the only unit that exhibits all of these types of alteration assemblages.

In the propylitic zone the alteration is manifested by chloritization of biotite, minor epidote (usually in the phyllites and meta-andesites), calcite, and fine-grained pyrite associated with chlorite. In the central mineralized area, this type of alteration affects the secondary biotite in the Creston granite. In centrally located drill holes, the chloritization of secondary biotite is generally found at depths below 250 meters and beyond significant molybdenum mineralization. It may also be stressed that the potassic alteration described above coexists with, and is overprinted by the propylitic alteration described here. This phenomenon was one of the factors that led to the interpretation of an earlier, prehydrothermal potassic event in the area.

The argillic alteration zone at Creston is characterized by pervasive clay (kaolin) alteration of feldspar and absence of ferromagnesian minerals. On the surface this alteration type is greatly enhanced by supergene effects. Generally, the Creston granite exhibits weak argillization, except where strongly sheared. The bulk of the argillic alteration present on the surface affects the Meztlí diorite. Drill holes in the southern portion of the area cut Meztlí diorite exhibiting strong, pervasive clay alteration of the feldspars with minor fracture-controlled quartz-sericite-pyrite zones. No significant molybdenum mineralization is associated with this alteration type.

A major zone exhibiting phyllic alteration occurs in the Creston area; it is centered around the Creston breccia and covers the mineralized body as known to date. The phyllic assemblage consists essentially of pervasive sericitization of feldspars and biotite. The intensity of the alteration ranges from slight sericite "dusting" of plagioclase to development of zones with massive, fine- to medium-grained sericite. The sericitization is most strong in and around the Creston breccia; in the breccia itself, the breccia fragments are strongly sericitized.

At least two separate silicification stages can be discerned at Creston. The first stage, which consists of white, clear quartz, was intimately associated with the emplacement of a major portion of the molybdenum mineralization in the area. The second stage is a barren, or nearly barren, white clear to milky quartz. The first-stage quartz is generally mineralized with molybdenite, pyrite and, rarely, chalcopyrite, and occurs as rather steep, cross-cutting veinlets commonly ranging from 0.1-5.0 cm in thickness, and as matrix to the type "c" breccia found at depth in the east-central portion of the deposit. Occasionally, veins of this nature can be as thick as 30 cm. The second-stage quartz is generally barren; typical examples would be the matrix of the central portion of the Creston breccia (bxi), and the veins of bull quartz. In the case of the Creston breccia (bxi), the second-stage quartz might have been emplaced using the same major conduits that were

used for the emplacement of the molybdenum mineralization, dispersing in the process any high-grade mineralization that could have been deposited in the Creston pipe. Both silica-rich stages seem to have caused silicification of the intruded wall rocks in just about the same manner. In some localities, silicification zones are completely devoid of sulfide mineralization, whereas in other localities the best mineralization is in zones with strongly silicified sections, usually accompanied by phyllic-type alteration.

Mineralization

Mineralization of economic interest at Creston consists chiefly of molybdenite with a small separate zone of secondary copper mineralization located on the southern flanks of the Creston hill. Accessory metallic minerals such as galena, sphalerite, and chalcopyrite were observed, but these are minor occurrences of no economic importance. Tungsten, silver and gold are very low within the Cerro Creston deposit. The average pyrite content is 3%; it occurs in quartz veinlets and disseminated. Locally, short premineral fault zones contain as much as 50% total sulfides, with pyrite being by far the dominant sulfide.

The depth to the bottom of the oxidized zone in the mineralized area ranges from near surface to 73 meters deep, but averages about 60 meters. Minerals in the oxide zone include ilsemanite, ferrimolybdenite, chalcantite, brochantite, and turquoise. Visible oxidized molybdenum minerals in the oxide zone are fairly sparse.

Molybdenum

Most of the molybdenum mineralization at Creston occurs as fine- to coarse-grained (0.1 to 3 mm.) molybdenum crystals set in steeply dipping quartz veinlets ranging from 0.5 to 5 cm. in thickness. Occasionally molybdenite-bearing quartz veins as thick as 30 cm. have been observed. The main host rock is the Creston granite. Occurrences of this type make up the bulk (+70%) of the Creston stockwork molybdenum deposit as presently known.

A fraction (+20%) of the molybdenum mineralization is in the quartz matrix of the type "c" breccia (Figure 5). Here the molybdenite mineralization occurs as fine-grained flakes (0.05 to 1 mm.) at the very contact of the quartz matrix with the breccia fragments. Occasionally high-grade blebs are found within the type "c" breccia; molybdenum mineralization of this type, while coarse-grained, seems to be highly erratic.

Still minor (5-10%) molybdenite occur as coarsely disseminated rosettes in short zones (less than 30 cm.) of massive coarse-grained sericite. These zones, however, are few and far between. The molybdenite rosettes range from 0.10 to 3 mm. in size. This type of molybdenite mineralization occurs in the mineralized Creston granite and, where present, it is usually high-grade.

Copper

Secondary copper mineralization was intercepted in fairly significant amounts on the southern flank of the Creston hill. It occurs as thin films of chalcocite coating fine-grained disseminated pyrite and pyrite in quartz veinlets. The occurrence of secondary copper is widespread in the district, but these are generally thin and of low grade; three holes drilled more than two kilometers west of the Creston area intercepted this type of sparse mineralization.

In the Creston area, the secondary copper mineralization appears to be mainly associated with the Meztli diorite, and to a certain degree with the more mafic rocks encountered in drill holes. Weak chalcocite mineralization is present in zones as thick as 50 meters, with short, localized intercepts with grades better than 0.15% copper.

Hypogene copper mineralization is very low (0.001 to 0.1% copper on the average). Chalcopyrite, while being the main copper mineral in the sulfide zone, is not abundant; it occurs as small discrete blebs in quartz veinlets, often accompanied by pyrite and, rarely, by sphalerite and molybdenite.

Evidence of oxidized primary copper minerals in the oxide zone suggests a low-grade, vein-controlled copper mineralization system, much like that encountered in the sulfide zone. It appears that a considerable column of oxidized and leached rock was eroded away as enrichment progressed. The present oxidized column is too thin to warrant the formation of the considerable thickness of enriched copper blanket encountered.

Alteration-Mineralization Associations

The various alteration assemblages described, including the earlier potassic event, have extensively affected the rocks in the area, especially the Creston granite. Furthermore, the Creston granite is the main host to the bulk of the molybdenum mineralization in the area. These two main factors influenced the selection of the Creston granite as a frame of reference for purposes of discussing the alteration-mineralization associations at Creston. Other rock types either are not affected by pervasive alteration (e.g. the phyllites and meta-andesites), or are affected by only one alteration assemblage (e.g., the Meztli diorite which is mainly argillized).

The early potassic event, as described elsewhere in this report, is not associated with significant metallic mineralization. In the Creston granite the effects of this type of alteration are widespread, but molybdenum mineralization associated with veins and with irregular, fracture-controlled segregations of orthoclase have only rarely been observed. The molybdenum occurrences of this type are not of economic grade.

The main hydrothermal alteration type associated with molybdenum of economic grade is the phyllic phase. Here the molybdenite mineralization occurs in quartz stockwork and disseminated in short, erratic zones of strongly sericitized Creston granite. In zones where the Creston granite is moderately pervasively sericitized, the molybdenite is generally found in quartz veinlets with minor dissemination in narrow zones adjacent to the quartz-molybdenite veins. Pyrite is ubiquitous (0.5 to 3%), and other sulfide minerals such as chalcopyrite and sphalerite are relatively minor. Thin coatings of chalcocite on pyrite are common in the enriched zone.

Mineralization in argillically and propylitically altered Creston granite are restricted to thin, scattered quartz-molybdenite-pyrite veinlets; but more often to quartz-pyrite veinlets. In general, no significant hypogene metallic mineralization of economic importance occurs in these alteration zones.

THE PROPOSED CRESTON MODEL

As presently known, the Creston molybdenum-copper deposit appears to be the uppermost portion of a much larger mineralized system which was segmented by at least two, major, en echelon, low-angle gravity faults of postmineral age. The original system was very large and it was emplaced near the margins of the quartz monzonite porphyry, which is a phase of the Tertiary batholith found elsewhere in Sonora. As portrayed on cross section A-A¹ (Figure 5), the known deposit is in the hanging wall of the Creston fault. Limited deep drilling done in the footwall of this fault has intercepted significant molybdenum mineralization which is interpreted as being a portion of a deeper segment of the original system.

The mineralization system in the Creston area can be classified as a porphyry molybdenum-copper system. Some of the factors that support this classification are:

- a. the style of alteration and mineralization,
- b. the relatively low-grade and large tonnage of hypogene mineralization,
- c. the presence of a supergene copper enrichment blanket,
- d. the existence of at least one well-defined breccia pipe, and
- e. although not of great importance, the Laramide age association of the deposit.

As mentioned elsewhere in this paper, an intrusive rock responsible for the mineralization at Creston has not been recognized as yet. As geologic work continues to be done as part of the evaluation of

the project, new information will be obtained; this, in turn, will provide the basis for a thorough understanding of the mineralization system and the genesis of the Creston area.

FLL-CPM, January 1981.

HYDROGEOLOGY OF THE COSTA DE HERMOSILLO REGION

(Roadlog not available at press time)

FIELD TRIP No. 8 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 28

TRIP LEADERS:

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OTE. COL. CHEPEVERA, MONTERREY, N.L., MEXICO

G. CELAYA, S.A.R.H., DR. NORIEGA E IGNACIO RAMIREZ, HERMOSILLO,
SONORA, MEXICO.

J. L. JARDINES, S.A.R.H., PASEO DE LA REFORMA 35, PISO 11 MEXICO
1, D.F. MEXICO.

NEOTECTONICS AND PLIO-QUATERNARY VOLCANISM
OF THE AREA OF SANTA ROSALIA, BAJA CALIFORNIA

FIELD TRIP No. 9 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 28-30

TRIP LEADERS:

- B. COLLETTA, INSTITUTO DE GEOLOGÍA, UNAM., AP. POSTAL 1039,
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ROADLOG FOR THE SANTA ROSALIA FIELD TRIP

DAY ONE (March 28)

Time	Increment in Kilometers

08:00 h.	Depart from Hotel Valle Grande, Hermosillo. Head south to Guaymas on the Highway 15.
00.0	Cerro de la Campana with TV and microwave station relay: Paleozoic metamorphic limestones.
6 to 46.5	Mesozoic granodiorite east of the Highway. To W Mid-Tertiary ignimbrite where is a shrine to the "Virgen de Guadalupe" west of the Highway.
54	La Pintada - Mid Tertiary ignimbrites, dacites and rhyolites of the Sierra Libre.
65.5	Estación de Microondas: Late - Mesozoic (?) altered and fractured andesitic rocks. Along the road to the "Estación microondas", volcanic series dipping eastward; dikes and sills of rhyolite and dacite intruding pyroclastites and ignimbrites (Cochemé, 1981).
70 to 90	Thick dacitic flows.
93	Cerro Huerfano: andesitic series overlapped by Tertiary basaltic flows.
122	Fork to San Carlos: stay on Highway 15, to Guaymas.
124.5	Turn off to the Estación de Microondas and to the airstrip of Guaymas. Thick Tertiary ignimbritic series. Guaymas city, built in a cauldron structure. Rocks consists of a calc-alkaline suite including basalts, trachyandesites, rhyodacites and ignimbrites (Johnpeer 1977).
11:30	Take off from Guaymas airport, Servicios Aéreos S. A. flight.
12:00	View of Isla Tortuga, the only tholeiitic basalt volcano of the Gulf of California, (Batiza, 1977, 1978). Easternmost part of the island is downthrown along a NE-SW main fault. This volcanic structure lies NW of the Guaymas Basin, a spreading center recently studied through DSDP Leg 64 (Curry and Moore, 1979; Moore and Curry, 1979), and with a submersible (Lonsdale and Lawver, 1980). Dolerite sills and dikes intrude thick young sedimentary sequences.

12:15 h.

General view of the coast north of Santa Rosalía. From N to S: Sierra de Santa Ana (Miocene Comondú Group), the central block of Cerro La Reforma encircled by the rim of the La Reforma caldera (10 km diameter), the gently sloping southern flank of the caldera, and the Santa Rosalía Basin; in the background the high cone of the Tres Virgenes.

- At Cabo Virgenes, spectacular sequence of staircase marine terraces cutting Pliocene volcanics. The banks of Arroyo Las Palmitas and the Las Minutas locality show dioritic intrusive bodies (dated 1.37 ± 0.5 m.y., Schmidt and others, 1977).
- In the Arroyo La Reforma: Pliocene grey ashes and prismatic lava flows (30 m thick) with pillows.
- Yellow non-welded tuffs related with ignimbritic eruptions are conspicuous on the SW flank of Cerro La Reforma.
- The outer part of the rim is capped with Early (?) Pleistocene scoriaceous basalt flows.
- Between the La Reforma Caldera and the Tres Virgenes volcano, wide plateau composed of ignimbritic tuffs covered with alluvium.
- The two peaks N of the 3 Virgenes cone consists of dacitic andesitic flows. The 3 Virgenes Volcano results from viscous dacitic flows.

Recent volcanic activity (Ives, 1962) is evidenced by several dark basaltic flows. Light-colored pumices and ashes can be seen all around the Tres Virgenes Volcano.

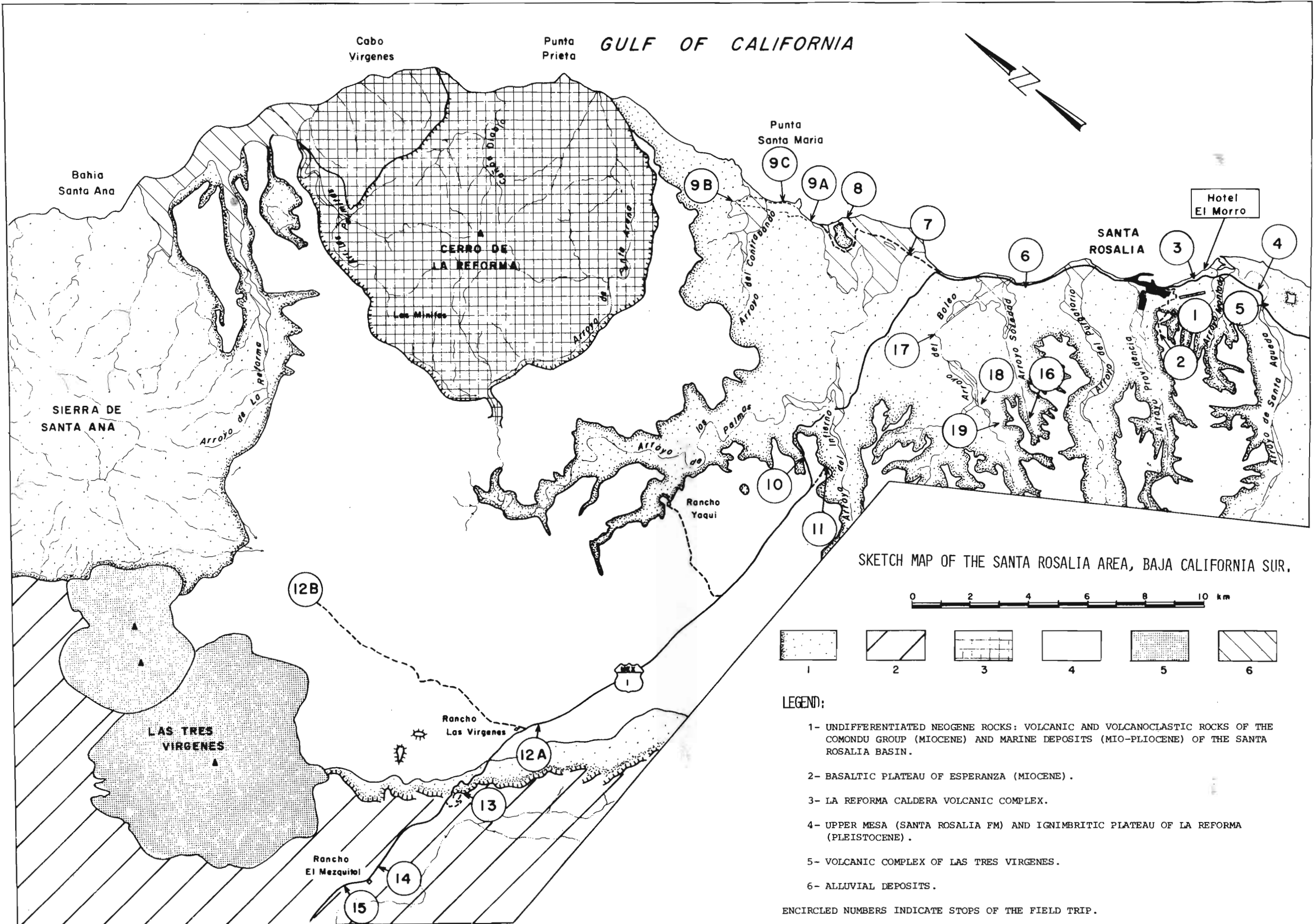
The main cone is built on the edge of a high escarpment cut in the Esperanza basaltic plateau and volcanoclastic andesites (Miocene Comondú Group).

- In the Santa Rosalía Basin: late Neogene marine embayment uplifted during the Quaternary and limited to the W by the Sierra Santa Lucía (Comondú Group).

Dissected upper mesa (Sta. Rosalía Formation and old abrasion marine terrace).

Mio (?) - Pliocene marine deposits exposed in the arroyo cuts.

Peaks of Comondú rocks rise through the Late Neogene



sedimentary series (paleoreliefs), and along the coast S of town.

Series of marine terraces particularly well-developed between Arroyos del Purgatorio and Santa Agueda.

13:00 h.

Landing at Sta. Rosalía airstrip (+ 100 m marine terrace).

Lunch.

00.0 Km.

Aboard vehicles.

Depart toward the upper mesa. Take dirt road at the northern extremity of airstrip. Small outcrops of Infierno Fm in roadcut.

0.6

Leave right road to cemetery, and turn off for cobble road uphill.

Overlook on Santa Rosalía, in left hair-pin bend. Coarse talus deposits and conglomeratic facies of Infierno Fm.

STOP 1

1.4

Top of the grade. Immediately before reaching the upper mesa, roadcut showing fossiliferous littoral deposit assigned to the Santa Rosalía Fm. Highest marine terrace of the area (elevation + 190 m): see section J in Ortlieb, this volume.

The road heads W. Small scarp before reaching stop 2.

STOP 2

2.0

Estación de microondas built in a small graben limited by two NW-SE normal faults. Obvious vertical offset of about 3 m. The northeastern fault appears as a calchified gap of about 0.50 m wide. This represents the most recent faulting activity in the Basin of Santa Rosalía.

View of the Arroyo Providencia. On other side of the canyon: pink tuffaceous Boleo Fm, yellowish sandy and conglomeratic Gloria Fm, greyish sandy and conglomeratic Infierno Fm.

Turn back

4.0

Airstrip.

If times permit, visit of + 132 m marine terrace above southern extremity of airstrip, and of + 86 m marine terrace (with very fossiliferous deposit) below the soccer field, E of the airstrip. (See section J, Ortlieb, this volume).

Descent to Highway 1.

5.0 Km

Right turn on Highway 1, towards S.

Roadcuts in Boleo Fm and Pleistocene marine deposits.

5.4

Boleo Fm overlying Comondú Group. Numerous faults on E roadcut; flat abrasion marine surface covered with well- rounded cobbles.

STOP 3

6.2

Dirt road to E. Pleistocene marine surface (+ 25 m) with successive ridges. Lower terrace at + 10 m, in sea-cliff.

Return to Highway.

6.5

Hotel El Morro. Pleistocene fossiliferous conglomeratic marine deposit overlying Comondú Group in road cut in front of the hotel.

7.1

Descent to Arroyo Montado.

To the W view of tilted blocks of Boleo Fm and Gloria Fm on the eastern compartment of the major Santa Agueda Fault (See Plate 1, Colletta and Angelier, this volume).

7.6

Arroyo Santa Agueda. The southeastern bank at the mouth of Arroyo Santa Agueda presents a conspicuous abrasion terrace (+ 10 m) cut in volcanoclastic rocks of the Comondú Group.

STOP 4

10.0

Brown calcareous crust, rich in Fe and Mn oxides, from the base of the Boleo Fm. This unit locally fossiliferous, which unconformably overlies the Comondú Group, has been interpreted as the first marine deposit of the Santa Rosalía Basin.

At the top of the southern bank of Arroyo Santa Agueda, low angle disconformity between Infierno and Gloria Fm. The upper surface corresponds to the Santa Rosalía Fm (morphostratigraphic unit) and is offset by a NW-SE normal fault.

Highway 1 crosses Arroyo Santa Agueda.

STOP 5

10.5

Exposures of the base of Boleo Fm.

- Talus breccia on Comondú rocks.
- Calcareous crust
- Gypsum beds overlain by tuffaceous sandstones.

The Santa Agueda fault can be observed in the northern bank of the arroyo: contact between Infierno Fm and Comondú Group (vertical displacement probably superior to 200 m).

If times permits climb up hill to reach the remnant of the upper mesa. Panorama to the N : 5 well-defined marine terraces (see sections J and K, Ortlieb, this volume). Note the general flat surface cut by the sea during Santa Rosalía time. In the background: the three peaks of the Las Tres Vírgenes and associated cones; to the N: the La Reforma Caldera.

The upper mesa (Sta. Rosalía Fm ?) is cut by a 4 m high scarp; this scarp is the superficial evidence of a WNW-ESE normal fault that cuts underlying sandy deposits of the Gloria Fm. Slickensides observed in Gloria Fm indicate pure dip slip movement.

Turn back toward Sta. Rosalía.

16.2 Km

Santa Rosalía airport turnoff.

Gloria Fm in roadcuts.

17.0

Km 0 of Highway 1, at the intersection with Santa Rosalía main street (old mine train monument).

17.3

Smelter of the copper company (Compañía Minera de Santa Rosalía).

17.9

Refuse dumps of the earlier "Compagnie du Boleo". Along the shore small ore trains.

19.6

Marine and fluvial terrace on Gloria Fm.

19.7

Cross the Arroyo Purgatorio.

STOP 6

21.8

Loma del Tirabuzon (Corkscrew Hill) fossiliferous locality in Gloria Fm : shark fauna (see Applegate and Espinoza, this volume).

Turn back to Hotel El Morro.

28.9

Hotel El Morro.

DAY TWO (March 29)

Highway 1 towards N.

0.0 Km

Km 0, Santa Rosalía.

- 5.3 Km Cross Arroyo del Boleo. Outcrops of Gloria Fm.
- 7.6 Leave Highway 1 and bear right just before the grade at sign: "No deje piedras en el Pavimento" (aviso a los Geólogos). Cliffs of greenish silty Gloria Fm capped with Pleistocene marine deposits, along the dirt road.
- STOP 7 9.2 Arroyo Infierno. Abundant tension gashes filled with gypsum crystals in Gloria Fm. Statistical microtectonic analysis reveals two directions of extension: E-W and WNW-ESE (see: Colletta and Angelier, this volume).
- 10.3 To the NW, view of two Pleistocene basaltic volcanoes.
- 10.5 Fork, take dirt road heading N.
- STOP 8 11.3 Reach coastline. General view of La Reforma. Succession of 6 marine terraces, in staircase disposition, at Punta Santa María (Section I, see: Ortlieb, this volume).
- 12.1 Turn back to fork.
- 13.6 Facing sandy Gloria Fm overlapped by Pleistocene basaltic flows. Turn seaward, to the NE.
- STOP 9 A 14.8 Hike to the Arroyo del Contrabando (don't forget to bring water).
- Approx.
walking time:
- 0:00 Follow trail along the beach toward N.
- Climb up hill just before reaching Punta Santa María cliffs.
- 0:15 Top of grade. Calichified rounded cobbles of the + 40 m marine terrace. Take left at fork.
- 0:20 Descent in canyon; Gloria Fm overlapped by Pliocene volcanic flows.
- 0:30 Climbing up on the other bank of canyon. Top of the grade (in situ marine fossils, + 40 m terrace).
- 0:45 General view of the northern cliffs in the Arroyo Contrabando; descent in a small arroyo.
- 1:00 h Arroyo del Contrabando. Head N, and cross the arroyo bed.

STOP 9 B

1:30 h Base of main cliff. Pliocene volcanic sequence overlying Gloria Fm: pumicitic tuffs, sills, lapillites and lava flows. Marine shells associated to pumicitic tuffs.

1:45 Follow arroyo to its mouth.

2:00 Reach the beach and follow it to the south.

STOP 9 C

2:15 Paleo-valley in Gloria Fm filled with pumicitic tuffs, and volcanoclastics. Arroyo-cutting in the latter deposits.

2:30 Exposures of basaltic flows in the sea-cliff.

2:40 Climb up steep grade and reach the lowest marine terrace of Punta Santa María.

3:00 Back to the vehicles. Lunch.

20.4 Km

Back to Highway 1: head W, up the grade. Roadcuts in Gloria Fm to the SE; in the background observe a paleorelief of Comondú rocks.

22.3

View of the two volcanic cones on the SE edge of the La Reforma volcanic complex.

23.8

Massive gypsum of the Boleo Fm.

24.9

Pink tuff overlying gypsum layers (Boleo Fm).

26.1

Descent in Arroyo del Infierno. Yellowish sandy Gloria Fm. Well-exposed subvertical fault plane in roadcut to the left.

26.9

Arroyo del Infierno

27.5

Numerous gypsum tension gashes in microconglomerates and greyish sands of the Gloria Fm. They indicate an E-W extension.

STOP 10

29.6

"Cuesta del Infierno". Ignimbritic flow of the La Reforma Caldera, covering the Pleistocene Santa Rosalía Fm and the Pliocene Infierno and Gloria Fm.

The complete volcanic sequence consists in (from bottom to top):

- Thick pumicitic layers and light coloured and thinly layered ashes that result of plinian-type eruption,

- A scoriaceous basaltic bed,
- Ignimbritic flows (20 m thick).

A close association of basalts and acid products appears in the magmatism of the Reforma caldera.

The Santa Rosalía marine Fm is conglomeratic and fossiliferous. This is the highest outcrop of Pleistocene marine deposits of the Sta. Rosalía Basin. It is attributed to Early Pleistocene (Wilson, 1948) and maybe to the late Early Pleistocene, around 1 M.y. (Ortlieb, this volume).

In the Gloria Fm two families of tensions gashes were found: the older one shows E-W extension, and the other corresponds to an WNW-ESE extension, probably related to the present extensional regime in the Gulf of California.

30.01 Km Top of grade. Estación de Microondas.

30.8 In major right bend of Highway, turn off to the E. Take the very bad old transpeninsular road.

STOP 11

31.7 Panorama of NW part of Santa Rosalía Basin and southern flank of La Reforma.

To the N, ignimbritic sheet overlying the conglomeratic Sta. Rosalía and Infierno Fm. Below, sandy Gloria Fm. and conglomeratic Boleo Fm. Observe normal fault in the latter formations. In the background the Cerro de la Reforma, and outer slope of the caldera covered with Pleistocene basalts.

To the NE, confluent of Arroyos Las Palmas and Infierno, and in the background, brownish massive beds of gypsum (Boleo Fm).

To the E, by clear weather, Isla Tortuga is visible.

To the SE, ignimbrites capping the Santa Rosalía Fm (upper mesa).

32.6 Back to Highway 1.

Flat surface of the ignimbritic plateau, with isolated peaks of Comondú basement rocks.

STOP 12 A

45.0 View of the SE flank of the las Tres Virgenes volcano (see: Fig. 3 Sawlan, this volume).

- 45.9 Km Rancho las Tres Virgenes. If time permits turn off to the ranch; follow dirt road for about 12 Km, heading N. After second cattle shack, arroyo cut (stop 12B) showing Pleistocene basalt overlain by ignimbritic flows from La Reforma complex and las Tres Virgenes older sequence.
- 47.0 Highway begins to climb up major scarp. Thick volcanoclastic and andesitic sequence of the Sierra Sta. Lucía Fm. (Comondú Group).
- 49.2 Several faults in roadcuts. Association of pure normal faults, oblique faults and tension gashes that indicate ENE - WSW extension.
- 50.2 Top of grade.
- 50.4 Turn off to the "Estación de Microondas".
- STOP 13 51.6 Estación de Microondas. Panorama of La Reforma and las Tres Virgenes (See geological map, Demant this volume).
- 52.8 Back to Highway 1.
- 54.1 Pleistocene pumice in roadcuts, and on the slopes of reliefs.
- STOP 14 56.2 Most recent basaltic lava flows, 200 m N of Highway.
- STOP 15 57.8 Highway crosses basaltic lava flows.
- "This lava contains plagioclase microlites, small olivines, clino-pyroxenes, and numerous magnetite grains...(it) has no obvious relationship with the Tres Virgenes dacitic magma chamber, but represents a small batch of basic magma rising from a deeper zone" (Demant, this volume).
- Way back to Santa Rosalía.
- 100.3 Back to Motel El Morro.
- DAY THREE (March 30)
- Highway 1 to the north.
- 00.0 Km 0, Santa Rosalía.

- 5.3 Km Leave Highway 1 to take dirt road entering the arroyos Boleo and La Soledad.
- 5.5 Fork: bear left to go upstream in the arroyo La Soledad. Arroyo cut in Gloria Fm; deposits dipping about 5° to the NNE.
- 6.0 To the north, on the left, peak El Morrito; yellowish sands capped by brown calcarenites, all attributed to Gloria Fm. Nannofauna collected at the base and on the flanks of the peak indicates an Early Pliocene age (NN12 Zone).
- 6.4 Pink clayey tuff of the Boleo Fm overlain by yellowish strata of Gloria Fm.
- 6.8 Dark brown ore beds rich in Mn and Cu. Mineralization is of hydrothermal origin and occurred during a NE-SW to E-W extensive stage contemporaneous with Boleo deposits. Normal faults led hydrothermal solutions to the surface.
- 8.4 Conglomerates of the Boleo Fm.
- 9.1 Normal faults cutting Boleo and Gloria Fm. Small outcrop on the left showing distinct density of faulting in Boleo Fm and in Gloria Fm.
- 9.7 Contact Gloria Fm / Boleo Fm : yellowish conglomerate and sandstones of the Gloria Fm lying over pink clayey tuff of the Boleo Fm in slight unconformity.
- 10.0 Boulders of Gloria Fm coquina can be observed along the road. On the south side of the road residues of ore beds with copper mineralization.
- 10.2 Mine gallery.
- Leave 2 dirt roads on the left and follow the road that crosses the arroyo and climb on the northern bank.
- STOP 16 10.5 Association of reverse faults and strike slip faults induced by a N-S compressive stress. Strike slip faults strike 155 to 170 and are mainly dextral. The strike slip faulting stage occurred between Boleo and Gloria deposition.

Observe the pink-red volcanoclastic layer that constitutes a key horizon called "Cinta Colorada" (Wilson, 1948).

Turn and go back to the mouth of the arroyo.

15.5 Km Back to the fork at the arroyo Boleo entrance. Turn left to go upstream in the arroyo Boleo.

17.00 Gypsum of the Boleo Fm.

17.7 Gypsum quarry on the left.

STOP 17 18.6 Huge crystal of gypsum in a "synclinal faulted zone". Gypsum strata are gently bent: each anticline is separated by a recrystallized fault zone without true syncline.

Directions of folds are parallel to the main faults of the region and are probably more related to underlying faulted blocks than to a compressive event.

18.7 Mouth of the "Cañada de la Gloria".

19.7 Opening of the gypsum cañon; valley get wider in the clayey tuff of the Boleo Fm.

On the southern bank, the brown ore beds are slightly offset by small reverse faults according to a N-S compressive stress. This minor reverse faulting is probably related to strike slip events that occurred at the limit between Boleo and Gloria time.

21.2 In the background to the south, angular disconformity between Boleo and Gloria Fm.

21.8 Fork: bear left and cross a small cañoncut in volcanic rocks of the Comondú Group.

STOP 18 22.2 Paleorelief of Comondú rocks covered by mineralized fossiliferous limestone and tuffs of the Boleo Fm.

On the slope of the structure, conjugated normal faults in Boleo deposits have been tilted showing that the present dip of the stratas does not only result of depositional processes and that the pre-Boleo paleotopography was somewhat milder.

22.3 Leave dirt road on the left.

22.4 Leave other dirt road on the left.

- STOP 19 22.7 Km Multistriated fault plane in Boleo and Comondú Fm. Leave vehicles and climb up the thalweg to the south. At about 5 mn hiking, reach obvious slickenside showing two sets of striation (a fainter third set can locally be observed) that indicate two successive movements on the same fault plane. The older (syn-Boleo) slickensides indicate a dip slip displacement while the younger (Post-Boleo) ones indicate oblique right slip with important lateral component.
- View of the antiform- paleorelief to the NE.
- 37.2 Turn and go back to the Santa Rosalía airstrip.
- 13:00 h Lunch.
- 14:30 Take off. Servicios Aéreos flight.
- 15:30 Guaymas airport.
- 18:00 Arrive Valle Grande Hotel in Hermosillo.

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THE GEOLOGY AND SELACHIAN PALEONTOLOGY OF LOMA DEL TIRABUZON (CORKSCREW HILL) ,
SANTA ROSALIA, B.C.S.

Shelton P. Applegate

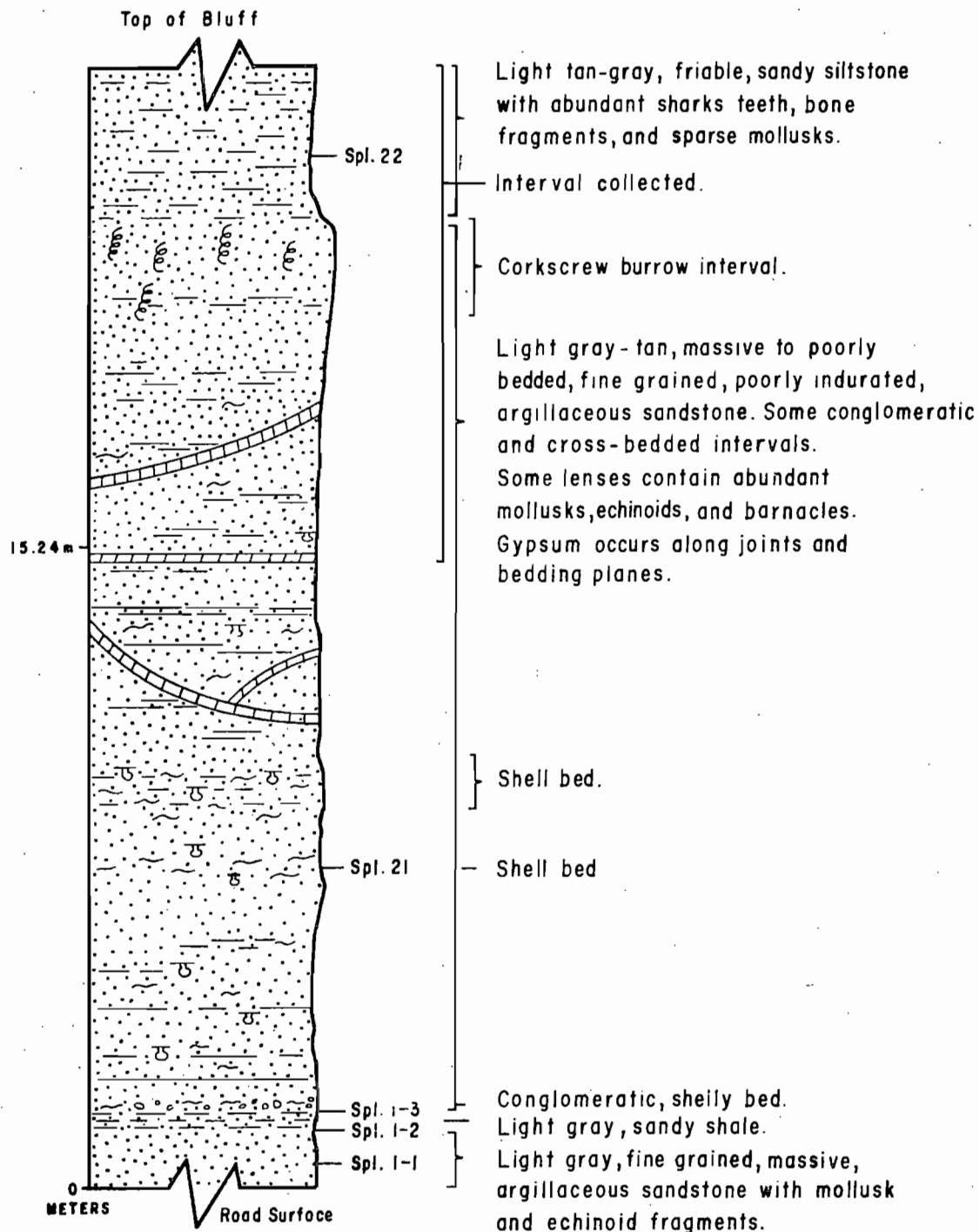
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The rich fossiliferous beds that exist at Loma del Tirabuzon have been discussed by Applegate, (1978). The locality is 5 km north of the town of Santa Rosalia. On the east the locality is bounded by sea cliffs that are on the west side of the road. The area covers about 1.5 hectares, where approximately 50 meters of section are exposed. The lower 30 meters form the bluffs at the side of the road, (Figure 1). The top 3 meters of these cliffs contain shark's teeth. Below the shark's tooth layer occurs the corkscrew-like burrows which give the locality its name. These burrows reach a length of one or two meters. Below the ichnofossils are shell beds and layers rich in echinoderm spines. The base of the formation is not exposed at this locality. The sediments consist of greyish tan silts and sands with interfingering conglomerates.

The lower beds including the shark's tooth layer are part of Wilson's, Gloria Formation, (1955). Since the name Gloria was preoccupied, we suggest that the formation be called the Cañada de la Gloria Formation, because the original name Gloria was taken from the Canyon known as the Cañada de la Gloria.

The fossil fauna consists of Foraminifera, ostracods, bryozoans, gastropods, pelecypods, echinoderm plates and spines, crab claws, as well as the corkscrew-like burrows, bony



fish teeth, head plates, vertebrae and spines, shark and ray teeth, vertebrae and dermal denticles, turtle bones and scutes, whale and dolphin ear bones and teeth and other mammal bone which is poorly preserved.

Only the shark fauna and one new species of gastropod has been described from this locality. A revised and expanded list of the sharks is as follows:

1. Hexanchus cf. griseus;
Common off of the California Coast but not known from the Gulf of California.
2. Notorhynchus sp.; The teeth are close to the living Notorhynchus cepedianus recently reported from the Gulf of California (Applegate 1979).
3. Heterodontus sp.; 2 or 3 living species exist in the Gulf of California.
4. Squalus sp. Known from the west side of Baja but not from the Gulf of California.
5. Odontaspis ferox; Known from southern Baja near La Paz.
6. Eugomphodus acutissima; Extinct, a related species living in the Western Pacific and North Atlantic.
7. Isurus benedeni; Extinct, known from the middle Oligocene to the lower Pliocene of Europe.
8. Isurus oxyrinchus; Living in the Gulf of California.
9. Carcharodon megalodon; extinct.
10. Carcharodon sulcidens; extinct.
11. Carcharodon carcharias; living in the Gulf of California.
12. Alopias pelagicus; Known from San Blas, Mexico Most likely living in the Gulf of California.
13. Cetorhinus sp.; Living Cetorhinus maximus known in the Gulf

of California today from a single record.

14. Mustelus sp.; Over 5 species living in the Gulf of California at our locality known by only two teeth.
15. Galeorhinus galeus?; Common on Pacific side of Baja rare in the Gulf of California.
16. Galeocerdo rosaliaensis; Extinct; probable ancestor of Galeocerdo cuvier which lives in the Gulf of California today.
17. Hemipristis serra; Extinct, a living related single species is confined to the Indian Ocean.
18. Rhizoprionodon longurio; Common at our locality and common in the Gulf of California today.
19. Prionace glauca; Rare as a fossil at our locality but living south of La Paz in the Cape area.
20. Carcharhinus albimarginatus?; Living in the Gulf of California, fossil teeth may represent a different species.
21. Carcharhinus altimus; Living in the Gulf of California.
22. Carcharhinus brachyurus; Living in the Gulf of California.
23. Carcharhinus falciformis; Living in the Gulf of California, a common fossil at our locality.
24. Carcharhinus galapagoensis; living in the Gulf of California.
25. Carcharhinus leucas; Living in the Gulf of California, a common fossil.
26. Carcharhinus limbatus?; Living in the Gulf of California, rare as fossil.
27. Carcharhinus obscurus; Living in the Gulf of California, a common fossil.
28. Carcharhinus porosus; Living in the Gulf of California,

a rare fossil at our locality.

29. Carcharhinus velox; Living in the Gulf of California.
30. Sphyrna lewini; Living in the Gulf of California today where it is very common.
31. Sphyrna media; Living in the Gulf of California.
32. Sphyrna mokarran; Living in the Gulf of California.
33. Sphyrna tiburo vespertina; Living in the Gulf of California.
34. Sphyrna zygaena; Living in the Gulf of California.

The lack of a recent species in a fossil fauna only represents negative evidence, yet it is of interest to analyze the recent pacific species which do not exist in this locality in relationship to their known habitats. The Loma del Tirabuzon lacks a number of deep water sharks known in the pacific waters off of Mexico, such as; Echinorhinus cookei, Alopias superciliosus, Apristurus brunneus, Apristurus kampae, Cephalurus cephalus, Galeus pipe-ratus, and Parmaturus xaniurus. In the Tirabuzon fauna we do have sharks known to live in fairly deep water today: Hexanchus cf. griseus, Notorhynchus sp. and Odontaspis ferox. All of these are known by only a few teeth. The open sea pelagic sharks that have not been found are: Carcharhinus maou, Alopias vulpinus and Rhincodon typus, on the other hand open ocean sharks such as: Prionace glauca, Carcharhinus falciformis, Sphyrna mokarran and Isurus oxyrinchus do occur. Two sharks that are found in extremely shallow water are also absent Negaprion fronto and Ginglymostoma cirratum. Rocky substrate forms such as: Triakis semifasciata, Cephaloscyllium ventriosum and one or two species of Heterodontus are not known. It is interesting that we have only two Mustelus teeth, a genus with 5 species living many of them are quite common in the Gulf of California. Perhaps the most surprising thing is the total lack of Squatina, a shark known from moderate depths.

From the above and the known distribution of living sharks that occur in our fauna we could expect a depth from 20 to 200 meters. The shallow depths being favored. The surface conditions must have been that of an open ocean, 2 or 3 km from the shore.

Various works have placed the age of the Cañada de la Gloria Formation, from Lower Pliocene to Upper Pliocene. Applegate, (1978), suggested a Middle Pliocene age. Recent work in progress at the Geological Institute on the Pliocene of Baja California has produced no evidence to change this determination.

Along the Pacific Coast, Lower Pliocene shark faunas are poorly known. The Pliocene of Simi Valley, California, USA, appears to be of this age. Here we find two species of Isurus. Isurus hastalis and I. plana. These species are common in Middle and Upper Miocene sediments, evidently these did not extend into the Middle Pliocene. North of San Felipe, Baja California Norte, in Pliocene beds above the Miocene diatomites we have one shark's tooth that seems to be intermediate between the Miocene Galeocerdo aduncus and the Middle and Upper Pliocene Galeocerdo rosaliaensis, which at least suggests the possibility of a Lower Pliocene age for these Imperial Formation beds.

The fauna at Rancho Algodones, Baja California Sur, is Upper Pliocene in age although sharks teeth are abundant the species appear limited. Isurus benedeni is absent as is Eugomphodus acutissima. All three species of Carcharodon are still present as is Hemipristis serra and Galeocerdo rosaliaensis.

Since the discovery of the fossils at Loma del Tirabuzon numerous amateur collectors have visited this area. Today it is almost impossible to picture the wealth of exposed teeth that

occurred there. We hope that this locality will ultimately become a national monument so at least the corkscrew burrows will be preserved.

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FAULTING EVOLUTION OF THE SANTA ROSALIA BASIN,
BAJA CALIFORNIA SUR, MEXICO

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RESUMEN

Se realizaron más de 600 medidas en fallas localizadas en la cuenca neógena de Sta. Rosalía. Estas medidas permitieron determinar la orientación de los esfuerzos principales $\sigma_1, \sigma_2, \sigma_3$. Los resultados muestran la existencia de un episodio de fallamiento lateral dextral el cual separa dos episodios de fallamiento normal extensivo. El primero de los episodios extensivos fué mucho mas intenso: produjo fallas de orientación general NNW-SSE que influenciaron las deformaciones posteriores y que resaltan actualmente como patrón principal. La existencia de desplazamientos laterales sugiere que la parte oriental de la Península de Baja California se comportó como una zona transformante extensional desde el Mioceno Superior-Plioceno Inferior.

INTRODUCTION

Over 600 faults planes and tension gashes were measured in the Neogene formations of the Santa Rosalía area. Measurements were performed at several distinct localities, each of them including 10 to 100 readings. For each faulted locality, the orientation of the 3 principal tectonic stresses. $\sigma_1, \sigma_2, \sigma_3$, were then computed with a statistical method analysis elaborated by Angelier (Angelier and Mechler, 1977; Angelier and Goguel, 1979; Angelier and Manoussis, 1980). The results are coherent and lead us to propose a tectonic history of the Santa Rosalía Basin. (These results are presented in a larger paper, by Angelier and others, in press).⁽¹⁾

GEOLOGIC SETTING

The oldest rocks of the area are quartz monzonites of Cretaceous age that crop out in the Arroyo de Las Palmas and in the central part of the Reforma caldera (Wilson, 1948; Schmidt 1975). This substratum is covered by thick volcanic and volcanoclastic series defined by Heim (1922) as the Comondu Formation. Overlying this formation disconformably are marine sediments, the deposition of which-

(1) Preliminary reports of this work were previously presented (Chorowicz and others 1980; Colletta and others 1980).

was periodically interrupted by tectonic movements and limited emersions. The sedimentary sequence was subdivided by Wilson (1948) into four formations, each of them slightly unconformable on the earlier one. The following sequence, can be observed (from bottom to top):

- the Boleo Formation: essentially conglomeratic with interbedded tuffaceous clayey sandstone and grading locally into thick gypsum facies (average thickness 100 m);
- the Gloria Formation: yellowish marine clayey sandstone and coarse conglomerates (average thickness 60 m);
- the Infierno Formation; conglomeratic and sandy with tuffaceous layers (average thickness 55 m);
- the Santa Rosalía Formation: marine and continental conglomerates (average thickness 10 m);

In the northern part of the basin, the whole sedimentary sequence is overlapped by ignimbritic flows from the Reforma complex (Demant and Ortlieb, 1979; Demant, this book).

Along the Gulf coast several marine terraces are well exposed (Ortlieb 1979; Demant and Ortlieb, 1979; Ortlieb, this book; Colletta and others, in prep.).

CHRONOSTRATIGRAPHY OF THE SANTA ROSALIA BASIN

Wilson (1948) proposed a chronostratigraphy based on marine faunal determinations. The three formations (Boleo, Gloria, Infierno) were thought to be, respectively, of Lower, Middle, and Upper Pliocene age, and an Early Pleistocene age was attributed to the Santa Rosalía Formation. Yet recent nannoplanktonic dating of the Gloria Formation is somewhat inconsistent with this chronology. A few samples collected in the lower and middle part of the Gloria sequence (BC 49, 50, 51, 52, 58, 59, see Plate 1 for location) carry nannofauna of Early Pliocene age probably the NN12 zone (determination C. Muller). Although this age is based on only a few samples and should be confirmed by further studies, we consider it reliable. Consequently we proposed a new chronostratigraphic interpretation:

- Comondu Formation (in the area of Santa Rosalía): no older than the Middle Miocene (?)
- Boleo Formation: Upper Miocene (?)
- Gloria Formation: Lower to Middle Pliocene
- Infierno Formation: Upper Pliocene (?)
- Santa Rosalía Formation: Pleistocene

Up to now, no direct evidence of the older age we suggest for the Boleo Formation has been found. Nevertheless we can argue that Miocene marine deposits are documented at various localities of the Gulf Province (Gómez, 1971; Andersen, 1969; Gastil and Krummenacher, 1977; Minch, 1979) and that, the very poor marine fauna reported by Wilson did not firmly date the Boleo Formation as Lower Pliocene.

As for the age of the Comondu Formation, K/Ar dates (Mc Fall, 1968; Gastil and others, 1975; Gastil and others, 1979) show that it could be bracketed between 21.6 and 6.6. My. In the San Ignacio area (about 70 km west of Santa Rosalía) the Comondu Formation grades in a marine facies (the San Ignacio Formation) that was dated by Gastil and others (1979) at 11 My K/Ar on tuffs.

Thus, the older ages we suggested for the Comondu and Boleo Formation appear quite consistent with the general chronological data in Central Baja California.

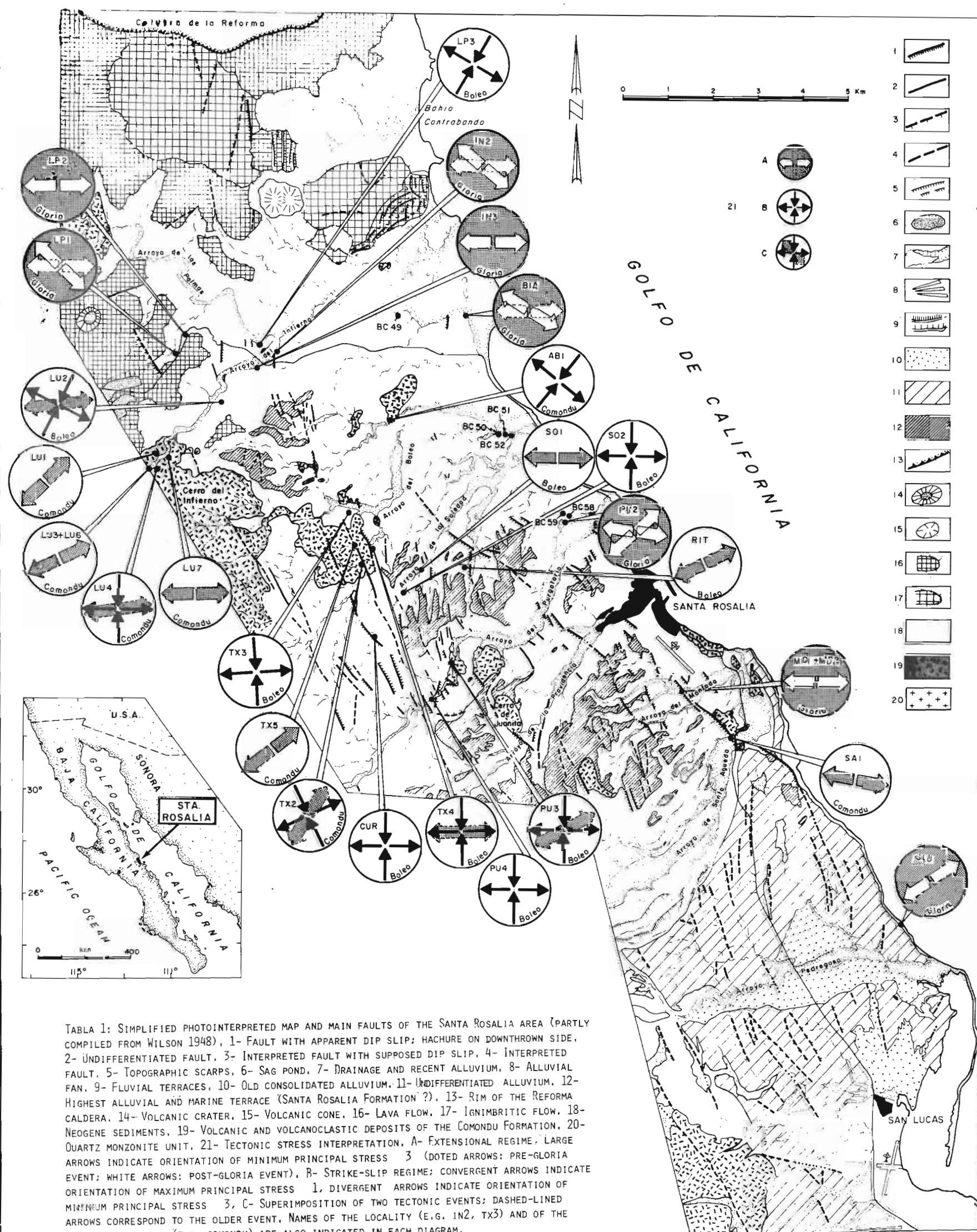


TABLE 1: SIMPLIFIED PHOTOINTERPRETED MAP AND MAIN FAULTS OF THE SANTA ROSALIA AREA (PARTLY COMPILED FROM WILSON 1948), 1- FAULT WITH APPARENT DIP SLIP; HACHURE ON DOWNTHROWN SIDE, 2- UNDIFFERENTIATED FAULT, 3- INTERPRETED FAULT WITH SUPPOSED DIP SLIP, 4- INTERPRETED FAULT, 5- TOPOGRAPHIC SCARPS, 6- SAG POND, 7- DRAINAGE AND RECENT ALLUVIUM, 8- ALLUVIAL FAN, 9- FLUVIAL TERRACES, 10- OLD CONSOLIDATED ALLUVIUM, 11- UNDIFFERENTIATED ALLUVIUM, 12- HIGHEST ALLUVIAL AND MARINE TERRACE (SANTA ROSALIA FORMATION?), 13- RIM OF THE REFORMA CALDERA, 14- VOLCANIC CRATER, 15- VOLCANIC CONE, 16- LAVA FLOW, 17- IGIMBRITIC FLOW, 18- NEOGENE SEDIMENTS, 19- VOLCANIC AND VOLCANOCLASTIC DEPOSITS OF THE COMONDU FORMATION, 20- QUARTZ MONZONITE UNIT, 21- TECTONIC STRESS INTERPRETATION, A- EXTENSIONAL REGIME, LARGE ARROWS INDICATE ORIENTATION OF MINIMUM PRINCIPAL STRESS 3 (DOTTED ARROWS: PRE-GLORIA EVENT; WHITE ARROWS: POST-GLORIA EVENT), B- STRIKE-SLIP REGIME; CONVERGENT ARROWS INDICATE ORIENTATION OF MAXIMUM PRINCIPAL STRESS 1, DIVERGENT ARROWS INDICATE ORIENTATION OF MINIMUM PRINCIPAL STRESS 3, C- SUPERIMPOSITION OF TWO TECTONIC EVENTS; DASHED-LINED ARROWS CORRESPOND TO THE OLDER EVENT, NAMES OF THE LOCALITY (E.G. IN2, TX3) AND OF THE FAULTED FORMATION (E.G. COMONDU) ARE ALSO INDICATED IN EACH DIAGRAM.

FAULTING PATTERN

As shown on Fig. 1 A, in the Santa Rosalía Basin, the faults have two predominant trends: one N-S (175) and the other NNW-SSE (160 to 155). Statistical analysis of all the measured brittle structures, including tension gashes, indicates a general N-S attitude and a lack of E-W trends; 75% of the strikes are included between 155 and 190. West-dipping fault planes are more numerous than east-dipping ones. This asymmetry which appears in Fig. 1 B, results from the high number of antithetical secondary faults induced by a few main gulfward syn-thetical faults.

About 90% of the faults planes have a dip over 45° (Fig. 2C) while pitches of the slickenlines are widely distributed between 0° to 90° (Fig. 1 D). This attitude corresponds to a general high angle faulting with all kind of movements from pure strike-slip to pure dip-slip.

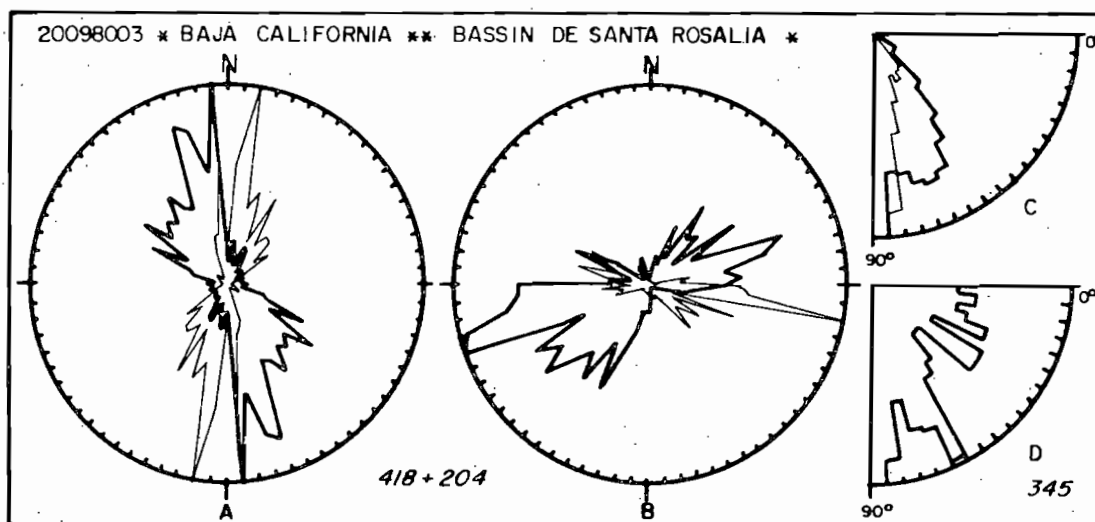


Fig. 1: Rose diagrams based on 418 fault planes (bold lines) 345 of which carry slickenslides, and on 204 tension gashes (thin lines) A- Strikes of planes. B- Dipping directions. C-Dips. D-Pitches of slickenslides.

Only very few true reverse faults, with pitch over 45° and reverse displacement, are observed (Fig. 2). Displacements are mainly right-normal-slips; in detail two populations of faults are predominant: (1) pure right-slip faults and (2) oblique-slip faults with pitches of about 60° (Fig. 2).

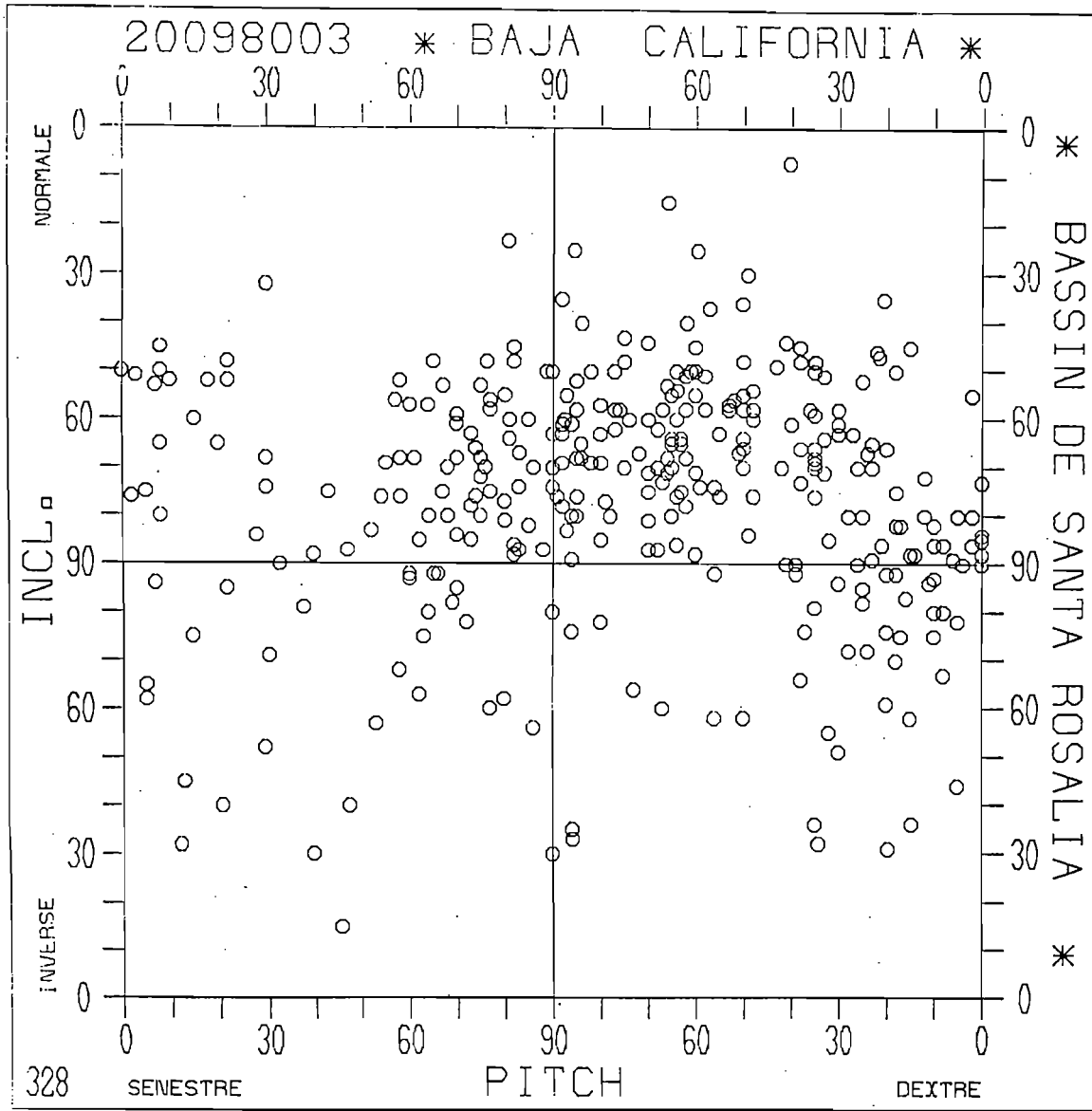


Fig. 2 : Plot of slickenslides pitches versus dips for 345 fault planes in the Santa Rosalia area. Observe the two predominant groups of pure strike-slips (pitch= 90°) and oblique slip faults with pitches of about 60°.

FAULTING EVENTS

The whole structural pattern is not the result of a single event. The faulting complexity increases with the age of the formations and indicates several superimposed tectonic stages. Three main successive faulting events are evident.

The extensive post-Comondu, pre Gloria event

The oldest tectonic regime registered in the Santa Rosalía Basin, is an extensional one: it postdate the Comondu volcanics, goes on through Boleo time and predates the Gloria time. A slight rotation of stresses might have occurred during this event and the σ_3 extensive principal stress does not appear uniformly oriented in the whole area; statistical study indicates a σ_3 ranging from NE-SW to E-W and the two major fault trends (NNW-SSE and NS) probably result of this extensive event. Although no major change in the tectonic regime is evidenced, two generations of faults were definitely recognized during this extensive event: one pre-Boleo and the other contemporaneous to the Boleo deposits.

The first generation (pre-Boleo) is well represented by numerous antithetical normal faults in the Comondu Formation particularly in the Arroyo del Infierno. Sets of conjugate normal faults are locally eastward tilted 10 to 30° (Fig.3). Both structures, faults and tilted strata were buried by the Boleo deposits. The tilting is thought to immediately postdate the faulting event and to predate the marine deposits. The paleotopography described in the Comondu volcanics by Wilson (1948) resulted probably from this first faulting stage.

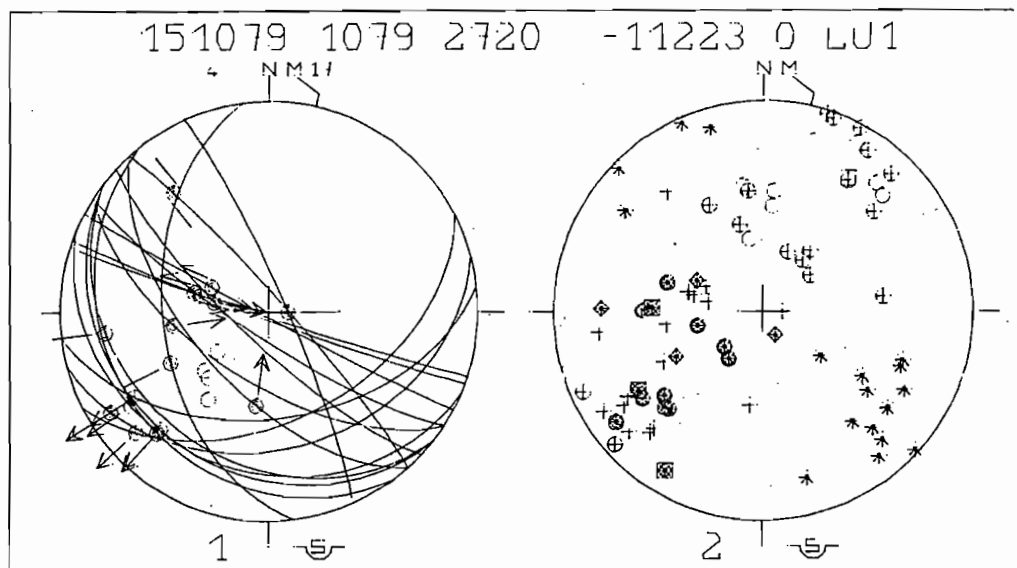


Fig 3: Example of tilted normal faults in the Arroyo del Infierno (Mina Lucifer). Schmidt net, lower hemisphere. A- Fault planes with slickenslides, small open circles are poles of strata. B-axes related to each fault plane (•, poles of fault planes; +, slickenslides; ⊙, ○, *, axes related to individual faults mechanisms). Observe that sub-vertical faults and reverse faults will get a west dipping normal attitude after rotating strata to an horizontal position.

The second generation (during the Boleo deposition) is more subdued but can be demonstrated by various features. In the Boleo Formation, normal faults due to a NNW-SSE extension are truncated by strike-slip and reverse faults of a younger tectonic event (LU 2). Several striated fault-planes show two generations of slickenslides; the first one indicates a normal faulting related to a NNW-SSE extension and the second one indicates strike-slip displacement belonging to the younger tectonic events (LU 2, TX 4, PU 3). Moreover, in the higher part of the Arroyo del Purgatorio and of the Arroyo del Boleo, several fault gaps which are filled with conglomerate indicate extensive tectonic movements contemporaneous with the Boleo deposits. Other evidences of these synsedimentary movements are a low angle unconformity and an intraformational conglomerate in the gypsum of the Arroyo del Infierno. Volcanic activity recorded by a layer of lapillis and cinders, called the Cinta Colorada, is doubtless related with these extensive tectonics. On the other hand, the ore deposits are thought to be of hydrothermal origin and also to be definitely related to the faulting (Wilson 1948). Ore beds are confined to the Boleo Formation and were offset by later tectonic events. Therefore the mineralization probably arose during this second faulting stage.

The post-Boleo-pre-Gloria strike-slip faulting event

At various localities, strike-slip movements are evidenced by horizontal - slickenslides. Right- lateral trends are more frequent but numerous left lateral faults were found, particularly in the Arroyo del Purgatorio. However, right or left trends are in agreement with a compressional principal stress (σ_1) roughly oriented N-S as deduced from the statistical study (Plate 1). A few small reverse faults related to this stress field can be observed in the Arroyo del Boleo. Fault planes of the first extensive stage were reactivated as shown by the superimposition of strike-slip slickenslides on earlier dip-slip slickenslides (PU 3, TX 4).

The amount of lateral offset cannot be precisely known but we think that no large strike-slip displacement occurred along any single major fault in the Santa Rosalía area. Yet, the general plot of lateral slip versus transversal (extensional) slip (Fig. 4) indicate a high number of lateral offsets. All the measurements showing strike-slip stress field (σ_1 and σ_3 horizontal) were performed in Comondu or Boleo Formation and no strike-slip system appears in the younger Gloria Formation. Therefore, we conclude that this tectonic event probably occurred at the boundary between the Boleo Formation and the Gloria Formation.

The post-Gloria extensive event

Gloria, Infierno and Santa Rosalía deposits are cut by important normal faults. Large faulted outcrops of the last two marine formations (Infierno and Santa Rosalía) are scarce and useful measurements could only be performed in the Gloria Formation. Post-Gloria brittle structures consist essentially of normal or oblique faults and gypsum tension gashes.

Veins of gypsum are sometimes very numerous and give good statistical results in computing the extensional stress σ_3 . Stress diagrams indicate two main extensional trends (BIA; LP2; IN2): One E-W and the other WNW-ESE. Detailed analyses of the attitude of gypsum crystals permits the establishment of the chronology of the two extensive stages, at least in the LPI locality.

First the E-W extension produced N-S veins with gypsum crystals perpendicular to the edges of the veins.

Then a WNW-ESE extension that reactivated older tension gashes, produced new oblique gypsum crystals and new NNE-SSW veins.

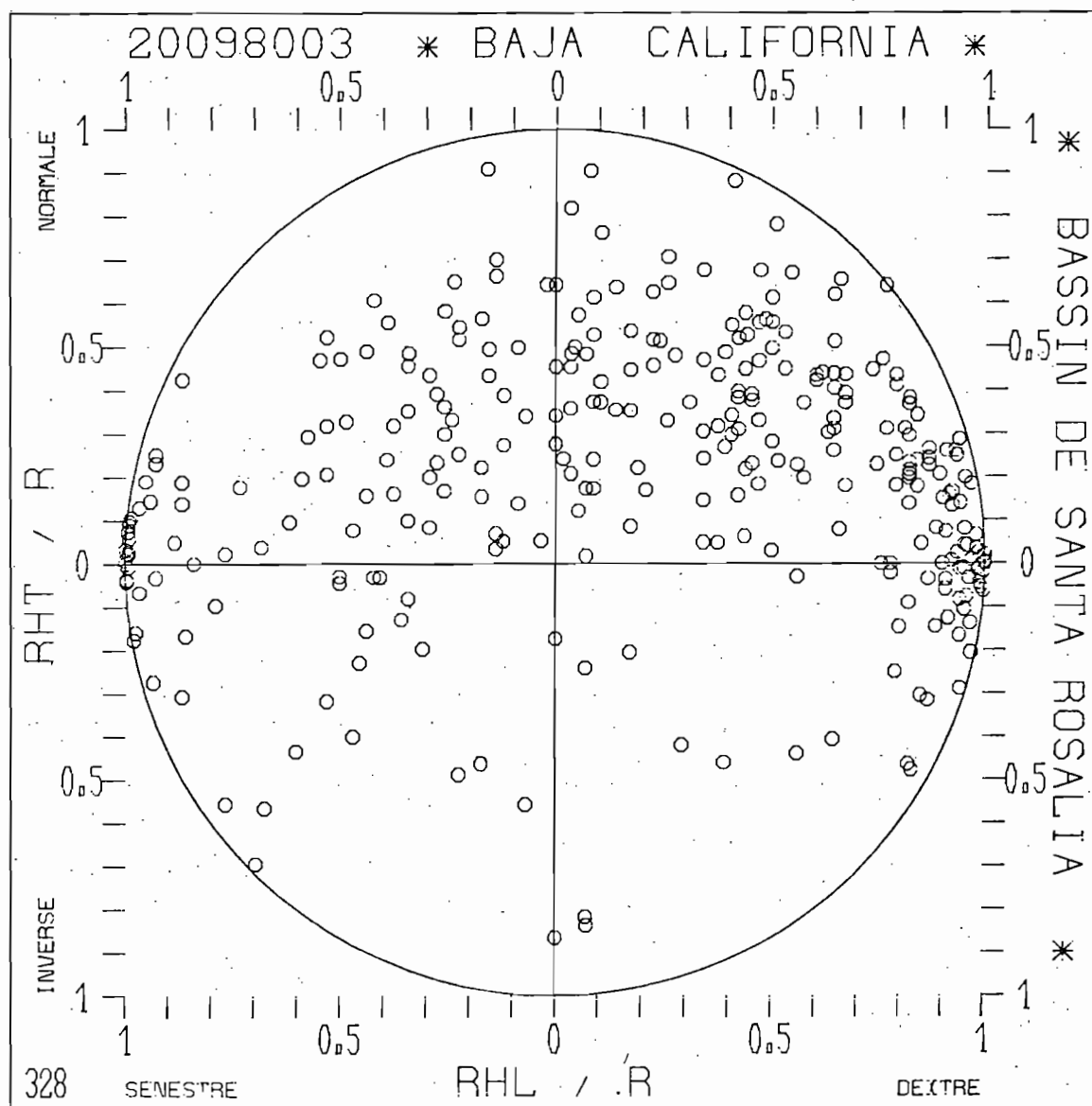


Fig. 4: Plot of the ratio RHL/R (RHL : horizontal -lateral component of the net slip, R : net slip) versus the ratio RHT/R (RHT : horizontal-transverse component of the net slip). Each quadrant of circle corresponds to reverse, normal, sinistral or dextral motion. Note the predominance of lateral movements, $RHT/R=0$ and $RHL/R=1$.

During these last two extensive events the earlier pre-Boleo NNW-SSE trends were reactivated and major faults of post-Gloria age were directly inherited from older zones of weakness. Therefore the general faulting trend appears unchanged. At the same time, the Gloria strata experienced an important eastward tilting in the southeastern part of the basin.

The timing of the two ultimate extensions could not be precisely established but it is obvious that tectonic movements went on during Pleistocene time.

Fault scarps which cut the conglomerates of the highest Pleistocene marine terrace (Santa Rosalía Formation ?) can be observed at the "Estación de microondas" of Santa Rosalía, and on the southern bank of the Arroyo de Santa Agueda, near its mouth.

HISTORY OF FAULTING

On the basis of the above mentioned chronostratigraphy, we can attempt to propose a timing of the faulting events. The tectonic history of the Santa Rosalía Basin may be subdivided in six successive stages:

- 1 . At the end of the Middle Miocene, about 10 My ago (?), occurred an important extension oriented NE-SW to EW associated to an eastward tilting. This first stage produced the major faults of the region that trend principally NNW-SSE.
- 2 . During the Upper Miocene, extensional processes went on concomitantly to the deposition of the Boleo Formation and to a volcanic and hydrothermal activity.
- 3 . At the boundary between Miocene and Pliocene, about 6 My ago, a main change in the tectonic regime appeared: strike slip movements occurred in response to a N-S compressional principal stress (σ_1). The earlier NNW-SSE normal faults were reactivated in right strike-slips.
- 4 . After Gloria time, at about 4 to 2 My (?), a new E-W extensional regime took place. Extension was milder than in pre-Pliocene events and only reactivated earlier faults while a first generation of gypsum veins occurred.
- 5 . More recently (not definitely dated), extensional stress changed to a WNW-ESE orientation. As a result of this change, a new set of tension gashes appeared and reactivation of the older structures went on.
- 6 . During the Quaternary, the Santa Rosalía area has experienced the most important uplift registered on the western coast of the Gulf. This exceptional - uplift is the result of two interrelated phenomena: (1) the regional rising of the Baja California block related to the opening of the Gulf and (2) the local resurgent doming related to the magmatic evolution of the Reforma caldera (Ortlieb, this book; Demant, this book).

CONCLUSIONS

In the Santa Rosalía area, detailed analysis of faulting demonstrates the existence, at about 6 My., of a strike-slip episode included in a dominant extensional regime. Other evidences of lateral movements during Pliocene and Pleistocene time were observed at several localities of the western Gulf coast (Mc Fall, 1968; Minch and James, 1979; and unpublished data). This strongly suggests that the Peninsula, in its eastern part at least, has worked as a complex transform extensional zone since Late Miocene-Early Pliocene time.

During later events (Upper Pliocene ?), the clockwise rotation of the extensional stress from E-W to WNW-ESE is probably related to the spreading of the main basins in the Gulf of California.

ACKNOWLEDGMENTS

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SEQUENCES OF PLEISTOCENE MARINE TERRACES
IN THE SANTA ROSALIA AREA, BAJA CALIFORNIA SUR, MEXICO

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ABSTRACT

The coastal area near Santa Rosalía and on the eastern flank of the La Reforma volcanic complex, in eastcentral Baja California, presents an exceptional sequence of Pleistocene marine terraces. These remnants of past high stands of sea-level have been studied along 11 traverses roughly perpendicular to the coastline. The shorelines interpreted to correspond to distinct interglacial sea-level are regularly distributed at elevations varying from about + 9 m to + 190 m above mean sea level. A number of nine main shorelines were identified. Since no radiometric dates are yet available, a tentative chronostratigraphy of these features has been inferred from the V28-238 deep-sea core paleoclimatic curve; according to these correlations nine "interglacial" isotopic stages would be represented: one in early Late Pleistocene, six in Middle Pleistocene, and two in late Early Pleistocene. The Santa Rosalía Formation previously attributed to the base of the Pleistocene is thought to be a late Early Pleistocene (about 1 M. y.) interglacial marine deposit. The vertical motions in the whole coastal area have probably been constant during the Quaternary although uplift rates are interpreted to have changed from a near value of 240 mm/ka to approximately 130 mm/ka, at about 350 000 years B.P.

RESUMEN

La costa cerca de Santa Rosalía y de la ladera oriental del complejo volcánico de La Reforma, en el noreste de Baja California Sur, presenta una secuencia excepcional de terrazas marinas pleistocénicas. Se han estudiado estas huellas de antiguos niveles del mar por medio de 11 secciones perpendiculares a la costa. Las antiguas líneas de costa, que son interpretadas como correspondientes a distintos períodos interglaciales, se encuentran en forma escalonada, estando la más baja a + 9 m aproximadamente, y la más elevada + 190 m sobre el nivel del mar.

Se identificaron en total nueve líneas de costas principales. Puesto que no existen todavía dataciones radiométricas de las terrazas, se intenta establecer una cronoestratigrafía de éstas a partir de la curva paleoclimática del núcleo V28-238; según estas correlaciones, nueve estadios isotópicos "interglaciales" serían representados: uno al principio del Pleistoceno Tardío, seis en el Pleistoceno Medio, y dos al final del Pleistoceno Temprano. La Formación Santa Rosalía, anteriormente atribuida a la base del Pleistoceno, bien podría ser un depósito marino interglacial de fines del Pleistoceno Temprano (cerca de 1 M.a). Los movimientos verticales en el conjunto de la zona costera han sido probablemente continuos durante el Cuaternario aunque parece que hace cerca de 350 000 años las tasas de levantamiento cambiaron de aproximadamente 240 mm/ 1000 años, a 130 mm/ 1000 años.

INTRODUCTION

The fluctuations of sea-level related to Pleistocene climatic variations have sometimes been registered and preserved from erosion as series of wave-cut benches, or of coral reefs, or as a succession of depositional features. In eastern Baja California the remnants of past high stands of sea level mainly consist of wave-cut platforms covered by thin littoral deposits. In Baja California peninsula, the region of Santa Rosalía (Fig.1) is known for its elevated Quaternary deposits (Wilson, 1948). The coast north and south of Santa Rosalía and the eastern flank of La Reforma volcanic complex prove to exhibit some of the most complete sets of staircase terraces in the Gulf of California area. These series of marine terraces which were only recently (1978-80) surveyed are briefly described here. Lateral and vertical correlations of the distinct shorelines are proposed in an attempt to decipher the amplitude and evolution of the Quaternary vertical movements of this region.

THE SANTA ROSALIA FORMATION

The Santa Rosalía Formation, described and named by Wilson (1948), is the most extended, and possibly the earliest, Quaternary marine unit in the region of Santa Rosalía. This formation crops out on top of the mesas to the W and to the NW of the town of Santa Rosalía. In the NW part of the Santa Rosalía Basin, the formation is overlapped by an ignimbritic sheet. These volcanics were formerly called "Tres Vírgenes volcanics" (Wilson, 1948; Wilson and Rocha Moreno, 1955; Schmidt, 1975), from the name of the Tres Vírgenes volcano (Fig.1) but are related to the volcanic history of La Reforma caldera (Demant and Ortlieb, 1979; Demant, this book). The Santa Rosalía Formation generally overlies unconformably the Infierno Formation. The latter formation was thought to be Late Pliocene (Wilson, 1948), but large scale regional correlations, particularly with the San Diego and Borrego Formations from southern California, suggest that it actually might be, at least partly, of Early Pleistocene age (Colletta and others, in preparation). Consequently the Santa Rosalía Formation would be late Early Pleistocene.

The Santa Rosalía Formation is a marine conglomerate with a sandy matrix. It contains a marine fauna indicating a nearshore or littoral environment. The fossil shells are notably better preserved than those of the underlying Pliocene (?) marine series. Lists of this fauna were published by Wilson (1948) and Ortlieb and Malpica (1978). The thickness of the fossiliferous deposits is commonly 1-3 meters. The marine beds are reported to grade laterally (?) and vertically to coarse terrestrial sediments (Wilson, 1948; Wilson and Rocha Moreno, 1955). In the present work, the Santa Rosalía Formation is referred to the marine facies only.

The lithology, the thickness and the faunal content of this sedimentary unit, as well as the location of its outcrops, indicate that the Santa Rosalía Formation is a result of a relatively short transgressive event, as are the Pleistocene interglacial stages. As a matter of fact, the Santa Rosalía Formation is very different, in many aspects, from the three underlying sedimentary formations (Boleo, Gloria and Infierno) of this basin; it would be perhaps better characterized as a morphostratigraphic unit rather than as a formation.

Except for two limited outcrops near the mouth of Arroyo Santa Agueda, Wilson (1948) did not identify the Santa Rosalía Formation along the present coastal area. Nevertheless fossiliferous marine conglomerates, such as those outcropping at about + 200 m, immediately W of the airstrip of Santa Rosalía, can be assigned to this formation (Fig. 2). The lithology and the state of preservation of the fossils in this locality are similar to those of the Santa Rosalía Formation type-



Figure 1.- Localization sketch-map of the Santa Rosalía- La Reforma area, in Baja California Sur.

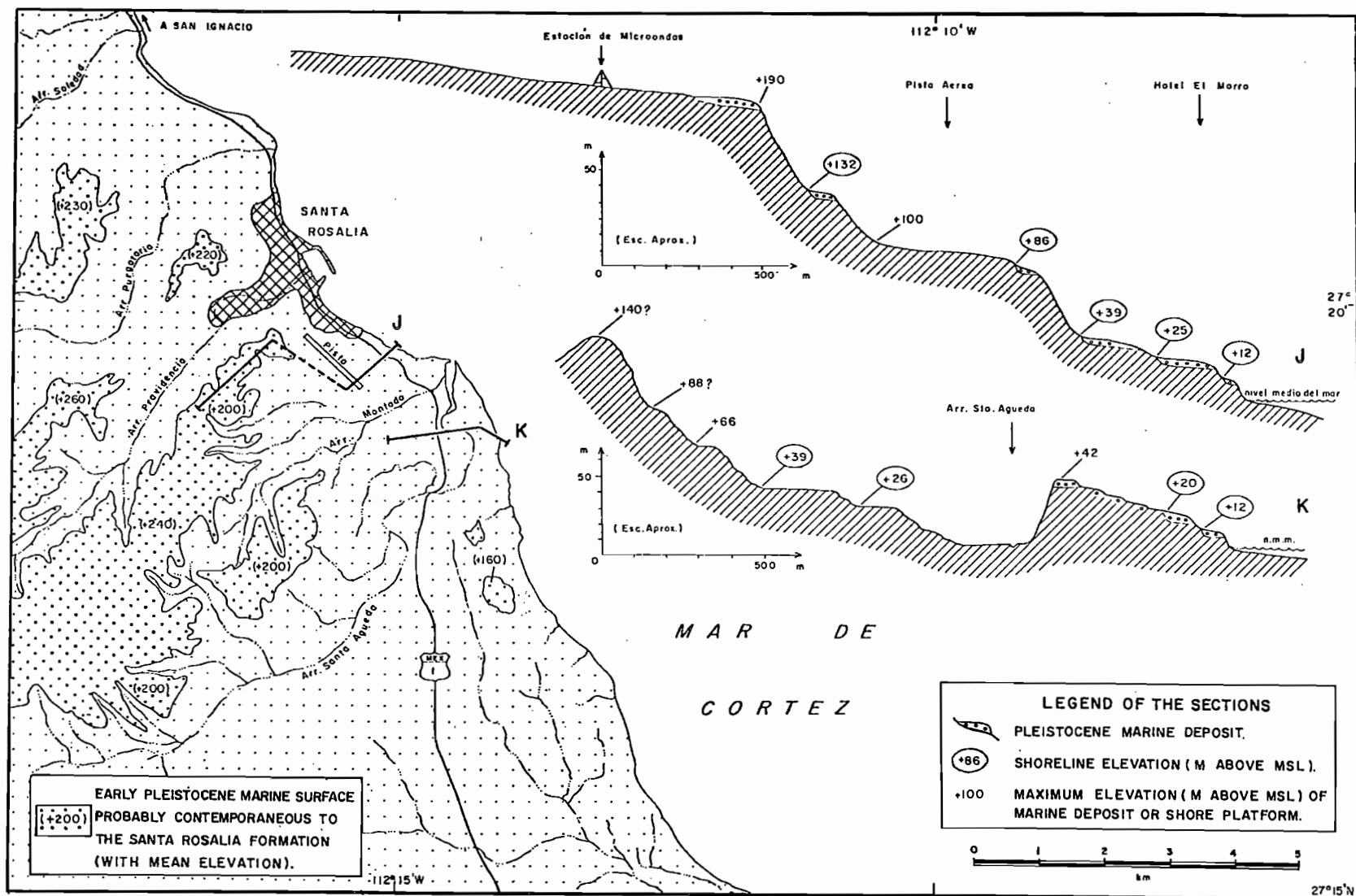


Figure 2.- Cross-sections J and K in Santa Rosalía area: localization and elevations of the Pleistocene shorelines identified south of the town.

localities of the NW part of the basin.

The upper surface of the mesas which dominate the coastal area of the basin is relatively flat and evenly inclined toward the SE; this surface was originally horizontal and has been formed by marine abrasion penecontemporaneously with the Santa Rosalía Formation. It has been affected by a few trending NNW-SSE, notably in the area of Santa Rosalía- Santa Agueda (Colletta and others, in preparation; Colletta and Angelier, this book; Angelier and others, in press).

RECONNAISSANCE STUDY OF THE MARINE TERRACES

The Santa Rosalía coast

Although the Santa Rosalía Formation was studied, and mapped, by Wilson (1948) and Wilson and Rocha Moreno (1955), these authors did not pay much attention to the marine terraces that fringe the littoral areas immediately N and S of the town. Only two of these conspicuous features were noted by Wilson and Rocha Moreno (1955). The highest marine terrace reported by them is between + 100 and + 80 m and corresponds to the wide and subhorizontal surface used for the airstrip, while the second one, between + 40 and + 20 m, constitutes the "Mesas" Francia and Mexico (respectively N and S of the Arroyo Providencia mouth) on which is built part of the town of Santa Rosalía. In a reconnaissance study of the low-lying marine terraces along the Gulf of California, Ortlieb (1978a) and Ortlieb and Malpica (1978) distinguished two terraces at Santa Rosalía, at + 8 -12 m* and at about + 20 m. Recently more complete series of marine terraces have been studied at the southern exit of the town (Ortlieb, 1979a, 1980; Colletta and others, in preparation).

Two cross-sections of the coastal area are presented here (Fig.2). Section J is perpendicular to the shore, starting near the Hotel El Morro, crosses the airstrip (Aeropuerto) and ends on the mesa near the microwave relay station (Estación de Microondas). Section K starts from the coastline, 1,5 km SE of the mouth of Arroyo Santa Agueda, heads NW and then W and terminates on top of the interfluvium between Arroyo Montado and Arroyo Santa Agueda (Fig.2). A total number of 9 distinct marine terraces is inferred from these sections (see below).

The marine terraces are identified by their morphology and the presence of typical Quaternary littoral deposits. The width of the platforms varies between a few meters to several hundred meters. The backshore angle, at the foot of the paleo-sea cliff, is not always clearly marked in the topography. The deposits associated with former high stands of sea level are either well preserved (up to 2 or 3 meters of fossiliferous sands and conglomerates) or simply consist of a lag of rounded pebbles including scattered marine shells fragments. The oldest, and highest, littoral deposits are frequently calichified, as a result of meteoric water circulation and mobilization of fossil carbonates. S of Santa Rosalía, the marine terraces have been cut in Miocene and Pliocene rocks. The most elevated ones were carved in the fine- -grained Gloria and Infierno Formations, and the lowest ones in the Boleo Formation, or in the volcanics of the Comondú Group.

This study is mainly concerned with the elevations of the distinct Pleistocene shorelines which were registered along the coast. The measurements have been performed with a precision aneroid altimeter, and the results were corrected for the variations of temperature and atmospheric pressure. When possible the shoreline elevations were measured at the backshore angle of the marine terraces. In most cases the maximum altitude reached by the paleo-sea level for each

* All the elevations reported in this work are relative to the present Mean Sea Level (MSL).

terrace could be determined either morphologically, or by interpreting the depth of deposition of the sublittoral, or littoral, sediments when preserved. When the terraces were poorly defined in the field, measurements were performed but these results are regarded as much less significant than the other ones. It must be emphasized that sound lateral and vertical correlations require that measured elevations always correspond to a common datum, which should be the highest level reached by the sea during the formation of the terraces (that is during the maximum stand of the respective interglacial periods).

The lowest and most recent paleo-coastline observed of Santa Rosalía, in both sections J and K, is at + 12 m, and the highest one is at + 190 + 200 m (section J, Fig. 2). The elevations of the 6 intermediate shorelines are found at + 20, + 25, + 39, + 66, + 86 and + 132 m (Fig.2). The two broadest Pleistocene marine platforms are the upper mesa (at about + 200 m), and the surface on which the airstrip is located (+100 m). The lateral correlation between terraces from sections J and I is treated together with that of the terraces found on the eastern flank of La Reforma complex.

The coast of La Reforma

The presence of marine terraces and benches on the coast E of the La Reforma volcanic complex was only briefly mentioned by Bowden (1973), Wright and others (1973), Schmidt (1975) and Schmidt and others (1977). In a preliminary survey, Ortlieb (1978 a) and Ortlieb and Malpica (1978) noted marine terraces at about + 10, + 20, + 50 and + 70 m in the areas of Caleta Santa María and Punta Contrabando, on the SE flank of La Reforma (See location Fig. 3). In 1978-79 additional field work was done on the coast of La Reforma, and more specifically, along nine cross-sections perpendicular to the coastline. Results of this study have already been reported (Ortlieb, 1979a, 1979b, 1980b; Demant and Ortlieb, 1979) but have not been published in full yet.

The La Reforma terraces differ from those studied just S of Santa Rosalía. The former ones generally lack fossiliferous deposits and can be described as true abrasion platforms cut in hard Plio-Pleistocene volcanic rocks. Fossil shells and beachworn organic debris are rarely found, except in some low-lying terraces, but the terraces are easily identified by unconsolidated layers of well-rounded cobbles or pebbles.

The nine sections studied on the flanks of La Reforma, about 2-3 km apart, are presented in Fig. 3. They show from 3 to 8 successive shorelines. The northernmost section (A), at Rancho La Reforma, exhibits three well-defined shorelines, between + 17 and + 70 m and another one more subdued at + 29 m. Immediately NW of Cabo Vírgenes is a spectacular set of staircase terraces, each wave-cut platform being terminated shoreward by the sea-cliff formed during a following transgressive cycle. As the terraces in the area of Cabo Vírgenes have an unusual lateral extension (3km), two sections were studied: sections B to the NW and C to the SE. Six definite shorelines were found between + 28 m and + 167 m in section B, while in section C, five definite shorelines and three indefinite ones were identified, between + 26 and + 154 m (Fig. 3). In the following section toward the S (D), at Cañon Diablo, the lowest shoreline is at + 12 m and the highest one at about + 185 m. Between Punta Prieta and Punta Contrabando, four sections (E, F, G, and H) have been surveyed; they show three to six distinct shorelines at elevations varying from + 9 to + 169 m (Fig. 3). The last section (I), at Santa María, is located at the boundary between the Santa Rosalía Basin and the La Reforma volcanic complex. Since it constitutes an important link between the two areas, it is fortunate that this section presents five definite shorelines (and two more poorly determined paleo-coastlines) at elevations ranging between + 12 and about + 166 m.

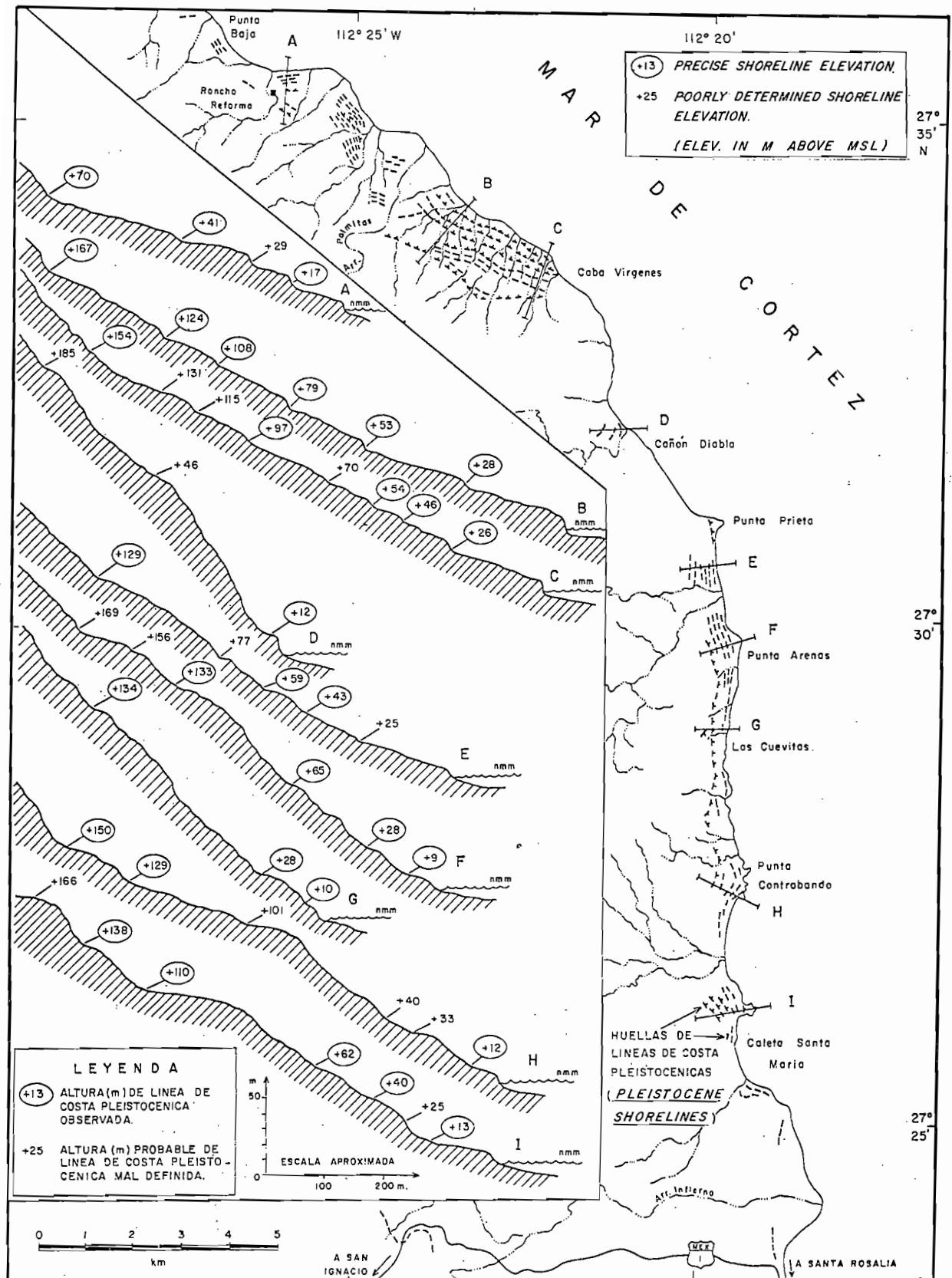


Figure 3.- Cross-sections A to I on the coast of la Reforma: localization and elevations of the observed Pleistocene shorelines.

LATERAL CORRELATIONS AND CHRONOSTRATIGRAPHY OF THE TERRACES

Each marine terrace forming part of a series, is interpreted as a remnant of a high stand of sea-level. These high stands are correlated with the warmer interglacial periods of the Pleistocene. According to deep-sea core data (oxygen isotope ratios, planktonic micropaleontology, ice-rafted debris, CaCO_3 ratios, etc.) about a dozen glacial/interglacial cycles occurred during the second half of the Pleistocene. Each cycle lasted between 80 000 and 100 000 years. The V28-238 deep-sea core isotopic curve (Shackleton and Opdyke, 1973) has been considered to provide a reliable chronostratigraphy, notably because it seems that the sedimentation rates have been constant (Fig.4). Thus it is worth trying to interpret the chronostratigraphic significance of the series of terraces observed in the area through this paleoclimatic and chronological framework.

At this moment, no radiometric dates which could help in the age determination of the shorelines are yet available. Fossil shells from terrace deposits in Santa Rosalía area have been submitted for U/Th dating, and several volcanic rocks, stratigraphically related to marine terraces features and deposits, are in process of K/Ar dating. Furthermore several samples of fossil material are being analyzed by the amino-acid racemization method. At present the only reported U/Th dates of marine terraces in the Gulf of California were obtained on the Sonoran coast, about 150 km NE of Santa Rosalía; these ages (80 000 - 120 000 years B.P.) concerned Sangamonian (last interglacial) deposits cropping out at less than + 6 m (Bernat and others, 1980). On the Pacific coast of southern Baja California, more Sangamonian fossil material, at a similar elevation, provided an average Uranium-series age of 116 500 years B.P. (Omura and others, 1979).

In this work, the Late Pleistocene is meant to cover the period 130 000 - 10 000 years B.P., and the Middle Pleistocene the period 700 000 - 130 000 years B.P. (Butzer, 1974). The age of the Pliocene/Pleistocene boundary, which has been discussed for so many years, is supposed to be 1,8 million years.

Rather than attempt to interpret directly the chronology of the marine terraces in each studied section, it seems wiser to examine, first, how do the shoreline remnants correlated from one section to the other.

Lateral correlations of the sections

The results of the altimetric measurements performed along the 11 sections of the Santa Rosalía- La Reforma area are indicated in Table 1. These data include the uncertainties in the measurements as well as for the definition of the shorelines in the field. The presentation of the values in Table 1 recapitulates the lateral correlations proposed for the distinct sections. Because the marine terraces have a restricted lateral extent, only a few of them can be directly correlated in the field. Photo-interpretation and, most of all, geometric considerations on the sequences of shoreline elevations had to be used.

In sections J and K, near Santa Rosalía, four shorelines at + 12 m, + 25 or 26 m, + 39, and + 88 or 86 m, can be easily correlated. Two more shorelines, found in section K, at + 20 m and about + 66 m, seem to have no equivalent in section J. In section J, three high well-marked terraces are observed, at + 100, + 132, and + 190 m. The latter one corresponding to the upper mesa of the Santa Rosalía Basin is tentatively correlated with the Santa Rosalía Formation (see above). To sum up, the area S of Santa Rosalía shows a total of nine distinct shorelines.

SANTA ROSALIA AREA		LA REFORMA COAST								
Santa Agueda (K)	Santa Rosalia (J)	Santa Maria (I)	Punta Contrabando (H)	Cuevitas (G)	Punta Arena (F)	Punta Prieta (E)	Cañon Diablo (D)	Cabo Virgenes S (C)	Cabo Virgenes N (B)	Rancho Reforma (A)
	190 +10 - 5				169 +2 -7		185 +7 -7			
		166 +8 -6	<u>150</u> +2 -2		156 +4 -4					
	<u>132</u> +5 -5	<u>138</u> +2 -5	<u>129</u> +2 -4	<u>134</u> +2 -2	<u>133</u> +2 -2	<u>129</u> +11 - 4		<u>154</u> +2 -2	<u>167</u> +2 -4	
	100 +6 -6	<u>110</u> +2 -7	<u>101</u> +3 -7				116 + 2 -13	131 +19 - 2		
88 +7 -5	<u>86</u> +6 -5					77 +2 -9		115 +10 -10	<u>124</u> +2 -3	
66 +6 -6		<u>62</u> +2 -6			<u>65</u> +2 -2	<u>59</u> +2 -2		<u>97</u> +2 -7	<u>108</u> +2 -5	
<u>39</u> +5 -5	<u>39</u> +3 -5	<u>40</u> +4 -2	<u>40</u> +2 -9			<u>43</u> +2 -2		70 +6 -6	<u>79</u> +2 -5	<u>70</u> +2 -2
<u>26</u> +4 -4	<u>25</u> +5 -2	25 +7 -7	33 +2 -8	<u>28</u> +2 -2	<u>28</u> +2 -8	25 +4 -5		<u>54</u> +2 -5	<u>53</u> +2 -5	<u>41</u> +2 -2
<u>20</u> +5 -5								<u>46</u> +2 -6		29 +2 -6
<u>12</u> +5 -3	<u>12</u> +2 -3	<u>13</u> +2 -5	<u>12</u> +2 -2	<u>10</u> +2 -4	<u>9</u> +2 -2		<u>12</u> +2 -2	<u>26</u> +2 -6	<u>28</u> +2 -4	<u>17</u> +6 -2

Table 1 Measured elevations (m above present MSL) of shorelines observed along 11 sections across the coastal area of La Reforma and the Santa Rosalia Basin (Localization in figs. 2 and 3). The accuracy of the measures of elevation mainly depends on the precision with which paleo-sea levels could be determined in the field (the most reliable data are indicated here by underlined values); the distinct ranges stated take account of the indefiniteness of the maximum elevation reached by paleo-sea levels, in each case. The horizontal lines show the suggested lateral correlation (see text).

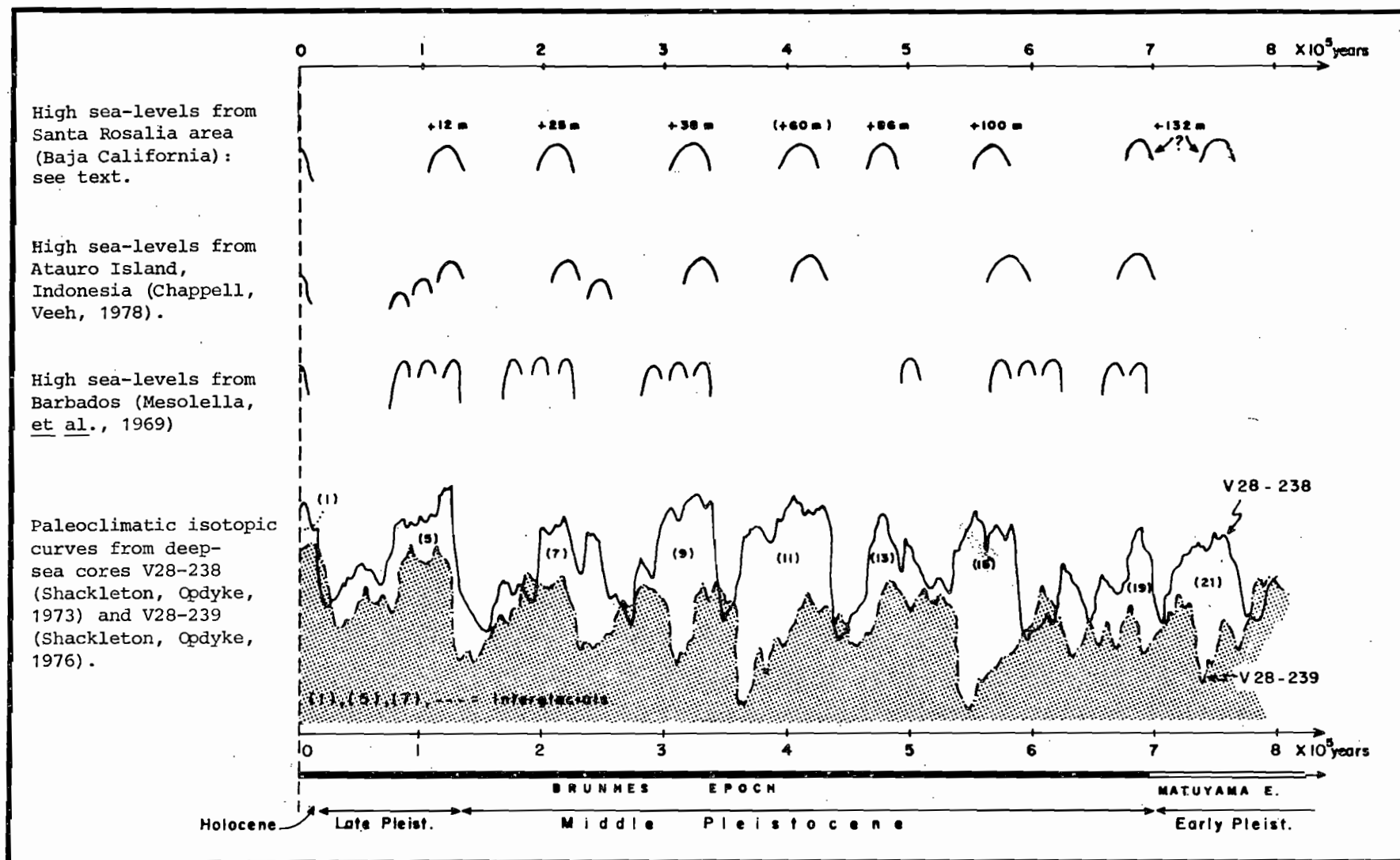


Figure 4.- Speculative correlation between the series of marine terraces from Santa Rosalia (Section J), from Atauro Island and from Barbados, and chronostratigraphic interpretation of these shorelines based on the V28-238 and V28-239 isotopic curves (the most reliable paleoclimatic scale is supposed to be provided by the V28-238 deep-sea core). According to this correlation, the lowest emerged terrace in Santa Rosalia area would be early Late Pleistocene, and the next five or six more elevated shorelines (between +25 and + 132 m) would be Middle Pleistocene.

Between sections J and I, respectively at Santa Rosalía and Santa María, the lateral correlation of the three lowest shorelines is easy to make, since their respective elevations are practically the same (+ 12 and + 13 m; + 25 m; + 39 and + 40 m). The shorelines found at + 110, and + 138 m, at Santa María are interpreted to correspond to those at + 100, and + 132 m, in the Santa Rosalía section, and thus would suggest a slightly more rapid uplift in the former area. The shorelines observed at + 62 and at about + 166 m, in the Santa María section, apparently have no lateral equivalent in the Santa Rosalía section; similarly the + 190 and + 86 m littoral deposits of section J are thought to have no equivalents in sections I. So, on the basis of the proposed correlations between sections I, J and K, there would be a total number of 10 shorelines

Section D, E, F, G, and H are more poorly documented than the preceding ones, and therefore correlation of the scattered shorelines data is rather more hazardous. However, two series of values are noteworthy: a higher series at + 129 m (section E), + 133 m (F), + 134 m (G) and + 129 m (H) and a lower series at + 12 m (D), + 9 m (F), + 10 m (G) and + 12 m (H). The former series is tentatively correlated with the + 138 m (I) or + 132 m (J) shorelines, and the latter with the + 13 m (I) or + 12 m (J, K) shoreline, both previously described in the southernmost sections.

Between sections C and B, at Cabo Vírgenes, most of the marine terraces can be followed in the field and on air-photographs. So the shorelines correlation is in this case better substantiated, even if the exact elevation of maximum stands of sea-level was sometimes difficult to determine (particularly in section C). The backshore angle of the lowest terrace was found at + 26 m (C) or + 28 m (B), and no other remnant of sea-level stand could be observed, in either section, between the present (Holocene) coastline and this elevation. The base of the next following paleo-sea cliff is at + 54 m (C) or + 53 m (B). A small littoral scarp is observed at + 46 m, in section C, but this shoreline progressively disappears to the NW and is thus interpreted as a remnant of a short stand of sea-level that occurred during the regressive phase of the last but one glacial/interglacial cycle. The next following more elevated shorelines are found systematically at higher elevations in section B with respect to section C: + 79 m (instead of + 70 m), + 180 m (+ 97 m), + 124 m (+ 115 m), and + 167 m (+ 154 m). The lateral equivalent of the + 131 m shoreline of section C was not identified in section B. If the + 46 m shoreline of section C is excluded, a total number of 7 shorelines were thus counted in the area of Cabo Vírgenes. Because of the regular spacing of the step-like terraces it is supposed that the coast at Cabo Vírgenes (sections C and B) recorded the last 7 high stands of sea-level, probably corresponding to the last 7 interglacials.

The lateral correlation between sections B and C, and the other sections to the S, depends on the interpretation made for the + 26 (C) or + 28 m (B) shoreline. If the elevations alone are taken into account, it would seem consistent to correlate the + 25 or + 26 m shoreline of sections I, J, and K with the + 26 m shoreline of sections B and C; nevertheless field data show that the latter is the lowest and youngest shoreline while the former is an older one. Consequently the same youngest paleo-shoreline is thought to be presently registered at about + 12 m in almost all the southernmost sections and at about + 27 m at Cabo Vírgenes. The correlation of the more elevated shorelines, between sections B and C on one hand, and sections D to K on the other hand, is made according to this pairing of the lowest terraces.

In the northernmost section (A), where only four shorelines were identified,

it seems that uplift rates have been intermediary between those of the Cabo Vírgenes and those of the more southern localities. The lowest observed shoreline is situated at + 17 m. The next higher shoreline was poorly determined in the field, and for that reason might be the lateral equivalent of the faint + 46 m shoreline of section C (and possibly, of the + 20 m shoreline of section K). The two highest definite shorelines are thought to correspond to the + 53 m and + 79 m shorelines of section B.

In conclusion, in the whole area, ten successive shorelines have been identified. One of these shorelines is probably due to a short stillstand during a regressive phase, and the nine others are interpreted as remnants of interglacial high stands of sea-level. The three youngest terraces and the seventh in elevation are generally better preserved than the other ones. Along the coast S of Cabo Vírgenes the shorelines suffered very little local deformations. At Cabo Vírgenes the vertical spacing of the shorelines is as regular as in the other sections but the height difference between successive shorelines is approximately double as compared to what it is in the more southern sections. To be able to compare uplift rates in this coastal area, it is necessary to interpret the chronology of the shorelines.

Chronostratigraphic and paleoclimatic interpretation

According to the V28-238 chronostratigraphy (Shackleton and Opdyke, 1973) the Late Pleistocene was characterized by 4 isotopic stages, and the Middle Pleistocene by 14. Each isotopic stage is considered to correspond to a half glacial/interglacial cycle (except, probably, for stages 2 and 3). So there were probably 9 interglacial high stands of sea-level during the period 700 000- 10 000 years B. P. On the V28-238 paleoclimatic curve, the early Late Pleistocene isotopic stage 5, and the Middle Pleistocene isotopic stages 7, 9, 11, 13 and 15 are well-marked and somewhat better defined than the earlier ones (stages 17, 19, 21 and 23) (Fig.4). The number of glacial/interglacial cycles of the Early Pleistocene is not yet firmly established but is probably low, the Early Pleistocene being often referred to as the "preglacial Pleistocene".

The total number of shorelines found in the Santa Rosalía-La Reforma area compares well with the number of interglacials as interpreted from the V28-238 chronostratigraphy, and with the number of successive marine terraces described in other parts of the world (e.g. Stearns, 1935; Colquhoun, 1965; Broecker and others, 1968; Ward and others, 1971; Bloom and others, 1974; Chappell and Veeh, 1978; Bender and others, 1979).

The lowest and most recent shoreline, which is generally well-defined, is at about + 12 m in the studied area. It is interpreted as early Late Pleistocene, as are the numerous remnants of a low terrace found all along the coasts of the Gulf of California (Ortlieb, 1978a; 1978b; 1979a; 1979b; Ortlieb and Malpica, 1979). It is noteworthy that this lowest terrace is at a significantly higher elevation than in any other locality of the Gulf, except probably at the northermost extremity of the gulf (Ortlieb, 1978b; 1980a). As noted above, the coast at Cabo Vírgenes exhibits its lowest shoreline, and thus probably Late Pleistocene in age, at an elevation much higher than to the E and the S of La Reforma. In fact, the + 28 m shoreline observed at Cabo Vírgenes is thought to be the most elevated remnant of the last interglacial sea-level from the whole Gulf of California. There is no general agreement as to whether the Late Pleistocene maximum stand(s) of sea level was (were) one- or threefold, and in the latter case what were the amplitudes of the fluctuations (Hollin, 1965 and 1980; Steinen and others, 1973; Bloom and others, 1974; Stearns, 1976). In the Gulf of California area no positive field evidence of more than one Late Pleistocene high stand of sea-level has been found, up to now.

The lack of any conspicuous subsequent paleoshoreline below the + 28 m elevation might be a further indication that, if the sea-level rose again after the + 28m stand, it did not reach an elevation comparable to the "maximum interglacial level".

In most of the studied sections a similar difference in height is found between the present mean sea level and the first (lowest) paleo-shoreline, as between the first and the second paleo-shoreline, and also as between the second and the third ones. These height differences are about 10-15 meters, except at Cabo Vírgenes where they are about 27 meters. Assuming that the last Quaternary climatic variations have been cyclic, that the interglacial high stands of sea-level have been eustatically equivalent, and that a relatively regular uplift took place in the area, the constant height difference observed along each section leads to consider that the last and lowest terrace is Late Pleistocene (isotopic stage 5), and that the next two higher terraces probably correspond to the isotopic stages 7 and 9 of the V28-238 chronostratigraphy (Fig.4). The approximate ages of the two sea-level maxima as deduced from the V28-238 curve (Figs. 4 and 5) are 220 000 and 320 000 years B. P. In the case of the highest shorelines, height differences are more variable than between the youngest ones. Besides, the successive older shorelines have not always been registered (or identified), and so the study of the height difference ratios, between all the sections, is more difficult. A chronostratigraphic interpretation of the series of terraces must take into account the missing shorelines, and so rely on the lateral correlation framework (Table 1).

Three sections are better documented than the others, and moreover are distributed along the whole studied area: Santa Rosalía (J), Santa María (I) and Cabo Vírgenes (B); at Cabo Vírgenes, section B was preferred to section C because its shorelines are more definite in the field, and thus the elevation data more reliable. The chronostratigraphic interpretation of the series of terraces from the Santa Rosalía-La Reforma area are based on these three sections.

Section J, when compared to section K and I seems to be lacking a shoreline at about + 60 m, and another one at about + 160 m. The best correlation fit between the shorelines data of section J and the V28-238 chronostratigraphy is indicated in Fig. 4. This correlation results in assigning each shoreline to an "interglacial" isotopic stage (5 to 15). Because of the relatively large vertical spacing of the +100 m and + 132 m shorelines, and because of the shortness of the peak 17 in the V28-238 isotopic curve, it is inferred that the high sea-level corresponding to that interglacial did not reach the full "interglacial eustatic level". The time correlation of the + 132 m shoreline with either (or both?) isotopic stages 19 or (and?) 21 (Fig. 4) is speculative. According to the preceding correlations, and assuming that the uplift rate has been approximately the same, both the probably missing shoreline at about + 160 m, and the + 190 m fossiliferous deposit, may be assigned to the late Early Pleistocene. The age of the + 190 m deposit, being interpreted to belong to the Santa Rosalía Formation, may then be estimated to be about 1 000 000 years B.P. To sum up, the shorelines at + 12 m, + 25 m, + 39 m, + 86 m, + 100 m and + 132, observed in section J are interpreted to correspond respectively to the V28- 238 isotopic stages 5, 7, 9, 13, 15, 19 and (or) 21, while the most elevated shoreline (at + 190 m) is thought to be late Early Pleistocene.

From the tentative chronology suggested for the sequence of shorelines in section (Fig.4), and the lateral correlation proposed in Table 1, possible ages of the shorelines of sections B and I (Fig. 5) are deduced. In section I, the + 62 m, + 110 m, + 138 m and + 166 m shorelines are respectively correlated with the V28- 238 isotopic stages 11, 15, 19 and 21. In section B, the + 108 m, +124 m, and + 167 m shorelines are thought to correspond to stages 11, 13 and 19.A

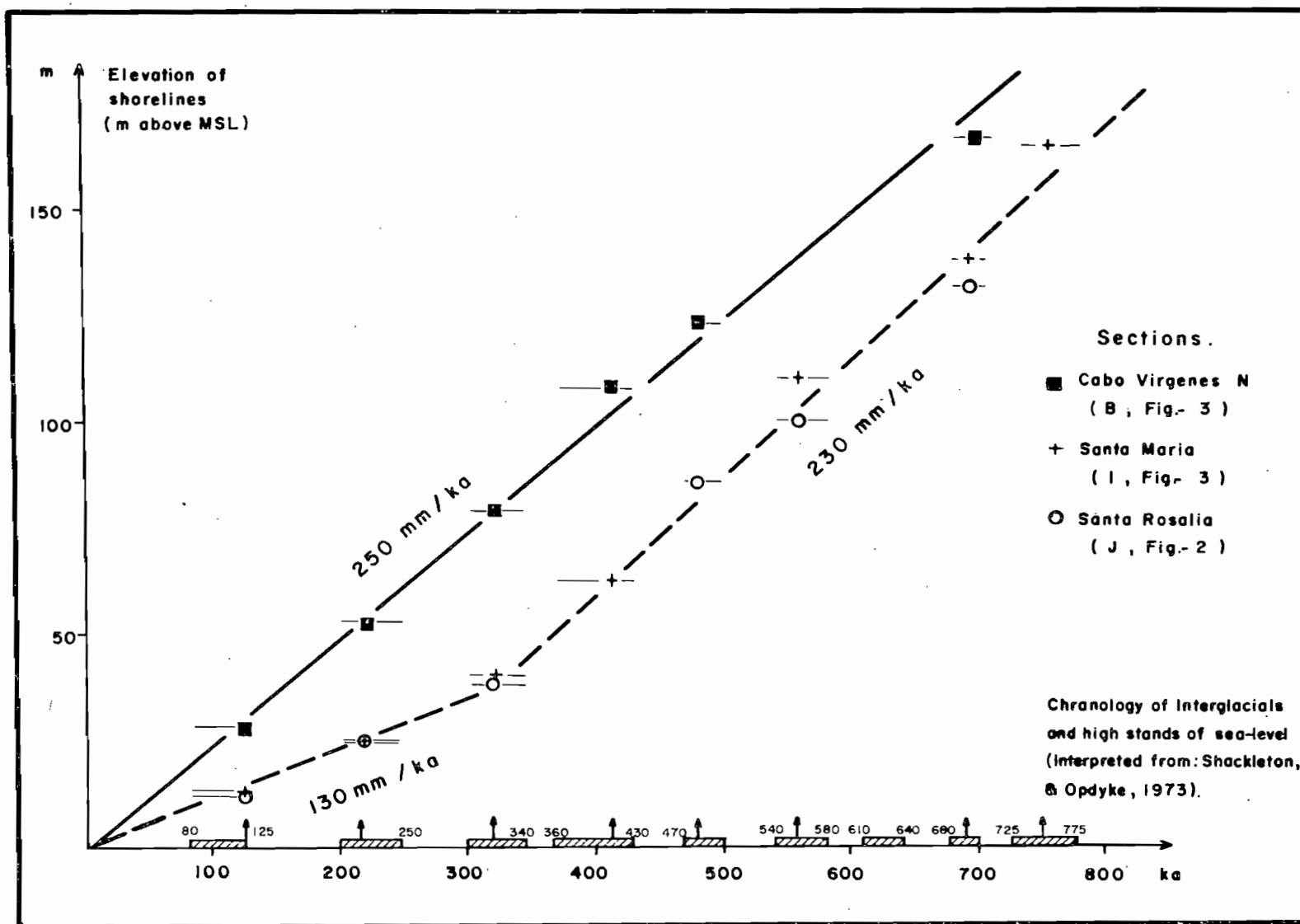


Figure 5.- Tentative elevation-age plot of shorelines from sections B, I and J. Vertical arrows on the time scale correspond to the peak of each "interglacial" stage deduced from the V28-238 isotopic curve (see Fig. 4); hatched boxes and thin horizontal lines figure the duration of the isotopic stages. The plot supports a fair correlation of the successive shorelines with most of the Middle- Late Pleistocene high stands of sea-level. Suggested uplift rates are indicated along the distinct correlation curves.

plot of shorelines elevations versus maximum interglacial ages based on the V28-238 chronostratigraphy (Fig. 5) shows a reasonably good correlative relationship in each section (B, I and J). This plot illustrates that the proposed correlations (both vertical and lateral) suggest different uplift rates between section B on one hand, and sections I and J on the other. The elevation-age plot of the shorelines of section B determines a straight line indicative of a constant uplift rate of 250 mm/ka, and the correlation curve of sections I and J is composed of two straight segments suggesting uplift rates of 230 and 130 mm/ka. Thus the Cabo Vírgenes area would have been submitted to a constant vertical motion since Early Pleistocene time, while the more southern coastal area probably has been uplifted, at a similar rate, during the Early Pleistocene and the first half of the Middle Pleistocene, and afterwards at almost half this rate (130 mm/ka). The break in the elevation-age correlative curve is situated between the isotopic stages 9 and 11, that is probably around 350 000 years B.P.

CONCLUSION

If many shorelines were registered in the coastal area of the Santa Rosalía Basin and La Reforma, it has been because vertical movements occurred during the major part of the Pleistocene. The relatively regular vertical spacing of marine terraces strongly suggests that these movements have been continuous. According to the proposed chronostratigraphic correlation of the distinct shorelines in the 11 studied sections, it is inferred that the whole coastal area has been uplifted, since about 1 000 000 (?) years B. P. until about 350 000 years B.P., at a rate averaging 240 mm/ka. After 350 000 years B. P., only one locality (Cabo Vírgenes) has continued to be uplifted at this rate, while the rest of the coastal area has been suffering reduced (rate about 130 mm/ka) but continuous vertical movements.

Marine terraces have not been found inland, except for the Santa Rosalía Formation. This formation is closely associated with a morphologic surface (the upper mesas) which evenly slopes down from more than + 300 m, in the NW of the basin (at "Cuesta del Infierno", km 17 of Highway 1), to less than + 200 m at the mouth of Arroyo Santa Agueda (see Fig. 2).

This tilting is related to the volcano-tectonic history of the La Reforma caldera. Since deposits of the Santa Rosalía Formation predate the principal ignimbritic eruption of La Reforma, this unit was most probably tilted during the last stages of the intumescence phase (Demant and Ortlieb, 1979; Demant, this book). This phenomenon accounts for at least part of the inclination of the upper mesas, but cannot be alleged to explain the series of uplifted shorelines observed on the coast.

The whole coastal area, including the La Reforma eastern flank and the surroundings of Santa Rosalía, has been uplifted at a uniform rate, in such a way that no gradient in the vertical motions could be detected. Evidence was provided that Cerro La Reforma, inside the caldera, has been strongly uplifted during the Quaternary; this uplift which might have reached 1 000 m in a million years, has been related to block-faulting and "resurgent doming" (Demant, this book). The mean uplift rate, of the order of 1000 mm/k.a., suggested for the Cerro La Reforma, is several times superior to the rates interpreted from the marine terrace data. Notwithstanding those discrepancies, a regional component of this "resurgent doming" is probably involved in the uplift of the studied coastal area.

Finally, part of the Quaternary vertical motions in the Santa Rosalía-La Reforma area is thought to be related to a regional upheaval of east central Baja California (Ortlieb, 1980b). Since Pleistocene shorelines are found at up to + 50 to + 100 m elevations, in distinct localities along the coast between Santa Rosalía and Loreto, it is clear that tectonic forces, either local or regional, have been active in the second half of the Quaternary. A global comparison of the tectonic behavior on both sides of the Gulf of California shows that the eastern coast of Baja California has been generally uplifted, including during Late Pleistocene time.

More studies on the sequences of marine terraces, and particularly geochronological results, presently underway, are still needed to substantiate the tentative lateral correlations and the chronostratigraphy of the Pleistocene shorelines.

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PLIO-QUATERNARY VOLCANISM OF THE SANTA ROSALIA AREA,
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INTRODUCTION

The Santa Rosalía area, located near the 27°30' N parallel on the western coast of the Gulf of California, has been known since the end of the last century for its ore deposits. A new copper mineral, the boleite, was described from this locality (Fuchs, 1886). The Boleo copper veins are interbedded in a thick Pliocene marine sequence (Wilson, 1948). Such marine sediments are more common along the Pacific coast (Darton, 1921; Heim, 1915 and 1922; Mina, 1956 and 1957) than in the eastern part of the peninsula.

From the volcanological point of view, two Plio-Quaternary structures are of particular interest: (1) the Sierra La Reforma caldera (Demant and Ortlieb, 1979), clearly recognizable on the ERTS-LANDSAT documents, and (2) the southernmost cone of the Tres Vírgenes Group (Mooser and Reyes, 1961), the only historically active volcano in Baja California (Ives, 1962). This volcanism is part of an Upper Miocene-Plio-Quaternary magmatic sequence well represented all along the peninsula (Gastil et al, 1975 and 1979).

Considering the general geology of Baja California, the Santa Rosalía area represents a key zone between the Cretaceous batholithic belt intrusive in a volcano-sedimentary sequence, the Alisitos Formation (Gastil et al, 1975), to the north, and the Middle Miocene (McFall, 1968; Gastil et al, 1979) volcanic and sedimentary rocks (Comondú Formation) of the Sierra La Giganta, to the south.

The oldest rocks found in the Santa Rosalía region are quartz monzonites (Wilson, 1948) dated by Schmidt (1975) 91 ± 2 , 1 m.y. Resting on the Cretaceous basement are the Santa Lucía volcanic rocks (Fig. 1), northern limit of the Comondú Formation. They constitute a typical calcalkaline sequence, from basalt to dacite, with some pyroclastic tuffs and breccias (Demant, 1975). In Pliocene time, this irregular paleotopography was invaded by a marine gulf extending as far as 10 Km west of the present coastline. The littoral sedimentation is rather complex and was first studied in detail in the mining district (Wilson, 1948; Wilson and Rocha, 1955). The post-Pliocene volcanism is referred by Wilson (1948) and Schmidt (1975) as Tres Vírgenes Volcanics, but I will not use this designation since these products came from two different volcanic centers; the Reforma caldera and the Tres Vírgenes volcano.

THE TRES VIRGENES VOLCANO

The name Tres Vírgenes designate a group of three volcanoes lying in a NE-SW direction (Mooser and Reyes, 1961), but only one of them, the southernmost, was active during the late Quaternary. This youngest cone is built on a Plio-

Quaternary basaltic plateau overlying Tertiary continental sediments and pyroclastic tuffs of the Comondú formation. The first lavas are relatively viscous and dacitic in composition. Some of these flows, to the north, show well marked "levées" and ridges. At the end of this late Quaternary episode, the volcanic cone reached 2,000 m high.

The historic activity of the Tres Vírgenes volcano is clearly owing to the existence of 18th century manuscripts written by the Jesuit Missionaries present at this time in Baja California and according to Ives (1962) an eruption occurred in 1746. Although there was no observation of this eruption, it is possible to reconstruct the succession of the volcanic events by field observations.

The eruption started with a major explosive phase responsible for the important pumice deposits that cover a wide area around the volcano. These pumices do not represent a plinian type eruption, as defined by Lirer et al. (1973) from the last eruption of the Vesuvius, but more likely a succession of gas blasts, a kind of mechanism that happens in strato-volcanoes in their mature stage. The Pumice beds 10 to 20 cm thick, alternate with fine ash falls which correspond to drops in gas pressure during the eruption. As a result of the explosive activity, a crater, 1 Km in diameter, was formed on the upper part of the Tres Vírgenes cone and big fractures notched their slopes. A dacitic lava was then poured out from a fissure on the southern flank, near the base of the volcano, forming a short (1.5 Km) but thick (≈ 100 m) flow, a small dacitic plug was filling up the vent.

Near the SW limit of the summit crater, a small cinder cone built up and the black ashy material it generated covers the top and the slopes of the volcano, - clearly demonstrating a change in the nature of the magma. The lavas, emitted through a NE-SW fracture, flowed down the flank and formed two malpais at the south-western and south-eastern base of the cone. These are basaltic in composition and have a typical block-lava morphology. The three small viscous dacitic flows (Fig. 1) observed south of the Tres Vírgenes Volcano are also recent, as deduced from their sparse bush cover and their morphology.

The other products emitted by the Tres Vírgenes Volcano are not related to the dacitic ones, but could be due to fissural eruption. Yet, considering the close relations in time and space of these two magmas, the question might be raised as to whether the basaltic lava is responsible of the triggering of the pumicitic eruption, as was considered in many similar events (Sparks, et al, 1977; Self and Williams, 1979).

THE SIERRA LA REFORMA VOLCANICS

This wide volcanic structure located 20 km NW from Santa Rosalía presents most of the elements typical of a caldera i.e.: (1) ignimbrite deposits, (2) a central collapse (10 km in diameter) and (3) viscous domes located near the margin of the caldera. The only particular phenomenon that can explain the morphology of the Reforma Caldera, is the large uplift of the central block. Nevertheless, Schmidt (1975) and Schmidt et al, (1977) assert that in spite of its circular outline, the Sierra La Reforma is not a caldera, but a complex system of tectonic blocks, the vertical movements involved being related to the Gulf of California rift system.

Two different periods can be distinguished in the evolution of the Sierra La Reforma: (1) a Pliocene sequence, mainly submarine, and (2) a Quaternary activity responsible for the present morphology.

1.- The pre-Reforma Pliocene volcanic sequence

It is rather difficult to reconstitute the Pliocene volcanic centers as the older rocks only outcrop in the big cañon cutting the more recent products (Fig.1),

GEOLOGICAL MAP OF THE SANTA ROSALIA AREA (BAJA CALIFORNIA)

MAPA GEOLOGICO DE LA REGION DE SANTA ROSALIA (BAJA CALIFORNIA SUR)

0 1 2 3 4 5 km

A. DEMANT, 1981

LEGEND

TRES VIRGENES VOLCANO

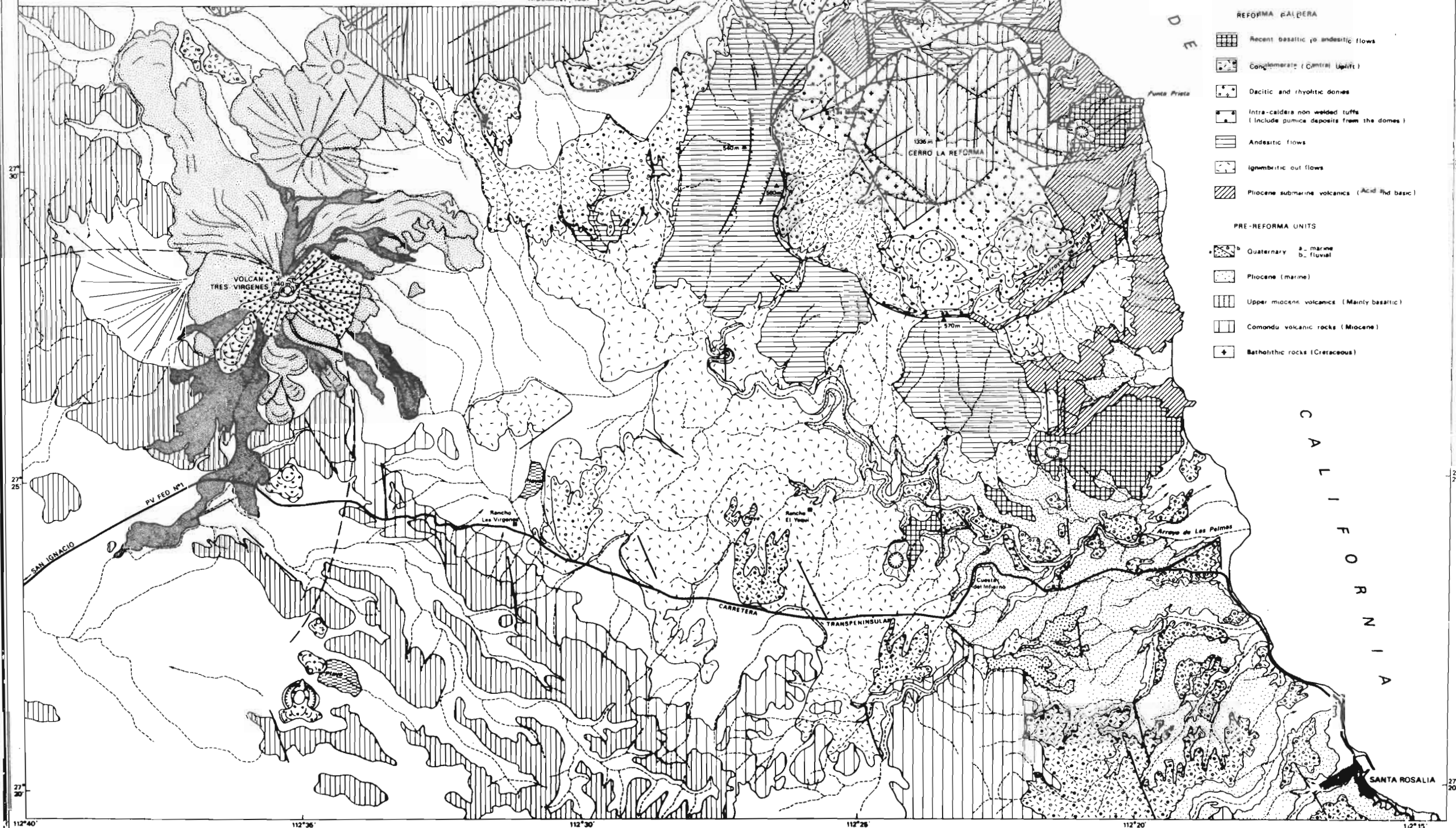
- Basaltic emissions a. flows b. cinders
- Recent dacitic flows and summit dome
- Limit of the pumice deposits
- Andesitic to dacitic flows (Tres Virgenes volcano and oldest cones to the NE)

REFORMA CALERA

- Recent basaltic to andesitic flows
- Conglomerate (Central Uplift)
- Dacitic and rhyolitic domes
- Intra-caldera non welded tuffs (Include pumice deposits from the domes)
- Andesitic flows
- Igneimbritic out flows
- Pliocene submarine volcanics (Acid and basic)

PRE-REFORMA UNITS

- Quaternary a. marine b. fluvial
- Pliocene (marine)
- Upper miocene volcanics (Mainly basaltic)
- Comundu volcanic rocks (Miocene)
- Batholithic rocks (Cretaceous)



and are included in the marine Pliocene sequence. Pliocene sediments are found up to the center of the caldera, in Las Minillas locality for example, and are mostly yellowish sands and clays, very similar in aspect with the middle Pliocene Gloria Formation defined in the Santa Rosalía area (Wilson and Rocha, 1955). The volcanic activity in the Sierra La Reforma region probably started some 3 m.y. ago.

The first Pliocene volcanic units are pumicitic flows found in the Tertiary sediments. Examples of acid products erupted in a submarine surrounding are rare and in this respect Sierra La Reforma is an interesting area. The pumicitic lava fragments are rounded, enclosed in the yellow Pliocene sands, and they often resemble the pillow structures produced by submarine basaltic eruption. Such pumice flow can be well observed near the coast-line, on the southern rim of the Arroyo del Contrabando, south of the Cerro La Reforma, and also near the Rancho La Reforma, to the north (Fig.1). So they are not localized events, but part of an important acidic eruptive sequence developed near to the gulf coast. Very thick (100-150 m) layers of ash pumice falla deposited in the sea, as demonstrated by the included pecten shells, are also present and are conspicuous in the southeastern part of Cerro La Reforma, cut by deep cañons. Pumice flow and pumice falls result from plinian-like explosive activity, and lava flow are relatively scarce. Nevertheless in the Arroyo de Punta Arenas, a small obsidian mass can be seen intruding the Pliocene sediments. The brecciated surface of the lava underline a rapid cooling due to the contact with water saturated material. Another outcrop of acidic lava is located on the northern rim of the Caño del Diablo, near the shoreline.

The next volcanic unit in the sequence is also submarine, but predominantly basic in composition. It forms a complex system of dykes, sills and lopolites intrusives in the Pliocene sediments as well as in the ash and pumice falls. Prismatic is always well developed. The explosive nature of the contact between magma and water explains the formation of brownish palagonitic tuffs, visible for example in the Arroyo de Punta Arenas. Lava flows are sometimes present, as in the northern rim of the Arroyo de La Reforma, where a 30 m thick aphyric lava unit appears among the Pliocene sediments. Pillow structures were formed in the upper part of the flow, while vertical prismatic are ubiquitous in the lower part. Schmidt (1975) describes in the Sierra La Reforma granodioritic stocks. One of them, located in the Las Minillas region (Fig.1) was dated 1.34 ± 0.50 m.y., galene deposits and native sulfur are associated with this "Plutonic event". Two other intrusives, the Cañon del Diablo and the Borrego stocks were distinguished by the same author in the north-eastern part of the Reforma caldera. The exact stratigraphic position of these rocks is not very clear on Schmidt's map.

They outcrop in an area near the gulf coast where numerous basic Pliocene lavas are visible, and in Pliocene sediments (eg. at Las Minillas) and may be considered as part of the pre-Reforma basic intrusive sequence. This interpretation is also supported by the close relationship between the plutonic stocks and the lava flows (Schmidt, 1975).

In the northern rim of Arroyo de la Reforma (Fig. 1), the submarine lavas are overlain, without angular unconformity, by aerial flows; the emersion of this zone was thus progressive. The latter sequence, named Aro volcanics by Schmidt (1975), was dated by the K/Ar method in 1.09 ± 0.12 m. y. The Aro volcanics are clearly related with the disappearance, in the Pleistocene time, of the Santa Rosalía-La Reforma marine gulf. This sequence outcrops essentially in the eastern and north-eastern part of the Sierra La Reforma; a black glassy lava of intermediate composition, with elongated feldspar crystals, is a very common facies.

The pre-Reforma volcanism is thus represented by important acid explosive

phases followed by numerous intrusions and associated lava flows of more basic composition that were emplaced under water. These volcanic products are not associated with a well defined vent, but the close geographic relation with the more recent Reforma lavas is probably not fortuitous. This sequence ends with aerial eruptions in the Middle Quaternary indicating that the Reforma caldera is very recent, and certainly less than 1 m. y. old.

2.- The Reforma caldera

The caldera forming eruptions are grey colored welded tuffs extending throughout the southern and western regional of Cerro La Reforma, covering an area of at least 100 km². In some places, a vitrophyric zone (Smith, 1960a ; 1960 b) can be observed in the lower part of this 10 to more than 50 m thick unit. The ignimbrite plateau is not horizontal, but radially inclined from the center of the structure, and corresponds with the present Cerro La Reforma. The altitude of the ignimbrite deposit varies from 590 m, near the limit of the caldera, to 400 m, 10 km farther south. Such a regular slope ($\approx 2\%$) is not the mark of an original topography, but the result of a regional uplift related to the development of the magmatic chamber, named intumescence stage by Smith and Bailey (1968).

In the upper part of the Cuesta del Infierno, 15 km WNW of Santa Rosalía, the ignimbrite unit covers a fossils rich marine conglomerate of the Santa Rosalía Formation (Wilson, 1948), situated at + 290 m. This deposit represents the highest Quaternary marine terrace in the Gulf of California (Ortlieb, 1977) and serves to indicate the amplitude of recent vertical movements in this region. In this same locality the first products issued from the Reforma caldera, pumice falls capped by white ash falls can be also observed; this 10 m thick unit is the result of a plinian-type eruption. The close association of plinian and ignimbritic deposits is well explain by the dynamic of these eruptions. As was pointed out by Wilson and Sparks (1979), the mechanisms of both explosions are rather similar, but gas rich eruptions give rise to plinian clouds climbing high in the atmosphere, whereas in the ignimbritic eruptions, the gas and magma mixture is denser and thus spreads horizontally. Between the ash and pumice deposits of the Cuesta del Infierno and the ignimbritic unit, basaltic scoriae are present. Thus basaltic magmas are closely associated with acid products in building up the Reforma caldera. This is also demonstrated by the large basaltic flows, 2 to 10 m thick, that cover the southern and western part of the ignimbritic plateau. In the western part of the Reforma caldera, the volcanic units have filled up a paleo-valley, and are recovered by talus deposits of the Middle Quaternary epoch. The volume of the acid pyroclastic deposits can be estimated in ≈ 5 km³, while the basaltic one is widely inferior ($\approx 0,5$ km³) due to the thinness of the lava flows.

As a result of the rapid emptying of the magma chamber, a caldera structure, 10 km in diameter and 100 m deep, was formed. A system of concentric fractures cuts the ignimbrites and the basaltic lavas. The structure of the central part is rather complex, as it was affected by later vertical movements and volcanic events. Yellow non-welded tuffs are widespread in this region; they overlie volcanic rocks of the Comondú Formation and the Pliocene sediments. As these tuffs were deposited in an aerial surrounding, they cannot be considered as equivalent to the eastern pre-caldera ash and pumice falls, but are probably related with the Reforma ignimbritic eruptions.

Three rhyolitic domes-coulées, very recent judging from their morphology (since the crests and other flow structures are well preserved) are located near - the southeastern rim of the caldera. Obsidian is very common on the front and -

margins of the flows. Ash and pumice fall deposits cover all the adjacent area and demonstrate the explosive character of the first stage development of these domes. Another lava dome is present on the north-western flank of the caldera, but is cut by the fracture of the Arroyo de Las Palmitas. The eruption of acid magma in caldera structures generally occurs after central subsidence and the resurgent doming phenomenon (Smith and Bailey, 1968), result of the hydrostatic re-equilibration of the collapsed mass, as seen in the classical example of Valles Caldera, New Mexico (Smith et al, 1970).

The most outstanding morphological feature of the Reforma caldera is the tremendous uplift of the Cerro La Reforma. This block of Miocene Comondú lavas is the highest point of the area (1,350 m) and it rises more than 800 m over the caldera rim. Yellow tuffs are preserved in some places on its top. The Cerro de La Reforma is limited by faults, and thus corresponds to a tectonic block uplifted by a piston-like movement. This vertical thrust was not uniform, but more pronounced in the south-western part. This difference explains the general tilt toward the north-east of the Cerro La Reforma block and the outcropping of Cretaceous quartz monzonite in the southwestern side of the mountain. The rapid uplift ($\approx 1,000$ m in 1 m.y.) has induced an important erosion and the formation of a wide detrital fan that fills up the southwestern part of the collapsed center (Fig.1) and conceals the topographical steep slopes of the caldera in this region. Some volcanic eruptions occurred during the uplift, as demonstrated by the presence of three pyroclastic units interbedded with the conglomerate named Palmas Terrestrial Conglomerate by Schimdt (1975). This author considers the detrital formation as Pliocene in age but it was in fact formed during the Middle Quaternary, after the development of the domes.

Such vertical movement was interpreted in the La Primavera caldera located near Guadalajara, Mexico, as renewed resurgence of magma that may carry the development of a second eruptive cycle (Mahood, 1980). In that case, the uplift (≈ 500 m) would concern the entire center of the caldera, so that the original collapsed zone would be completely concealed and replaced by a convex morphology. In the Reforma caldera, however the vertical movement is rather distinctive. Only part of the central zone, the Cerro La Reforma, probably located upon the magma chamber, was strongly uplifted, but as a consequence of this rise, blocks of Comondú rocks were also tilted in the western, southern and eastern margins (Fig.1). The northern flank of the caldera also moved upwards, thus the Arroyo de Las Palmillas flows to the north, and not alongside the edge of the caldera as the Arroyo de Punta Arenas does in the south-eastern region (Fig.1). Other evidences of the uplift of the northern part are the higher altitudes of the caldera rim (700 m) and of the marine terraces of Cabo Vírgenes (Ortlieb, this book). So, the Arroyo de Las Palmillas does not underline the edge of the caldera, but rather the limit of a more recent local tectonic movement. The important uplift of the Cerro La Reforma has the same volcanic significance than the intumescence stage, and would indicate the strong probability of a new ignimbritic eruption. In any case, the very recent age of the Reforma caldera does not allow to consider this structure as extinguished.

The most recent volcanic eruptions in the Reforma region are basaltic. They built up three scoria cones and associated lava flows: the Palmas and Punta Santa María cones, on the southern and south-eastern flanks of the caldera, and the Punta Prieta volcano to the north-east (Fig.1), which is the youngest. Different lava flows were emitted from this cone, and the last one covers a marine terrace of Middle Pliocene age (Ortlieb, 1979). These basaltic eruptions, located along N-S to NNW-SSE fractures, are not the result of a new period of activity of the caldera, but indicate that the interference of important volumes of such a basic material with

the rhyolites in the magma chamber is probable. Such a mechanism is generally considered as the cause of a rejuvenated stage in acid systems.

So as shown by field work, Sierra La Reforma is really a caldera structure, whose formation started in the Late Quaternary (≈ 1 m.y.). After a regional vertical uprise related to the intumescence stage ignimbritic eruptions occur immediately followed by basaltic emissions.

As a consequence of the partial emptying of the magma chamber a circular zone collapse forms a caldera. Following the hydrostatic re-equilibration of this central part, acid lavas erupt near the caldera rim, building up four domes. The outstanding characteristic of this caldera is yet the important uplift of the Cerro La Reforma block which also moved up the adjacent areas.

Only few K/Ar data are available (Schmidt, 1975; Schmidt et al, 1977) and they concern only the pre-caldera volcanics. These ages date the drying up of the marine gulf in ≈ 1 m.y. Age determinations on the Quaternary lavas are in progress, and it would be particularly interesting to correlate the marine terraces corresponding to the first regional rise (intumescence stage), and the last piston-like uplift of Cerro La Reforma which are the cause of the staircase wave-cut terraces well preserved on the eastern side of the caldera (Ortlieb, this volume).

PETROGRAPHY AND CHEMISTRY OF THE LAVAS

1.- Volcanic rocks from Tres Vírgenes volcano

The lavas emitted by the Tres Vírgenes volcano are porphyritic and they have a rather constant mineralogy. Plagioclase phenocrysts are more than 1 cm long; clino-pyroxene + orthopyroxene + hornblende are the ferro-magnesian minerals. The glassy matrix contains few feldespars and pyroxene microlites. In one of the viscous lava flow located on the southern flank of the cone, biotite, quartz and olivine are also present. This last mineral is rimmed by ortho-pyroxene. Such a complex association is probably the sign of the mixing of the dacitic magma filling the chamber with an uprising basaltic lava. The pumice deposits surrounding the volcano, and the more recent dacitic flow located on the southern foot of the cone have the same mineralogy and are certainly connected with the Tres Vírgenes magma chamber. Three samples were analyzed (Table 1). All are dacitic in composition and show a weak evolution from 63 to 66% of silica. Considering their mineralogy and chemistry these rocks are typically calc-alkaline dacites.

The 1746 black lava is on the contrary very different as its chemical composition is basaltic (Table 1). This rock is aphyritic and only a few small olivine phenocrysts with quenched morphology can be observed. The brown glassy matrix contains plagioclase microlites, small olivines, clino-pyroxenes, and numerous magnetite grains. Chemically this basalt (52% of silica) has a relatively high alumina content, calcium and alkali values are high also and it has also a definite calc-alkaline character. It contains 4.5% of normative olivine in the C.I.P.W. norm, but no nepheline. This basaltic lava has no obvious relationship with the Tres Vírgenes dacitic magma chamber, but represents a small batch of basic magma rising from a deeper zone.

2.- The Reforma volcanic rocks

The succession of volcanic events is complex in the Sierra La Reforma, but as a result of the volcanological study, two sequences can be distinguished: the pre-caldera sequence, and the Reforma volcanics.

2.1.- The pre-caldera sequence

The pumicitic rocks, conspicuous near the Gulf coast, contain only a few -

phenocrysts of plagioclase and greenish clinopyroxene. Magnetite, apatite, and zircon are accessory minerals, and the groundmass is glassy. A small acid lava pile outcropping in the eastern region has the same mineralogical composition, but the pumice flows interbedded in the Pliocene sediments are completely aphyric. A fragment sample in the center of a "pumicitic pillow" was analyzed, giving a rhyolitic composition (Table 2), but more detailed chemical studies on the pumice falls deposits are difficult because of their contact with sea water.

The structure of the submarine basic Pliocene lavas is rather variable as the cooling conditions change according to the size of these intrusions. In the palagonitic tuffs, the basaltic fragments have a glassy quenched structure while in the thick lava flows a finely crystallized groundmass is observed. The porphyritic intrusives stocks described by Schmidt (1975) represent in fact the largest Pliocene emissions, and this explains their microcrystalline structure, the similar petrography and the close connection with the lava flows. Olivine is not common in these rocks though relics of peridot are in some case observed included in pyroxene crystals. Plagioclase is the most abundant mineral, and is associated with orthopyroxene, clinopyroxene and small magnetite grains. Calcite is very frequent, probably as a result of the contact with Pliocene sediments. In the intrusive stocks (Schmidt, 1975), hornblende and biotite are present but generally in the groundmass. These lavas are mostly andesites, and true basalts are rare. In the two analyzed samples (Table 2) the amounts of Al_2O_3 are low and those of FeT relatively high but the chemical diagrams show a well defined calc-alkaline character.

Among the Quaternary Aro volcanics, a plagioclase rich dacitic facies is frequent. The plagioclase crystals present large glassy inclusions. Clinopyroxene, orthopyroxene and magnetite are the others phenocrysts. The matrix is glassy and contains small feldspar microlites. The chemical analyses (Table 2) confirms the silica-rich nature of these lavas. The three basic Pliocene samples show a regular evolution of the different oxides and their mineralogical association are similar. These rocks are probably the products of a common parent magma.

2.2.- The Reforma volcanics

The ignimbrites covering the flanks of the Reforma caldera are typically aphyric. The more common minerals are euhedral plagioclases and greenish clinopyroxenes, but some fayalite crystals can also be observed. The pumices from the Cuesta del Infierno pumice falls deposits have exactly the same mineralogy; they are beyond doubt part of the caldera forming eruptions. The peralkaline character of these rocks is indicated by the presence of acmite in the C.I.P.W. norm (Table 2). Using the nomenclature of MacDonald (1974), the FeO^T and Al_2O_3 values classify these rocks as comendites.

On the flanks of the caldera, the ignimbrite units have thereafter been covered by more basic flows. These lavas contain a plagioclase + clinopyroxene + orthopyroxene phenocrysts association. The same minerals are present as fine microlites in a brown glassy matrix. The analyzed rock (Table 2) has an andesitic composition, very similar to one of the Pliocene lava.

Two samples of the post-caldera domes were studied with more detail. One of them (1083) has a mineralogical association similar to that of the ignimbrite but the fayalite crystals are completely transformed. The chemical composition is almost identical, though no acmite is formed in the C.I.P.W. norm. The other lava (1094) has higher Al_2O_3 and CaO contents, and has a calc-alkaline character.

The last volcanic eruption in the Reforma region is the Punta Prieta scoria cone and its associated flows. These lavas are porphyritic and they contain numerous-

brown amphiboles which have totally reacted to form a fine grained magnetite-rich pseudomorph. Clinopyroxene is also abundant and olivine is present in the center of some pyroxene clusters. Plagioclase phenocrysts are not common, but the groundmass contains numerous fine feldspars, and pyroxene microlites, magnetite grains, brown glass and tridymite. This lava is an alkali-rich, low-silica andesite (Table 2).

From this mineralogical and chemical study of the Reforma and Tres Vírgenes volcanics, some results can be emphasized.

(1) There is no difference between the pre-caldera Pliocene rocks and the Quaternary Reforma sequence. The chemistry of the lavas is rather similar, and a bi-modal distribution of andesites and rhyolites is observed in both cases. The close geographic relation of the Pliocene products with the Reforma caldera, and the absence of such an activity in the Santa Rosalía area also demonstrate that the Pliocene events represent the first stage of the caldera development.

(2) The La Reforma ignimbritic products have a comenditic affinity, but this peralkaline character is not observed in the dome lavas or in the Pliocene pumicitic eruptions. In fact, a loss of alkalis as a result of hydrothermal alteration or hydration of the glass is frequent in this type of rocks and leads to a lowering and a loss of the agpaitic character (Araña et al, 1974). Yet one of the rhyolitic lava domes has a well-defined calc-alkaline tendency.

(3) In fact, the great majority of the Plio-quaternary Reforma and Tres Vírgenes volcanic rocks show a calc-alkaline affinity. No tholeiitic lavas were found in the Santa Rosalía area. Schmidt (1975) and Schmidt et al, (1977) have pointed out an important change in the Reforma products from Pliocene calc-alkaline rocks to Quaternary tholeiitic products, but in fact none of their chemical analysis exhibit a tholeiitic tendency. On the contrary, the latest basaltic eruptions from the Tres Vírgenes volcano and from the Punta Prieta cone show a mildly alkaline character according to the diagram of Schwarzer and Rogers (1974).

(4) The close association of basic magmas with the dacitic products of the Tres Vírgenes volcano or with the rhyolitic ignimbrites of the Reforma caldera implies the intervention of phenomena triggering an acidic explosive eruption and magma mixing is probable.

(5) Comenditic rocks normally occur in oceanic islands or continental zones of marked tensional tectonics, as oceanic islands near crests (Baker, 1974) or continental rifts like the Afar depression (Barberi et al, 1974). The La Reforma comendites are related with such a tensional tectonic, as they lie in the margin of the Gulf of California rift system, but they are also closely associated, as are the comendites of Sardinia (Araña et al, 1974) or those from La Primavera in the Trans-Mexican volcanic belt (Mahood, 1977; Demant 1978 and 1981), with a calc-alkaline sequence. In fact the volcanism of the Santa Rosalía area has to be considered within the more general framework of the magmatism related with the evolution of the Gulf of California.

DISCUSSION: TECTONICS AND MAGMATISM IN THE GULF OF CALIFORNIA AREA

About 30 m.y. ago, the East Pacific Rise reached the continental plate of North America; the Farallon plate subduction stopped and right lateral motion took place along the north-western coast of Mexico (Atwater, 1970). South of La Paz, oceanic crust appeared between 6 and 3.5 m.y. as a result of the jump of the East Pacific Rise to its present position (Mammerickx, 1979). Before Pliocene time the Gulf of California was most probably an extensional zone (Karig and Jensky, 1972; Chorowicz et al, 1980), and may be considered as a continental rift in an early stage of its history.

The 10 m.y. old basalts lying upon the Comondú Formation (Gastil et al, 1979) seem to be related to this continental opening. Studies provided evidence that the Gulf of California is now a region of lithospheric plate separation (Moore, 1973); spreading zones were recognized in several basins, and tholeiitic basalts were cored in the Guaymas Basin. Isla Tortuga, east of Santa Rosalía is a recently formed island with such a tholeiitic affinity (Batiza, 1978). Yet, on the peninsular block, tholeiitic rocks are not present, and a complex association of calc-alkaline rocks, on the Gulf side, and alkaline rocks near the Pacific side was recognized by Gastil et al, (1979). The calc-alkaline magmatism is not related with a subduction zone, and is probably atypical. Until now, no satisfactory explanation of this sequence has been provided.

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	1106	1076	58	1078
SiO ₂	65.42	64.09	62.74	52.02
TiO ₂	0.73	0.73	0.90	1.48
Al ₂ O ₃	15.07	14.35	16.07	17.77
Fe ₂ O ₃	1.73	1.05	2.01	2.77
FeO	2.42	1.84	2.49	4.45
MnO	0.07	0.07	0.07	0.12
MgO	2.36	1.54	1.96	5.26
CaO	4.36	4.12	4.45	9.83
Na ₂ O	4.36	4.82	4.92	4.39
K ₂ O	2.50	2.34	2.06	0.78
P ₂ O ₅	0.17	0.04	0.23	0.39
H ₂ O ⁺	0.44	4.54	1.00	0.10
H ₂ O ⁻	0.09	0.20	0.12	0.04
Total	99.72	99.73	99.02	99.40
C. I. P. W. Norms				
Q	18.60	18.46	14.78	-
Or	14.79	14.57	12.18	4.61
Ab	36.85	42.89	41.58	37.10
An	14.13	11.14	15.65	26.44
Di	5.21	8.14	4.07	15.98
Hy	5.34	1.65	4.51	3.01
Ol	-	-	-	4.42
Mt	2.51	1.60	2.91	4.01
Ilm	1.39	1.46	1.71	2.81
Ap	0.40	0.10	0.54	0.92

1106: viscous dacitic flow, eastern flank of the cone
 27°27'00" N 112°34'00" W

1076: pumice fall deposits from the Tres Virgenes
 27°24'30" N 112°34'45" W

58: small recent dacitic flow, southern foot of the cone
 27°24'50" N 112°35'15" W

1078: 1746 basaltic flow
 27°24'40" N 112°36'30" W

Table 1: Chemical analysis, CIPW norms, petrography and localization of Tres Virgenes volcanics.

	749	1985	1098	1082	1074	1079	1083	1094	1104
SiO ₂	72.29	56.36	58.63	66.50	70.80	59.69	70.99	65.69	55.56
TiO ₂	0.20	2.16	1.25	0.90	0.55	1.42	0.32	0.49	1.33
Al ₂ O ₃	12.78	14.97	15.61	14.65	11.75	15.06	12.42	13.57	15.97
Fe ₂ O ₃	0.26	1.27	2.53	1.34	2.35	2.20	0.67	1.78	2.96
FeO	1.12	8.22	4.87	2.90	0.97	5.07	2.10	1.07	3.81
MnO	0.01	0.13	0.12	0.07	0.10	0.12	0.01	0.06	0.12
MgO	0.08	4.66	3.74	0.99	0.17	3.30	0.07	0.51	4.26
CaO	0.54	5.37	6.45	1.94	0.90	5.66	0.57	1.91	7.32
Na ₂ O	3.77	3.99	4.18	5.47	4.95	4.64	4.90	4.22	4.55
K ₂ O	4.65	1.36	1.72	2.82	3.82	1.78	3.72	2.90	1.92
P ₂ O ₅	0.01	0.55	0.33	0.19	0.04	0.39	0.01	0.02	0.41
H ₂ O ⁺	3.97	0.14	0.68	1.59	2.98	0.75	3.08	6.18	0.94
H ₂ O ⁻	0.30	0.07	0.03	0.10	0.18	0.06	0.15	0.99	0.12
Total	99.98	99.25	100.14	99.46	99.56	100.14	99.01	99.39	99.27
C.I.P.W. norms									
Q	32.02	7.00	9.35	17.79	27.90	9.74	27.28	27.41	3.14
Qr	28.74	8.04	10.17	16.68	23.44	10.52	22.97	18.60	11.35
Ab	33.29	33.72	35.32	46.23	40.53	39.21	43.23	38.67	38.45
An	2.74	18.89	18.71	7.06	-	14.97	0.92	10.15	17.44
Ac	-	-	-	-	2.53	-	-	-	-
Di	-	3.74	9.19	1.21	-	8.82	1.75	-	13.09
Hy	1.80	20.50	9.88	4.73	-	9.29	2.17	1.38	6.99
Wo	-	-	-	-	1.32	-	-	-	-
Mt	0.39	1.84	3.66	1.94	1.93	3.19	1.01	2.41	4.29
Hem	-	-	-	-	0.23	-	-	0.27	-
Ilm	0.40	4.10	2.37	1.71	1.08	2.69	0.63	1.01	2.52
Ap	0.02	1.30	0.78	0.45	0.10	0.92	0.02	0.05	0.97

- 749: pumicitic flow in Pliocene sediments, southern rim of Arroyo de Punta Prieta
27°26'30" N 112°19'45" W
- 1085: thick lava flows in Pliocene sediments, northern rim of Arroyo de las Palmillas
27°34'30" N 112°26'15" W
- 1098: Pliocene lava flow, unnamed arroyo north of Cañon del Diablo
27°32'00" N 112°21'30" W
- 1082: aerial dacitic lava of the Aro sequence, north of Las Minatas
27°32'00" N 112°25'15" W
- 1074: ignimbritic vitrophyric level, Cuesta del Infierno
27°23'30" N 112°22'45" W
- 1079: basic lava flow overlying the ignimbrites, southern flank of the caldera
27°27'30" N 112°25'00" W
- 1083: rhyolitic dome, north-western rim of the caldera
27°27'30" N 112°25'30" W
- 1094: rhyolitic dome inside the caldera, southern region
27°29'45" N 112°24'30" W
- 1104: recent lava flow from the Punta Prieta cone
27°31'00" N 112°20'00" W

Table 2: Chemical analysis, Cipw norms, Petrography and localization of La Reforma volcanics samples.

LATE CENOZOIC VOLCANISM IN THE TRES VIRGENES AREA

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ABSTRACT

Four periods of late Cenozoic volcanism are represented by the volcanic rocks in the Tres Virgenes area, northern Baja California Sur. Neogene volcanism consists of calc-alkalic andesite that is succeeded by tholeiitic basaltic rocks. Pleistocene bimodal, basaltic-rhyolitic volcanism predates the Tres Virgenes calc-alkalic, basalt through dacite suite. At La Virgene volcano, the principal vent of the Tres Virgenes rocks, two major sequences of basalt through dacite testify to the two-stage evolution of this volcano. The Tres Virgenes vents are aligned north-south and parallel the extensional, Agujito fault zone which may be a product of northwest-southeast right-lateral shear couple.

RESUMEN

Cuatro épocas de volcanismo del Cenozoico último son representados por las rocas volcánicas en la área de las Tres Virgenes en el parte más norte de Baja California Sur. Volcanismo del Neogeno se compone de andesita calc-alcalina y el más joven basalto tholeiítico. Volcanismo basáltico-riolítico de dos modas del Pleistoceno antecede basalto a dacita calc-alcalina de las Tres Virgenes. Al volcán La Virgene, la abertura principal de las rocas Tres Virgenes, dos sucesiones de basalto a dacita testifica al evolution de esta volcán en dos series. Las aberturas de las Tres Virgenes compone una línea del norte al sur y siguen paralelo a la zona de faltas extensional de Agujito que pueden ser un producto de esfuerzo derecha-lateral en una dirección noroeste-sudeste.

INTRODUCTION

Tres Virgenes rocks compose two stratovolcanoes, La Virgene and La Soufre, and satellite vents that crop out in a north-south linear array west of las Virgenes valley (Fig. 1). This Tres Virgenes volcanism is Pleistocene and/or Holocene in age, and it was preceded by three magmatic episodes since Miocene time (Fig. 2). Oldest to youngest, these 3 episodes are represented by: (1) the early Miocene Andesite of the Sierra Santa Lucia; (2) the (?) mid to late Miocene Basalt of Esperanza; and (3) the Pleistocene La Reforma ignimbrites and basalt.

The Tres Virgenes area discussed herein is approximately 20-25 km wide and approximately 40 km in length (north-south). The eastern margin of this area is the La Reforma volcanic complex to the north and the village of Santa Agueda to the south (Fig. 1). The Boleo basin and La Reforma are thus not included.

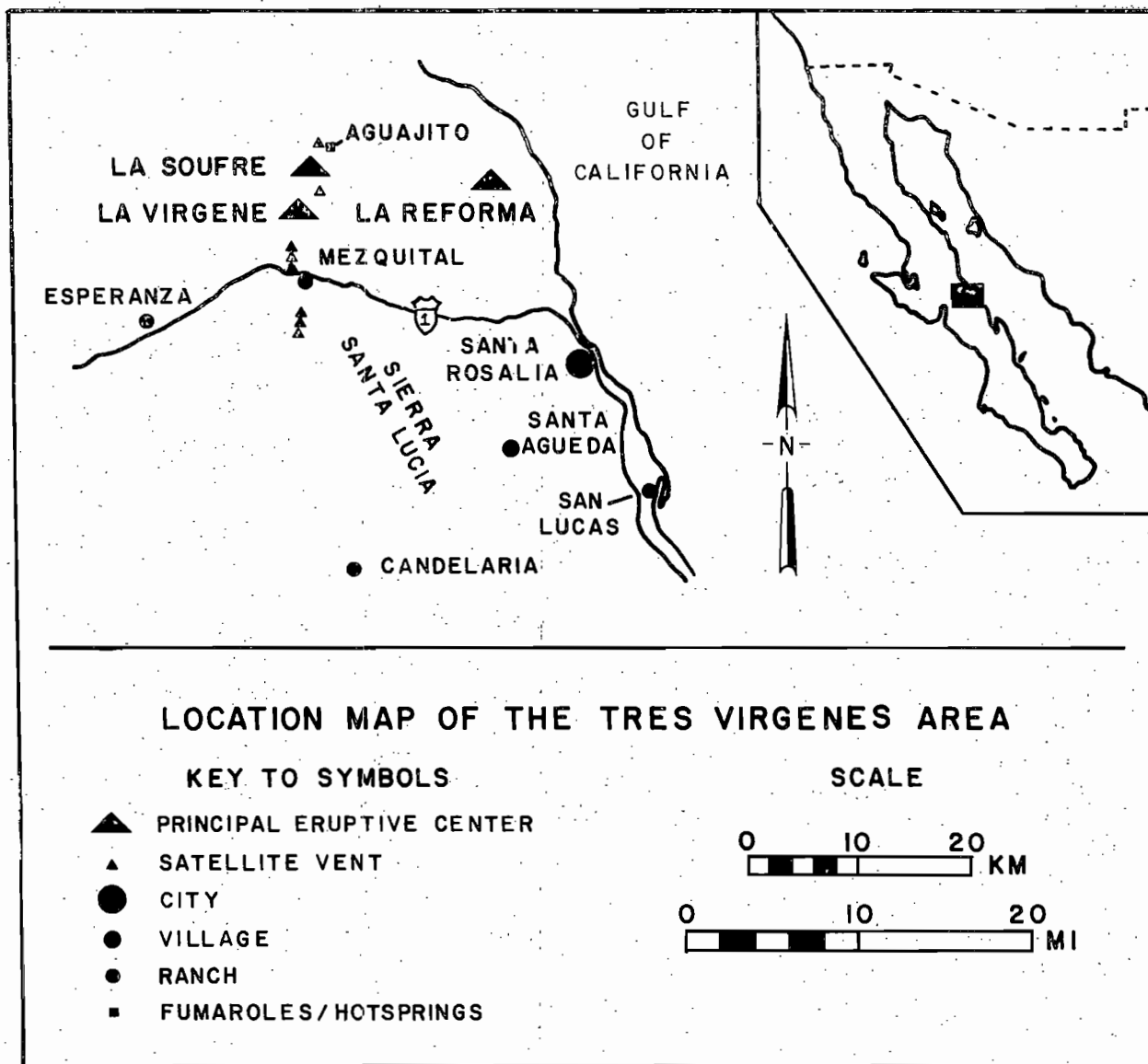


Figure 1. Location map. Rectangle in inset sketch map of Baja California represents the area of the location map. The area between La Soufre, La Reforma and highway 1 is Las Virgenes Valley.

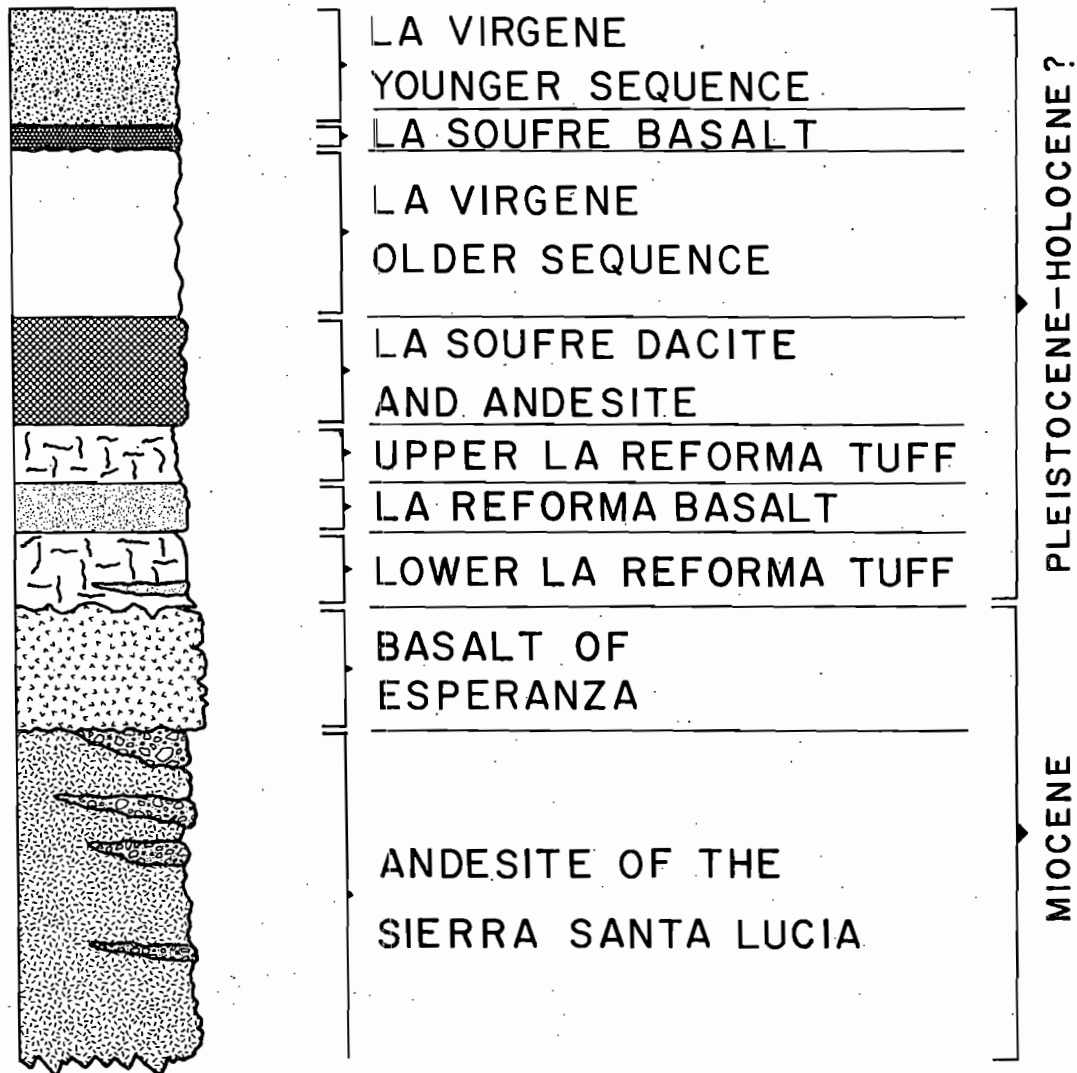
PRE-TRES VIRGENES ROCKS

Granitic rocks of Cretaceous age crop out in the area between highway 1 and La Reforma and they probably form a plutonic basement in the Tres Virgenes area. Overlying these granitic rocks is a late Cenozoic volcanic cover that includes the Andesite of the Sierra Santa Lucia, the Basalt of Esperanza, and the La Reforma volcanic rocks.

The Andesite of the Sierra Santa Lucia is exposed mostly south of highway 1 (Fig. 1). This unit comprises gently-dipping (10° or less) andesite, and mudflows and minor associated fluvial and eolian sediments that increase in abundance to the west. Hornblende is the characteristic phenocryst of the andesite. X-ray fluorescence analyses of 13 samples show that the volcanic rocks of this unit are calc-alkaline with respect to variations in FeO^*/MgO vs SiO_2 (Miyashiro, 1974), have a limited range in silica of 57-64% SiO_2 , and have $\text{K}_2\text{O} \sim 1.3$ at 60% SiO_2 (Table 1). An early Miocene age is assigned on the presumption that this unit correlates with the 22-17 m.y. hornblende andesite of Gastil and others (1979).

The ?mid to ?late Miocene Basalt of Esperanza crops out mostly north of highway 1 and west of the Tres Virgenes volcanoes in the Tres Virgenes area (Fig. 1). It consists of mesa-forming basalt flows that apparently issued from nearby, monogenetic cones. The lava flows tend to be laterally extensive in contrast to the localized, intercoalescing flows of the Andesite of the Sierra Santa Lucia. The contact between these two Miocene volcanic units is observed only in an area south of La Virgene volcano where the Basalt of Esperanza disconformably overlies the Andesite of the Sierra Santa Lucia (Figs. 3A,B; 4A,B). Petrographic studies indicate the most common rock type is basaltic andesite rather than true basalt. Typically, the rocks lack olivine, and the common phenocryst assemblage is Plag + Opx + Cpx + Fe-Ti oxides. A single chemical analysis indicates that K_2O is very low and total iron is markedly higher than in the Sierra Santa Lucia and Tres Virgenes rocks (Table 1). The chemistry is tentatively characterized as tholeiitic. It is similar to an analysis of basalt from near San Ignacio, a town about 30 km west of Esperanza, that was dated by Gastil and others (1979) at 11.1 ± 0.8 m.y.

The youngest of the pre-Tres Virgenes rocks are Pleistocene (Wilson, 1949) rhyolite ignimbrite (ash-flow tuff) and basalt that erupted from the La Reforma complex and peripheral vents, respectively (Fig. 2). La Reforma tuffs consist of two ignimbrite sheets exposed mostly north of highway 1 (Fig. 1) and east of the Tres Virgenes (Fig. 3A,B). In addition, remnants of intracanyon flows of the older ash-flow crop out in arroyo Santa Agueda, 4-8 km southwest of Santa Agueda, and in arroyos west of the Tres Virgenes. This ash-flow spread across las Virgenes valley, and in one instance, surmounted a ridge with a minimum height of 100 m., and flowed down the east-west canyon 0.8 km south of Mezquital, nearly reaching Esperanza. The lower tuff is characteristically partially welded, with large pumice lumps or *fiámme* (to 40 cm) common. Pumice airfall, locally overlain by cross-bedded surge deposits, is present beneath the lower tuff along highway 1 at Infierno grade (see locality description for 79BMS16B, Table 1). Fayalite, ferroaugite, anorthoclase and Fe-Ti oxides are the characteristic phenocrysts in the lower ignimbrite. Chemically, the rock is a comendite (Table 1).



GENERALIZED STRATIGRAPHIC COLUMN OF THE TRES VIRGENES AREA

Figure 2. Composite stratigraphic column. Thicknesses of units shown here are schematic only and are not to scale.

TABLE 1. Representative Major Element Analyses of
Volcanic Rocks in the Tres Virgenes Area

Sample No.	79BMS67	79BMS93	79BMS16B	79BMS103	79BMS88	79BMS104
Major elements (wt. %) (anhydrous)						
SiO ₂	58.95	54.33	74.33	52.20	59.35	65.85
TiO ₂	0.76	1.43	0.23	1.32	1.02	0.74
Al ₂ O ₃	18.17	15.20	12.29	17.50	17.09	16.02
Fe ₂ O ₃ ^a	2.27	2.93	1.75	2.82	2.54	2.25
FeO*	3.46	7.12	0.91	4.30	3.53	1.94
MnO	0.10	0.14	0.06	0.12	0.11	0.08
MgO	3.63	6.33	0.35	6.11	3.86	1.66
CaO	6.62	8.96	0.70	10.01	6.70	4.10
Na ₂ O	4.38	3.31	5.39	4.58	4.40	5.17
K ₂ O	1.48	0.15	3.96	0.72	1.21	1.99
P ₂ O ₅	0.19	0.11	0.03	0.32	0.20	0.19
L.O.I. ^b	0.98	0.0	n.d.	0.02	0.61	0.90

Sample descriptions and localities:

79BMS67 - Andesite of the Sierra Santa Lucia; olivine and hornblende-bearing two pyroxene andesite from base of lava flow in ridge outcrop on southeast side of arroyo Santa Agueda, approximately 4km southwest of Santa Agueda, 1.7km northeast of Rancho El Bule (elevation: 440m).

79BMS93 - Basalt of Esperanza; hypersthene basaltic andesite from outcrop at intersection of spur from highway 1 and the old highway (dirt), 1.1km north of highway 1, 2.5km east of Esperanza.

79BMS16B - Lower La Reforma tuff; fayalite ferroaugite rhyolitic ignimbrite (devitrified) from roadcut along highway 1 at the crest of Infierno grade, 0.2km east of the Almejas microondas station.

79BMS103 - La Virgene basalt of younger sequence; basalt from roadcut along highway 1, 0.7km west of Mezquital.

79BMS88 - La Virgene andesite of younger sequence; olivine - two pyroxene andesite from outcrop of the elongate lava flow southeast of La Virgene adjacent to the base of the cone.

79BMS104 - Satellite vent dacite; two pyroxene - hornblende dacite from outcrop of coulee 100m north of Mezquital.

^a Fe₂O₃ was calculated according to the method of Irvine and Baragar (1971)

^b L.O.I. = loss on ignition produced by fusion of sample.

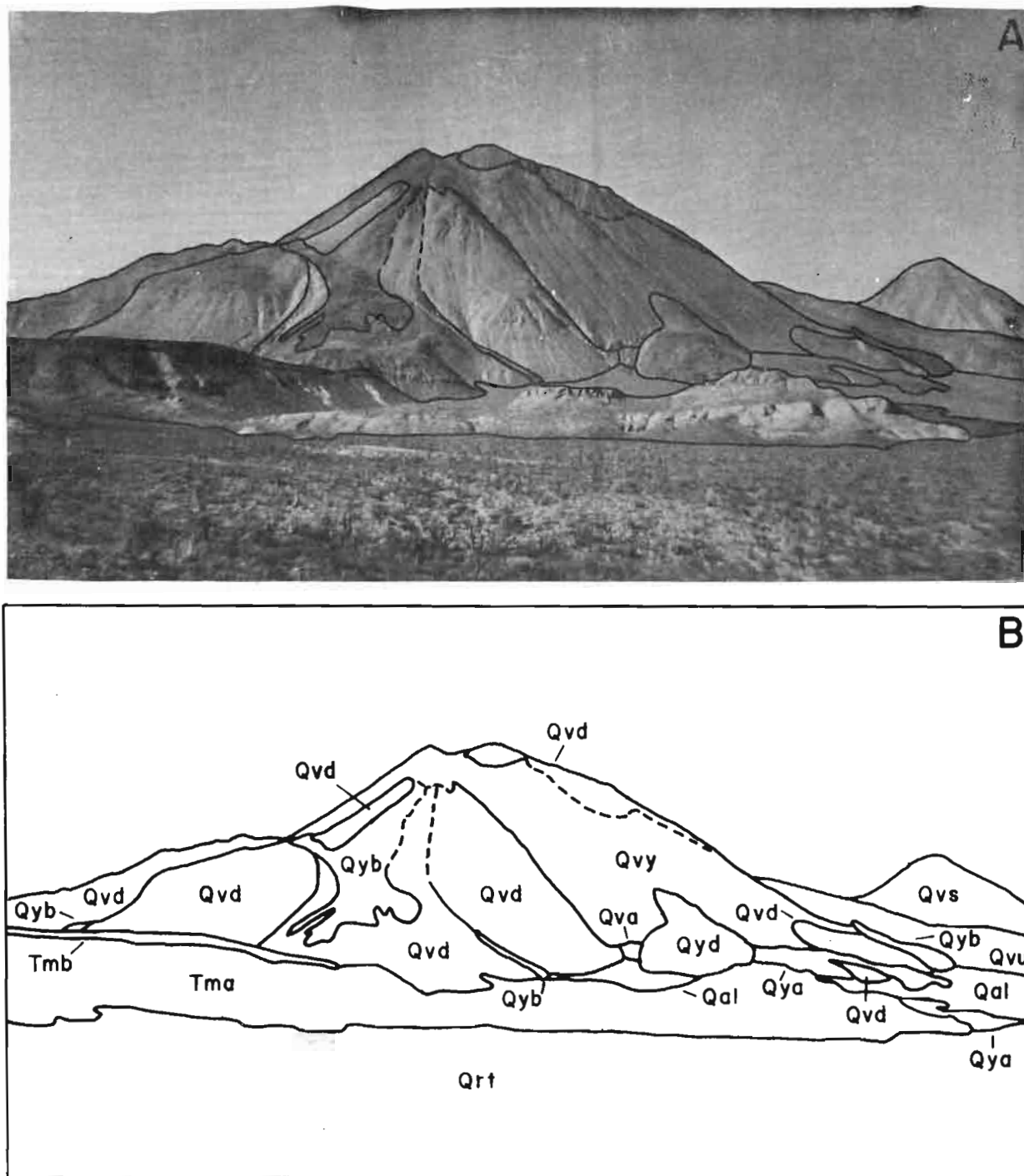


Figure 3A, B. View of the southeast flanks of La Virgene volcano (center) and La Soufre volcano (right) from highway 1 near Rancho Las Virgenes. Units are as in Figure 5 with the following exceptions. Tmv of Figure 5 is differentiated into Tma - Andesite of the Sierra Santa Lucia and Tmb - Basalt of Esperanza. Qvy is differentiated into Qyb - La Virgene younger sequence basalt, Qya - La Virgene younger sequence andesite, and Qyd - La Virgene younger sequence dacite. An additional symbol is Qvu - undifferentiated La Virgene volcanic rocks.

(McDonald, 1974).

The upper tuff is separated from the lower by basalt flows that erupted from peripheral vents, one of which is located in the eastern portion of the mapped area. In contrast to the lower tuff, it is poorly welded throughout and contains generally smaller (1-4 cm) pumice lumps. The phenocryst assemblage of the upper tuff, hypersthene, plagioclase and Fe-Ti oxides, is also quite distinct from that of the lower tuff.

TRES VIRGENES ROCKS

Nearly the entire volume of the Tres Virgenes rocks was erupted from the four principal vents shown in Figure 5. Two occur in the summit area of La Virgene, one coincides with the peak of La Soufre, and the fourth lies to the northeast of La Virgene. The northern La Virgene vent comprises dissected feeder dikes within hydrothermally-altered pyroclastic deposits. The southern La Virgene vent is a crater that marks the vent for the youngest La Virgene rocks. Although La Soufre peak was not visited, the position of the La Soufre vent is inferred from the attitudes of the fragmental deposits that dip radially away from the summit. The third "virgin", the northernmost peak of the Tres Virgenes, is not a composite cone as are La Virgene and La Soufre. It is instead a thick accumulation of dacite apparently erupted as flows from La Soufre volcano. The northeast vent (Fig. 5), the most productive of the satellite vents, extruded andesite flows that were accompanied by little, if any, pyroclastic activity. The eruptions from the north-south array of satellite vents were volumetrically minor, and with the exception of the southernmost vent, which produced basaltic andesite, all formed either dacite coulees or domes.

Eruptions at La Virgene volcano occurred in two stages represented by basalt through dacite sequences. The locus of activity of the first cycle was the northern La Virgene vent. Basalts of this older stage are exposed in two distal flows northwest and northeast of La Virgene (Fig. 5). Older cycle andesites compose much of the northwest flank though they are distributed around all sides of the volcano. The thick dacite coulees on the south side of La Virgene (Fig. 4A,B) are the ultimate products of the first cycle of activity. The present height of the cone was achieved early in the history of La Virgene, before extrusion of the older dacites, on the basis that older andesites can be traced nearly to the older vent. In addition, a thick, dacitic, pumice airfall, distributed in a lobe southwest of La Virgene, was deposited as a product of a violent eruption that postdated the older andesites but predated some of the dacite flows of the older sequence.

The extrusion of La Soufre basalt (Fig. 5), which contains cumulate gabbroic inclusions, initiated the second cycle of activity. This eruption was followed almost immediately by extrusion of basaltic flows at La Virgene that overlie La Soufre basalt and that drape over the earlier-formed dacite coulees (Figs. 3A,B; 4A,B). The limits of the youngest andesite and dacite are outlined in Figure 4A,B.

Thin section examination of the La Virgene rocks shows that phenocryst phases vary regularly with rock type. Typical assemblages, in decreasing

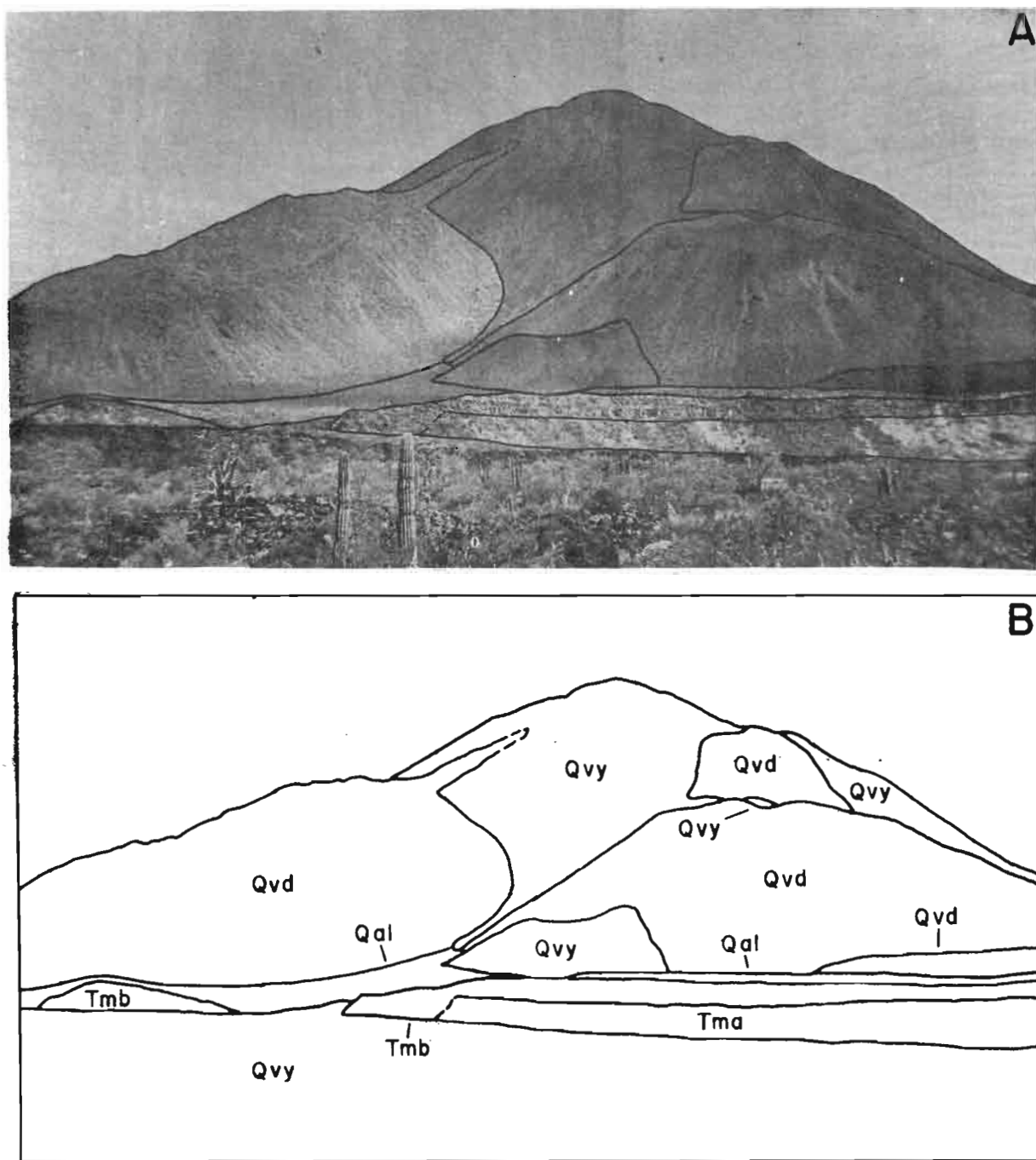
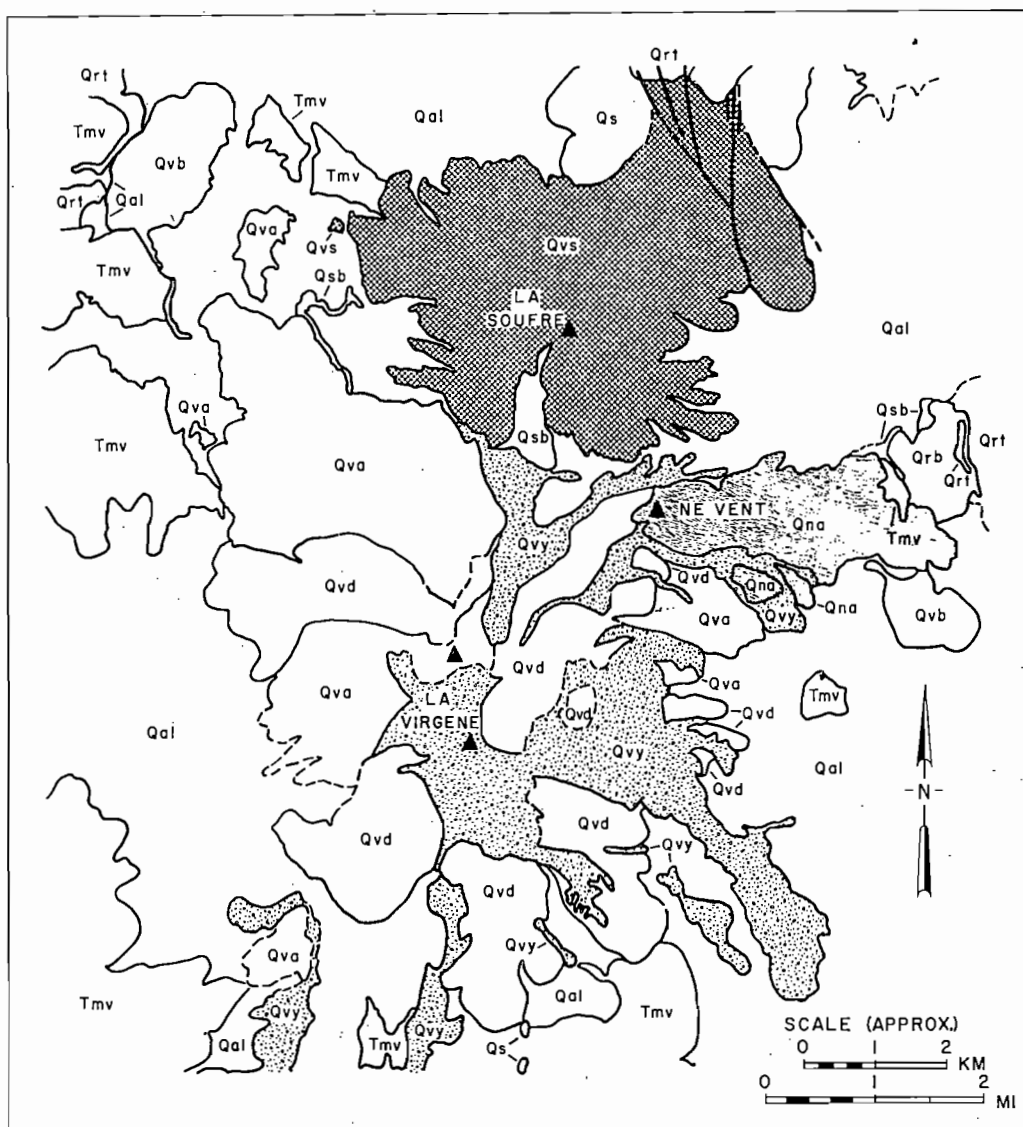


Figure 4A, B. View of the south flank of La Virgene volcano from highway 1 approximately 3km west of Mezquital. Units are the same as in Figure 5 with the following exception. Tmy of Figure 5 is differentiated into Tma - Andesite of the Sierra Santa Lucia and Tmb - Basalt of Esperanza. Qal also includes primary and reworked pumice airfall deposits.



GENERALIZED GEOLOGIC MAP OF LA VIRGENE AND LA SOUFRE VOLCANOES

GEOLOGY BY MICHAEL G. SAWLAN

EXPLANATION

Qs	SATELLITE VENT FLOWS AND DOMES (UNDIFF.)	Qvd	OLDER SEQUENCE LA VIRGENE DACITE	Qrb Qrt	LA REFORMA BASALT AND LA REFORMA TUFF
Qna	NORTHEAST VENT ANDESITE	Qva	OLDER SEQUENCE LA VIRGENE ANDESITE	Tmv	BASALT OF ESPERANZA AND ANDESITE OF THE SIERRA STA. LUCIA (UNDIFF.)
Qvy	YOUNGER SEQUENCE LA VIRGENE FLOWS (UNDIFFERENTIATED)	Qvb	OLDER SEQUENCE LA VIRGENE BASALT	Qal	ALLUVIUM
Qsb	LA SOUFRE BASALT	Qvs	LA SOUFRE DACITE AND ANDESITE		

Figure 5. Geologic map prepared using a 1:50,000 CETENAL air photo enlarged 2x for a base map. Solid triangles denote main vents.

order of abundances of phases, are: basalt, Ol + Cpx + Plag; basaltic andesite, Plag + Cpx + Opx + Ol; andesite, Plag + Opx + Cpx + Hb + Qtz; and dacite, Plag + Hb + Opx + Cpx + Bt + Qtz. Eight samples analyzed from La Virgene volcano are calc-alkaline on the basis of FeO^*/MgO vs SiO_2 variations. SiO_2 ranges from 52-68%, and K_2O is about 1.3 at 60% SiO_2 . Analyses of the La Virgene andesite and the Miocene Andesite of the Sierra Santa Lucia are virtually identical (Table 1).

STRUCTURAL SETTING OF THE TRES VIRGENES AREA

The trends of faults and folds within the area mapped suggest that these features may be a product of a northwest-southeast directed, right lateral shear couple. The Tres Virgenes vents are aligned in a north-south array along the western margin of the Aguajito fault zone. This zone is defined as the 1-2 km wide north-trending, multiple horst and graben sets that occur from Aguajito in the north to 6-8 km east of Mezquital. Vertical displacements are small, approximately 1-10 m. Axes of the folds within fault blocks trend at large angles to the northwest.

RECENT ACTIVITY OF TRES VIRGENES VOLCANOES

The Tres Virgenes volcanoes were cited as historically active in 1746 and 1857 (Wilson, 1949, 1955; Mooser and Lagos, 1961). The reported activity in 1746 is based on a Spanish missionary's labelling of the Tres Virgenes as "volcanoes of fire" on his sketch map of the region bordering the Gulf of California (Ives, 1962). Mooser and Lagos (1961) presented unreferenced reports of "clouds of smoke and vapor" that apparently were observed in 1857, but they doubted that an eruption had occurred. Inspection of the present La Virgene crater, clearly the source of the youngest La Virgene lavas, showed it to be in a state of degradation, partially infilled with colluvium and lacking recent pyroclastic debris. This author's qualitative appraisal of the crater suggests that there has not been magmatic activity in the past several hundred years although small phreatic eruptions cannot be ruled out.

ACKNOWLEDGEMENTS

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METAMORPHIC CORE COMPLEXES IN NORTHERN SONORA
AND SOUTHERN ARIZONA

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STRUCTURAL AND STRUCTURAL-PETROLOGICAL CHARACTERISTICS
OF SOME METAMORPHIC CORE COMPLEX TERRANES
IN SOUTHERN ARIZONA AND NORTHERN SONORA

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INTRODUCTION

Cordilleran metamorphic core complexes in southern Arizona and northern Sonora owe their character to disharmonic structural relationships. Four distinctly different structural levels produce the disharmony: these are occupied by protolith, tectonite, décollement zone, and detachments (Davis, 1980b) (Fig. 1).

The lowermost level is country-rock protolith. The protolith may consist of any rock which falls within the age-range of Precambrian through (at least) early Tertiary. Structures in rocks of the protolith conform to whatever habit typifies rocks of the region. Upward, the country-rock protolith is progressively or discontinuously transformed into tectonites, commonly mylonitic, characterized by low-dipping foliation and penetrative unidirectional lineation (Fig. 1). Minerals, textures, and structures of the protolith are overprinted thoroughly by the tectonite fabrics. Granites and quartz monzonites are transformed into mylonitic gneisses (Fig. 2A); sedimentary rocks are converted to carapace-like assemblages of tectonite marble, quartzite, calc-silicate rock, and flattened-pebble conglomerate (Fig. 2B); and low-grade metasedimentary and metavolcanic schists are deformed into tectonite schists with superposed, foliate fabrics (Fig. 2C). The tectonite foliation and lineation indiscriminately overprint all rocks and structural components of the protolith assemblage.

The next level upward is distinguished by intensive microbrecciation and chlorite-iron-epidote alteration of mylonitic tectonite within relatively thin décollement zones (Figs. 1 and 2). The décollement zones thus consist of microbrecciated and altered mylonitic tectonite. Commonly the original subhorizontal foliation and lineation is rotated to moderately or steeply dipping attitudes. The top of the décollement zone is usually capped by a massive one to two-meter thick ledge of fine-grained microbreccia, the upper surface of which is the décollement surface (Figs. 1 and 2).

Above the décollement are detachments of allochthonous non-tectonite country rock. As used here the term "detachment" refers to allochthonous, commonly rotated, hanging-wall rocks resting in very low-angle fault contact on

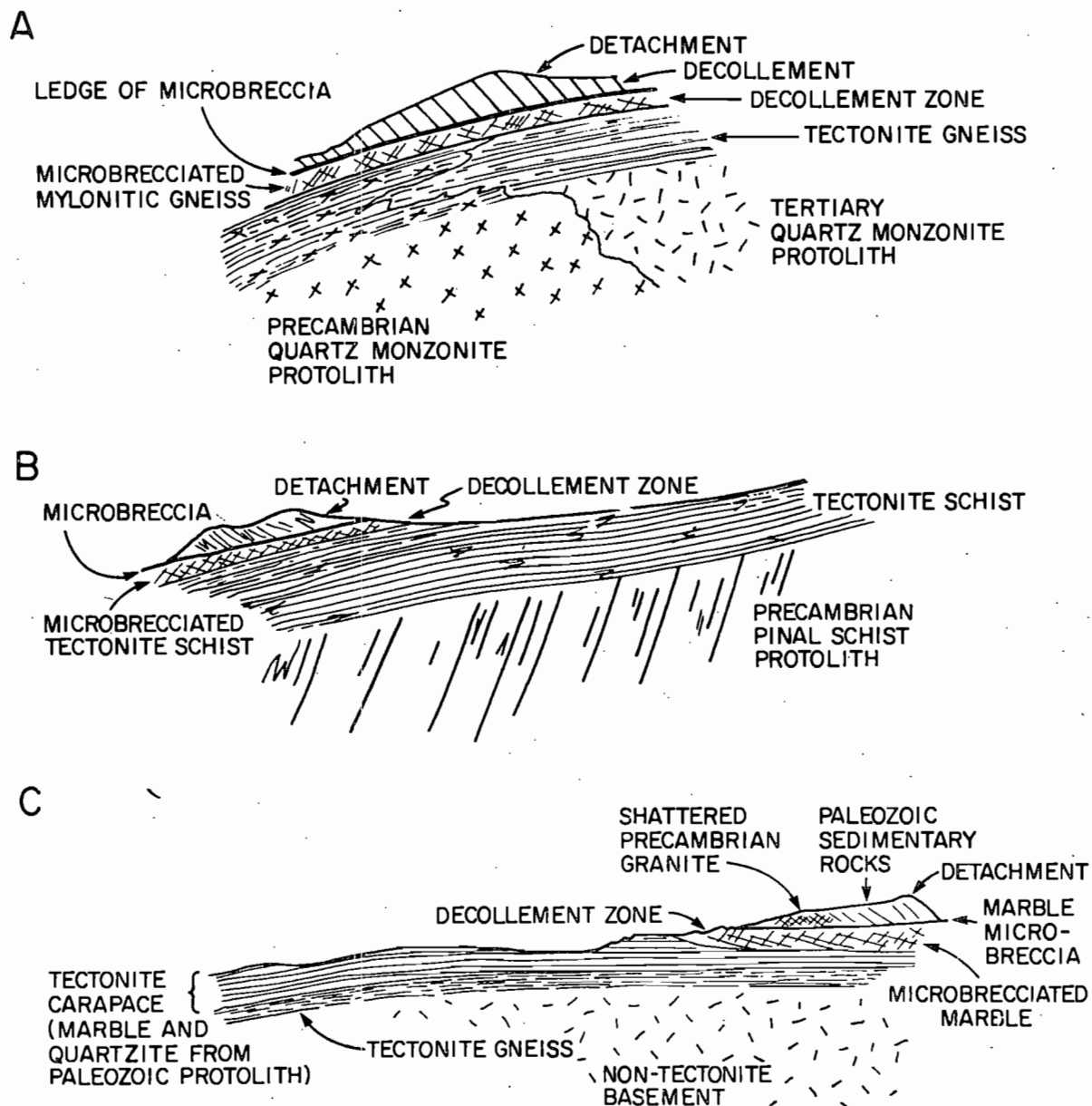


Figure 1. Structural components of metamorphic core complexes.

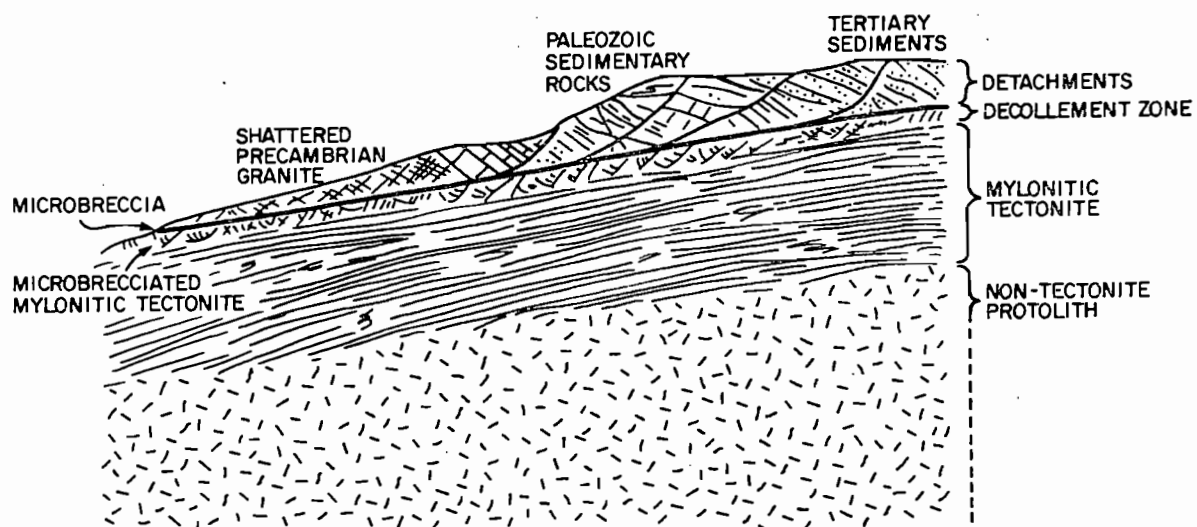


Figure 2. Expression of structural components of metamorphic core complexes in varied geologic settings.

the structurally deeper, generally older, footwall rocks. The term detachment is extracted from the expression "detachment fault", introduced by G.A. Davis and others (1979, 1980) to describe low-angle faults which mark the base of mid-Miocene allochthons in the Whipple-Buckskin-Rawhide Mountains terrane in southeastern California and western Arizona. The detachment rocks clearly have been faulted from their original underpinnings. They may be of any lithology and may range in age from Precambrian to mid-Miocene. Structural deformation in the detachment rocks is diverse, including faulting and rotation and/or shattering and/or folding.

Given the unusual nature of the disharmonic relationships, it should be no surprise that the origin of the Cordilleran metamorphic core complexes is hotly debated. Central issues are (1) time(s) of formation of mylonitic tectonite, (2) dynamic conditions under which the mylonitic tectonite evolved, (3) tectonic significance of décollement, and (4) mode of emplacement of detachments. Pertinent descriptive and interpretive details are available in a number of published articles (Coney, 1974, 1979, 1980; Crittenden and others, 1980; Davis, 1975, 1977, 1980a,b; Davis and Coney, 1979, 1980; Davis and others, 1979, 1980; DeWitt, 1980; Drewes, 1978, 1979; Rehrig and Reynolds, 1980; Reynolds and Rehrig, 1980; and Thorman, 1977).

The purpose of this paper is to try to bring into sharper focus the structural and structural-petrological properties of the disharmonic components of the core complexes. We attempt this through presentation of details of some core-complex study areas in northern Sonora and southern Arizona (Fig. 3): Sierra Mazatan, Sierra Magdalena, Pozo Verde Mountain, Coyote Mountains, Tanque Verde Mountains, and the Rincon Mountains. The descriptions are preceded by a section on "Interpretive Remarks" so that the chief components of core complexes (protolith, mylonitic tectonite, décollement zones, and detachments may be viewed in a broader, tectonic context. The interpretive remarks are drawn largely from Davis (1980a).

INTERPRETIVE REMARKS

The effect of major plutonic invasion into the upper crust in the early Tertiary set the stage for the unusual style of penetrative deformation recorded in the metamorphic core complex terranes. There are few localities where metamorphic core complexes lack any sign of known or inferred Tertiary intrusive bodies. The thermal input, in effect, softened the country rocks, more in some places than in others. The rheology of the cooling plutons and the locally heated country rocks made the systems vulnerable to ductile flow. The effect of intrusion was to convert the regional country rock into domains of cool rigid rocks and contrasting domains of hot ductile rocks.

Thermal effects were superimposed on a mechanically anisotropic crust. The upper crustal rocks did not possess laterally continuous lithotectonic units. Instead, the combination of Jurassic and Laramide faulting in the southern Basin and Range placed Precambrian basement laterally against Paleozoic and/or Mesozoic layered strata; folded, thrust strata against domains of homoclinal strata; and complete sections of Precambrian-Paleozoic-Mesozoic rocks against section where Cretaceous strata rest nonconformably on Precambrian basement (Davis, 1979). In fact, parts of southern Arizona and northern Sonora even seem to lack Precambrian rock altogether (Anderson and others, 1980). The combined influences of thermal and mechanical anisotropy seem to have predetermined the location of individual core-complex terranes.

Given the thermal and mechanical condition of the upper crustal

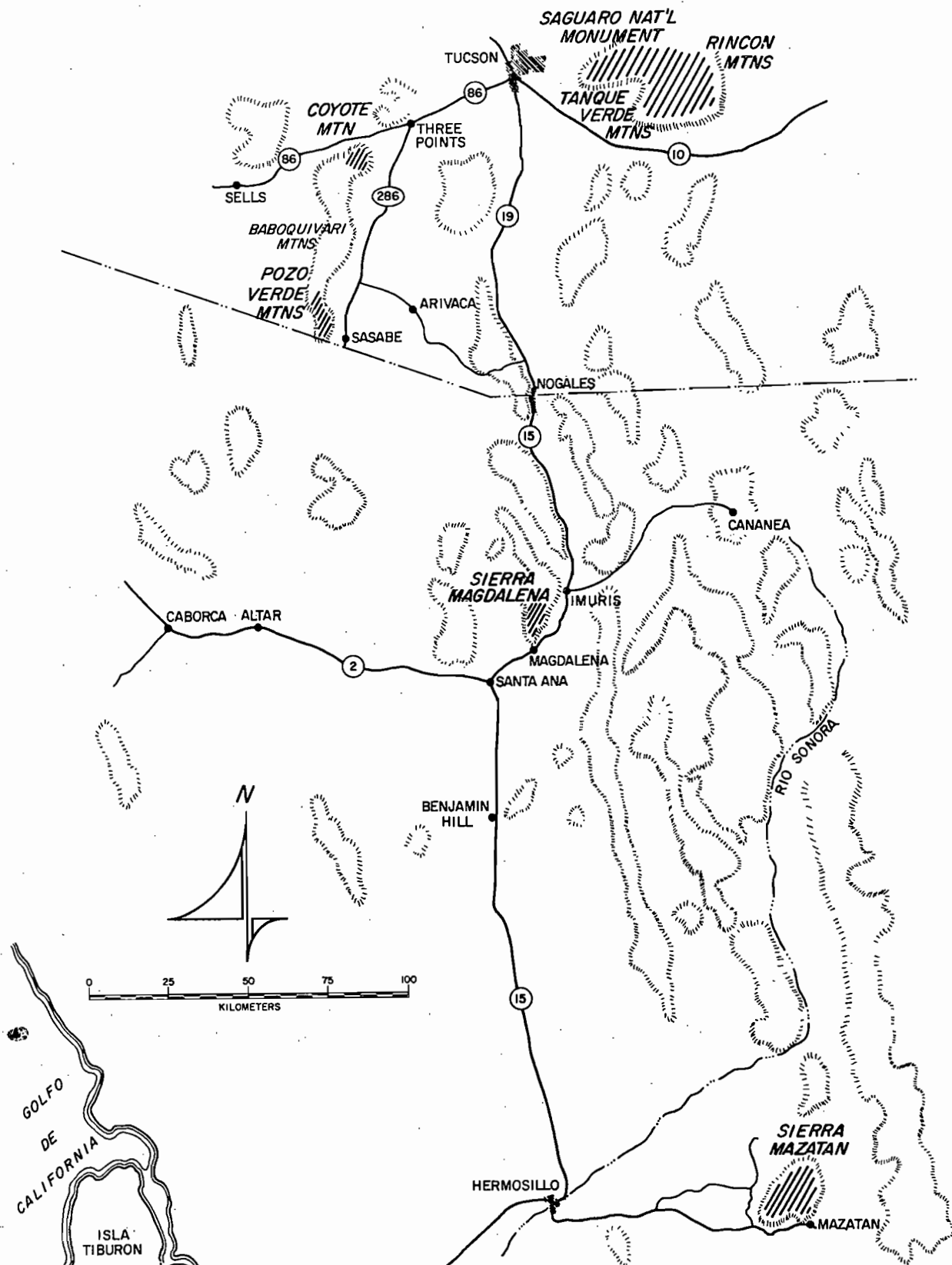


Figure 3. Map showing core-complex study areas in northern Sonora and southern Arizona.

rocks, what regional tectonic movements gave rise to the penetrative componental movements which produced mylonitic tectonite? The mylonitic tectonite fabric is considered to be a flattening/extension fabric, with direction of extension parallel to penetrative mineral lineation (Davis, 1977, 1980a; Davis and Coney, 1979; Davis and others, 1975; Coney, 1979, 1980). The tectonite fabric is interpreted to have formed by ductile normal shearing within gently dipping curvilinear zones of simple-shear (Davis, 1980b, 1981). These zones reflect regional tectonic movements which stretched the upper-crustal rocks, producing major extension (Davis and Coney, 1979). The cause of the stretching is not known. It could be a "passive" product of regional strike-slip movements. It could be a result of a collapse of what had been a shallow-dipping Benioff zone at the culmination of major Tertiary plutonism (Coney, 1979, 1980; Davis and Coney, 1979). Any interpretive theory must explain the consistent trends of fabric elements across more than 100,000 km² in rocks as young as 55 m.y. (Anderson and others, 1980) and maybe younger (Reynolds and Rehrig, 1980).

As a response to stretching, the heterogeneous upper crust was partitioned into an array of rigid blocks with ductile margins. "Megaboudin"-like rigid crustal blocks became marked by shoulders and necks of normal shearing characterized by significant thinning and flattening (Davis and Coney, 1979). At depth, these zones of normal shearing formed mylonitic tectonite at the expense of Precambrian basement. At higher structural levels these zones converted Paleozoic and Mesozoic strata into tectonite. The projection of the shear zones to the surface may have been marked by distributed, imbricate normal faulting, producing growth-fault basins of sediment accumulation (Davis and Coney, 1979). Downward, the zones of tectonite may have been transitional into horizontal regimes of lamellar flow.

As cooling of the system took place, some already-formed mylonitic tectonite was disrupted, rotated, and microbrecciated by closely spaced, mesoscopically penetrative, normal-slip listric (?) faults. This process could have been an upper-level, brittle counterpart of still-active mylonite tectonite formation at greater depth. The kinematic coordination is perfect. While mylonitic tectonite and microbrecciated mylonitic tectonite were being formed at depth, the rocks which now occupy detachments began to move downward and laterally, and thus rocks once far removed from deep normal shear zones were translated ultimately to positions directly above the mylonitic tectonite (Davis, 1977). This tectonic denudation began to fashion the dramatic disharmony between rocks of the mylonitic tectonite and the rocks above.

Abrupt cooling of the mylonitic tectonite rocks is recorded in mid-Tertiary K-Ar and fission-track ages (about 25 m.y.) (see Crittenden and others, 1980). Most workers have ascribed this cooling event to "uplift". The structural implications of such uplift are not clear. Perhaps the 25 m.y. $\pm$ cooling ages record development of the décollement and its ledge of microbreccia as a result of major imbricate normal faulting. The normal faults coalesced at or near the uppermost level of mylonitic tectonite. The trigger for this denudational faulting may be resurgence (or continuation) of faulting, especially near the boundaries of megaboudin-like tectonic units. The faulting accommodated further extensional strain in the upper crust as a whole. Ignimbritic volcanics (25 m.y. $\pm$) exploded out of deeply rifted parts of the crust throughout the entire belt of metamorphic core complexes (Davis and Coney, 1979).

It is important to consider that all of the events which proceeded from about 55 m.y. (?) to 25 m.y. may have accompanied and/or may have been produced by the concurrent formation of mylonitic tectonite (at some depth). However, the final emplacement of detachments in post-25 m.y. time, largely mid-

Miocene, postdated formation of all exposed tectonite. Oddly enough, the slip-line direction for translation of the detachments identically parallels the trend of penetrative lineation in nearby or underlying tectonite (Davis and Hardy, 1981).

The final emplacement of the detachments in mid-Miocene time probably accompanied continued regional extension. The faulting may in part have been influenced by geometric and/or dynamic conditions produced during the earlier inferred normal-slip descent of hanging-wall terranes above zones of mylonitic tectonite. The pure mid-Miocene detachments, i.e., those containing only mid-Miocene strata, were translated perfectly parallel to the direction of the penetrative lineation in the closest mylonitic tectonite, and for southeastern Arizona examples apparently (?) in the same sense of slip as that of simple-shear in the tectonite (Davis, 1981). The structural profiles of the mid-Miocene detachments convey the properties of slump- or toreada-block tectonics. Tempting as it is to explain the fault profiles as the result of stretching of brittle multilayers atop a ductile stretching medium, the geologic relationships disclose that many of the fault-block detachments now rest on footwall rocks which show no sign of concomitant penetrative flow movements: differential translation took place along low-dipping fault contacts between rigid but not necessarily stationary, basement and overlying sedimentary and volcanic rocks. Rocks nearest the margins of boudin-like crustal blocks were translated along curved (?) normal faults in the direction of the earlier stretched and descended terranes. All of these events appear to have postdated Laramide compression-induced shortening and preceded classical Basin and Range high-angle normal faulting.

SIERRA MAZATAN

Introductory Comments.

Sierra Mazatan lies 70 km east of Hermosillo, the capital city of Sonora (Fig. 3). It is a roughly circular range with a diameter of 12 km; the crest of the range is marked by knobs of granite that rise a few tens of metres above a rolling surface of low relief which in places is dissected by shallow stream valleys. As seen on aerial and orbiter photographs, Sierra Mazatan is one of the most distinctive and easily recognized physiographic features in northwestern Mexico. It rises nearly 1,000 m above the surrounding terrane and its circular flanks which outline the crest are emphasized by a cover of dark oak trees above 600 m elevation, in contrast to the lighter shades of brown and yellow of the adjacent desert.

Sierra Mazatan exhibits a strikingly smooth profile that is broken along the western flank by jutting outcrops of carbonate beds of Paleozoic age. These beds are structurally the highest elements of the range. They rest upon a strongly lineated surface developed upon a thin layer of mylonitic gneiss which is transitional downward into somewhat less strongly sheared porphyritic granodiorite gneiss. U-Pb isotopic analyses on zircon from the granodiorite gneiss yield an interpreted age of 58 ± 3 m.y. (Anderson and others, 1980). This Tertiary granite records the southernmost known extent of Tertiary deformation characteristic of more northerly Cordilleran metamorphic core complexes.

Protolith.

the major rock elements most affected by deformation. The extent of the gneissic fabric is restricted largely to the west flank of Sierra Mazatan, whereas the rocks on the top and eastern flank of the range contain scant evidence of the strong Tertiary overprint.

Tertiary granitic protolith for the gneiss is composed of rather uniform porphyritic granodiorite which is locally transitional into a more leucocratic muscovitic phase. Along the west flank of Sierra Mazatan the porphyritic granodiorite becomes gneissic and interfingers with layered quartzofeldspathic, biotite-rich gneiss and locally with quartzitic gneiss. These heterogeneous units are interpreted to be country rock into which the granodiorite was emplaced. The protolith of these gneisses is unknown. The gneisses do not contain abundant marble, quartzite, or graphitic schist and therefore it seems unlikely that they were derived from the Paleozoic carbonates or the coal-bearing Triassic beds which are found in this part of northern Sonora.

Sierra Mazatan lies near the southern extreme of a region dominantly underlain by granitic rocks. The granitic rocks appear to form a batholithic terrane which underlies a 100+ km x 25+ km region marked by north-trending ranges. The batholith is approximately bisected by Rio Sonora, and the walls of its canyon reveal excellent exposures of the granite. Granitic rocks along the flanks of this batholith record weak to moderate foliation.

Mylonitic Tectonite.

Mylonitic rocks are well exposed along the west flank of Sierra Mazatan particularly where intermittent streams have incised arroyos. The mylonitic granodiorite gneiss is characterized by classic mortar texture in which porphyroclasts of feldspar are set in a fine-grained aggregate of quartz. The porphyroclasts are from 1 mm to a few centimeters in diameter and display undulose extinction, deformation twinning, and abundant fractures. The nature of the porphyroclasts indicates that they are pre-tectonic and have persisted as relicts from primary igneous texture. Fractures that traverse the larger crystals are filled with granulated feldspar, and foliation defined by quartz aggregates wraps around the large feldspars. Quartz is abundant and its grain size, texture, and degree of strain are commonly a function of proximity to grain boundaries. Biotite, commonly partially altered to chlorite, and rare muscovite occur as (1) fine-grained aggregates that wrap around coarser feldspar crystals; (2) bent and twisted flakes where preserved from the effects of high strain; and (3) lenticular and ribbonlike aggregates of fine-grained, parallel flakes and prisms distributed among the well-oriented, strongly granulated recrystallized quartz that forms the matrix. Less abundant constituents such as sphene, apatite, zircon, and opaque minerals are commonly associated with the biotite. Of these, the opaque minerals in places show distinct granulation, whereas other minor components surrounded by more easily deformed mica seem to have escaped much of the deformation. The quartzofeldspathic gneisses are generally characterized by low-dipping foliation and penetrative lineation which commonly plunges less than 20°. Impressive swarms of aplitic and pegmatitic veins locally cut the layered country rock, and although they are not transposed into concordance with compositional layering, they are consistently slightly foliated. Locally migmatitic gneisses record extensive flowage (Fig. 4), spectacular boudinage (Fig. 5), and folds and faults. Folded layers of lineated mylonitic rock are not uncommon. Foliation and lineation become less distinct as the top of the range is approached. The rock is cut by striated surfaces, but it is not deformed to any unusual degree.

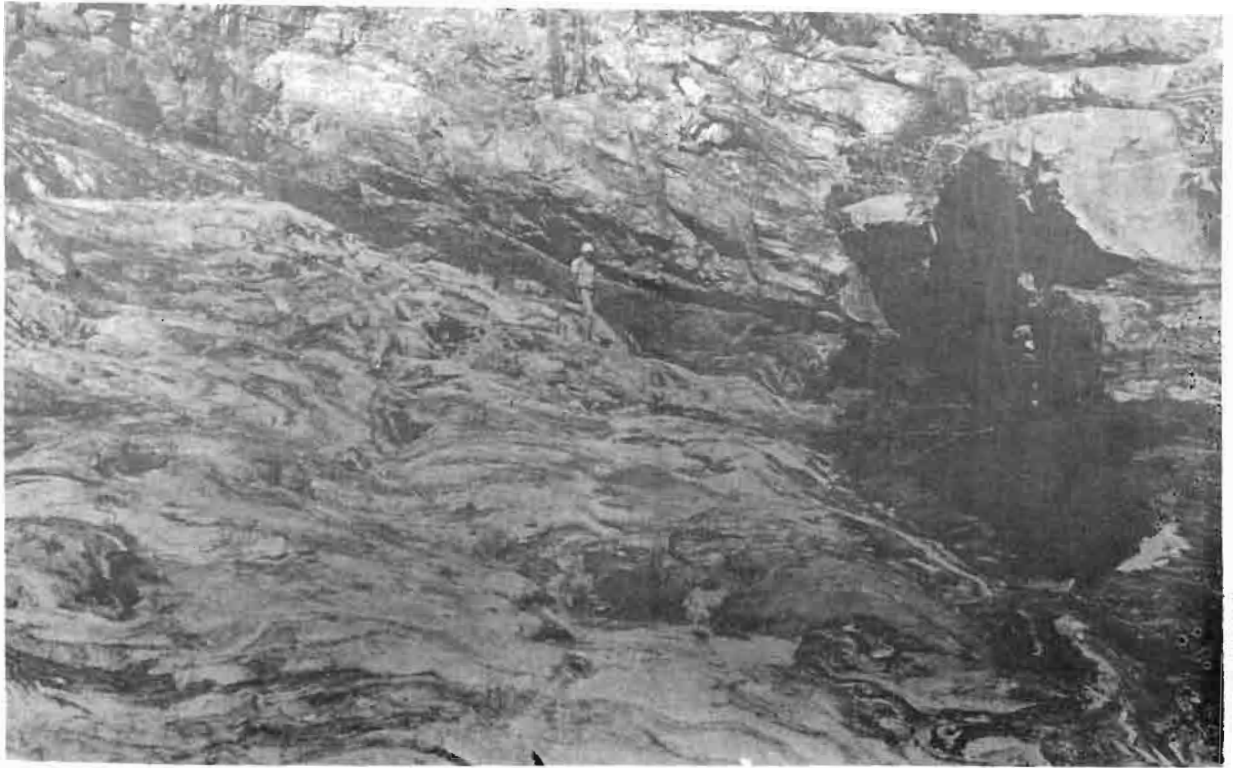


Figure 4. Migmatitic gneisses in Sierra Mazatan.

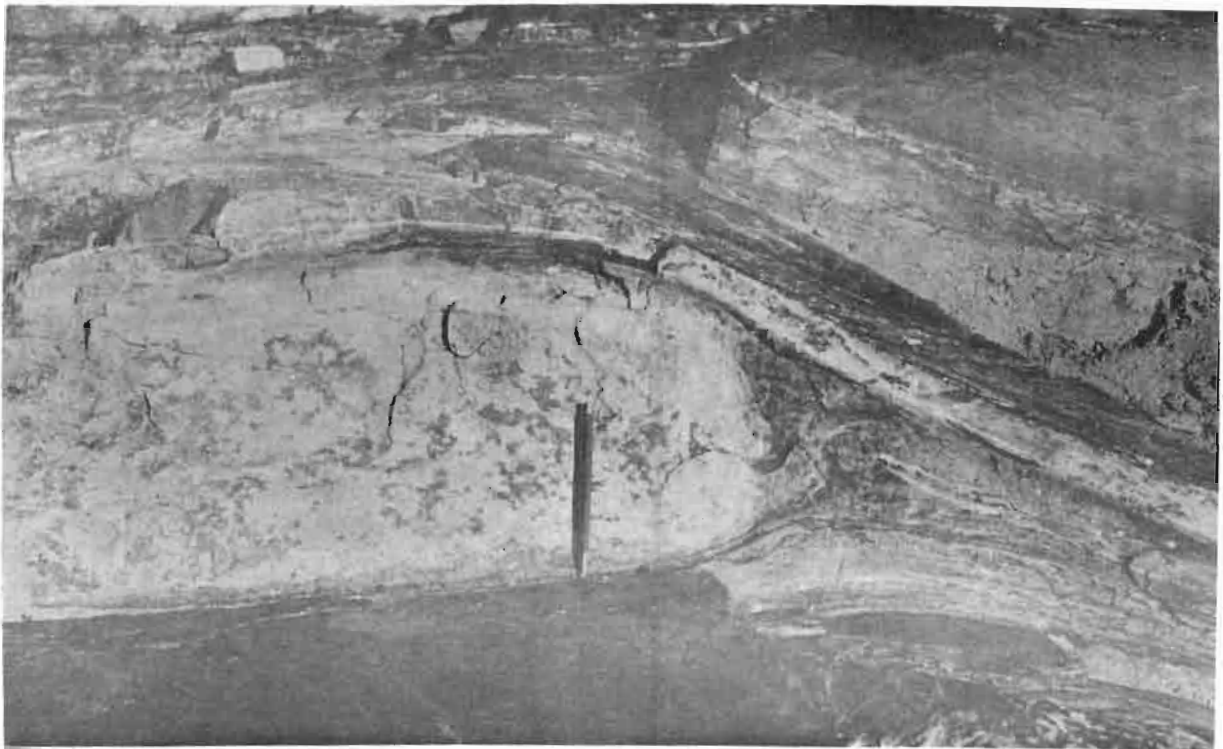


Figure 5. Boudinage at Sierra Mazatan.

Décollement:

Carbonate beds of Paleozoic age rest upon a distinct, strongly lineated, platy surface of strongly mylonitic granodiorite gneiss and quartzofeldspathic gneisses. The carbonate detachment rocks are locally folded. The intense microbrecciation and chlorite-epidote alteration so commonly seen in décollement zones have not been observed at Sierra Mazatan although detailed mapping of the décollement zone has not yet been undertaken.

MAGDALENA REGION

Introductory Comments.

In north-central Sonora, strongly deformed tectonites comprised of low- to medium-grade metasedimentary and metavolcanic rocks and associated silicic plutons underlie major north-trending sierras. These metamorphic tectonites are extensively developed along a north-south line in the region between Santa Ana and Imuris (Fig. 3). The tectonites extend from Sierra Madera on the east to at least as far west as the town of Altar. The zone of intensive deformation appears to be bounded by major NW-trending faults.

The town of Magdalena lies within this terrane of metamorphic core complex rocks. West of town, rocks on the eastern flank of Sierra Magdalena record the structural and metamorphic effects of Tertiary deformation.

Protolith.

Initial reconnaissance field mapping of a large area in north-central Sonora by Salas (1968[1970]) resulted in discovery of an extensive suite of greenschist-facies rocks composed of phyllite with subordinate marble, metaconglomerate, quartzite, and metagraywacke. In places, mineral assemblages characteristic of the almandine-amphibolite facies occur. Salas (1968[1970]) concluded that the layered rocks were composed of two suites, both of which were interpreted to be Precambrian in age. His conclusion was based on the fact that metamorphism recorded in this region is comparable to that reported from areas of Precambrian crystalline rocks that crop out to the west, south of Caborca.

Radiometric data (Anderson and others, 1980) indicate that one of the strongly deformed plutons considered to be Precambrian is actually of latest Cretaceous age. To date, no paleontologic or radiometric studies that bear directly on the age of the layered metamorphic rocks have been conducted. However, lithologically similar beds on regional strike from these metamorphic units yield mid-Jurassic ages.

For rocks on the eastern flank of Sierra Magdalena, metamorphism has obscured most primary features. The rocks are quartzofeldspathic gneisses and granitoid rocks. The gneisses are considered to be representative of the deformed silicic volcanic rocks of Jurassic age. Similar gneisses seen at other localities grade transitionally into Jurassic quartz porphyry. The strongly porphyritic granitic gneisses may have been derived from bona fide intrusive plutons, but they may have been derived as well from igneous material mobilized in place through anatectic processes.

Mylonitic Tectonite.

The tectonites derived from the Cretaceous plutonic rocks and the

Jurassic volcanic and sedimentary rocks share a common Tertiary (?) deformational fabric. Folds inherited from younger, pre-Tertiary (?) deformation can be discerned in the metamorphic rocks, but these are overprinted by a consistently aligned foliation/lineation fabric.

POZO VERDE MOUNTAIN

Introductory Comments.

Pozo Verde Mountain lies just west of the border-town of Sasabe, Arizona (Fig. 3). It occupies the southernmost end of the Baboquivari Mountain complex, which includes in the broadest sense the Pozo Verde, Baboquivari, Quinlan, and Coyote Mountains. Within the Pozo Verde Mountain terrane there is a 100 km² area of lineated mylonitic tectonite, formed at the expense of granitic protolith. The only other exposure of mylonitic tectonite within the entire Baboquivari Mountain complex is at the northernmost end, where the lineated tectonites of the Coyote Mountains crop out. Arched foliation in mylonitic tectonite in the Pozo Verde Mountains plunges gently south; that in the Coyote Mountains dips northerly. In between, the 50-km long Baboquivari Mountains consist entirely of non-mylonitic country-rock protolith. It is possible that the mylonitic tectonite may have passed as a continuous tectonite sheet over the entire stretch of terrain now occupied by the Baboquivari Mountains. However, it seems more likely that the northerly and southerly zones of mylonitic tectonite were never regionally continuous. Rather, they represent separate zones of shearing within a common system.

Protolith.

Haxel and others (1980) have discussed the geology of the Baboquivari Mountains in the context of the geology of the southern part of the Papago Indian Reservation. The main rock in the Pozo Verde Mountains was named (informally) by them "granites of the Presumido type", where Presumido refers to Presumido Peak in the northern part of the Pozo Verde Mountains. These granites are muscovite-garnet-biotite granites which generally are fine- to medium-grained, hypidiomorphic in texture. Haxel and others (1980) distinguish an older leucocratic phase which contains little garnet, and a younger, white-weathering leucocratic phase which usually has garnet. Compositionally the rocks are leucocratic monzogranites (Haxel and others, 1980).

The Pan Tak granite in the Coyote Mountains, one of the granites of the Presumido type, was dated at 58 ± 3 m.y., based on Ur-Pb on zircon (Wright and Haxel, 1980). Haxel and others (1980) believe that the granite in the Pozo Verde Mountain is also early Tertiary in age. A K-Ar determination on muscovite from granites of the Pozo Verde Mountain revealed a 24 m.y. cooling age (W.A. Rehrig and S.J. Reynolds, pers. comm., 1978).

Granites of the Presumido type intrude a country rock framework of dominantly Mesozoic rocks (Heindl and Fair, 1965; Haxel and others, 1980), although Mesozoic rocks are not part of the Pozo Verde Mountain. The Mesozoic rocks include metavolcanic and metasedimentary rocks, dominantly rhyolites and interbedded quartzites; unmetamorphosed clastic sedimentary rocks and rhyodacitic to andesitic volcanics; and syenogranitic to granodioritic plutons (Haxel and others, 1980). The plutons range in age from 170 to 140 m.y. and intrude the inferred Triassic (?) to early Jurassic sequence of metamorphic, sedimentary, and volcanic rocks (Haxel and others, 1980). The Mesozoic country-

rock framework makes up most of the Baboquivari Mountain complex. For example, the imposing Baboquivari Peak is undergirded by a 150 ± 10 m.y. granite (Haxel and others, 1980).

Mylonitic Tectonite.

Mylonitic tectonite in the Pozo Verde Mountains is wholly fashioned from granites of the Presumido type. Although the rocks are monotonously uniform compositionally, it is interesting to examine the upward increase in intensity of foliation and lineation. Granitic protolith with scarcely any distinguishable foliation and lineation is progressively transformed into penetratively deformed tectonite. The physiographic expression of this transformation is quite striking. Non-mylonitic outcrops of leucocratic rocks are weathered to rounded, exfoliated forms. In contrast, zones of mylonitic tectonite weather to subhorizontal layers and benches of moderately fissile rock.

Where most deformed, rocks of the Pozo Verde Mountain terrain are fine- to medium-grained quartz monzonitic augen gneisses. Aplite and quartz veins are ptgymatically folded into tight to isoclinal recumbent forms. In essence they are transposed into foliation-parallel alignment.

Mapping by Davis (1980b) has disclosed the general arch-like nature of mylonitic tectonite in the Pozo Verde Mountains. Foliation shows broad warping about the gently south-plunging axis (Fig. 6). Dips range in inclination from horizontal to approximately 35° . Foliations in the mylonitic tectonite do not conform to a simple arch, but one that is locally irregular in shape. Smooth curvilinear pinch and swell relationships are apparent both in map and outcrop relationships.

Mineral lineation is physically defined by alignment of unequidimensional chips of feldspar and elongate aggregates of plastically deformed quartz. More brittle counterparts of the lineation are slickenside striae. Lineation trends consistently north-northeast, with plunge values and directions congruent with local orientations of the mylonitic foliation in which the lineation is found. The trend of lineation in the Pozo Verde Mountains is transitional between the dominantly north-south trends which characterize the Papago Reservation domain to the north and west (Davis, 1980b) and the $N60^\circ E$ trends which typify both the Rincon-Catalina complex near Tucson and the complexes in northern Sonora.

The structural level of the upper part of the mylonitic tectonite is very close to the décollement zone, for rocks of décollement-zone affinity are exposed at westwardmost exposures along the base of the western flank of the Pozo Verde Mountains (Fig. 6). There, pegmatites and leucocratic granitic rocks are microbrecciated, shattered, and altered.

COYOTE MOUNTAINS

Introductory Comments.

The Coyote Mountains lie at the north end of the Baboquivari Mountain complex. The mountain is located south of State Highway 86 west of Three Points (Fig. 3). The northern flank of the Coyote Mountains has geologic properties which conform to characteristics of metamorphic core complexes (Davis, 1980b). Details of the structural geology and petrology of the mylonitic tectonite and décollement zone were investigated by A.F. Gardulski (1980). Unless otherwise noted, the descriptions and interpretations which follow are taken from her work.

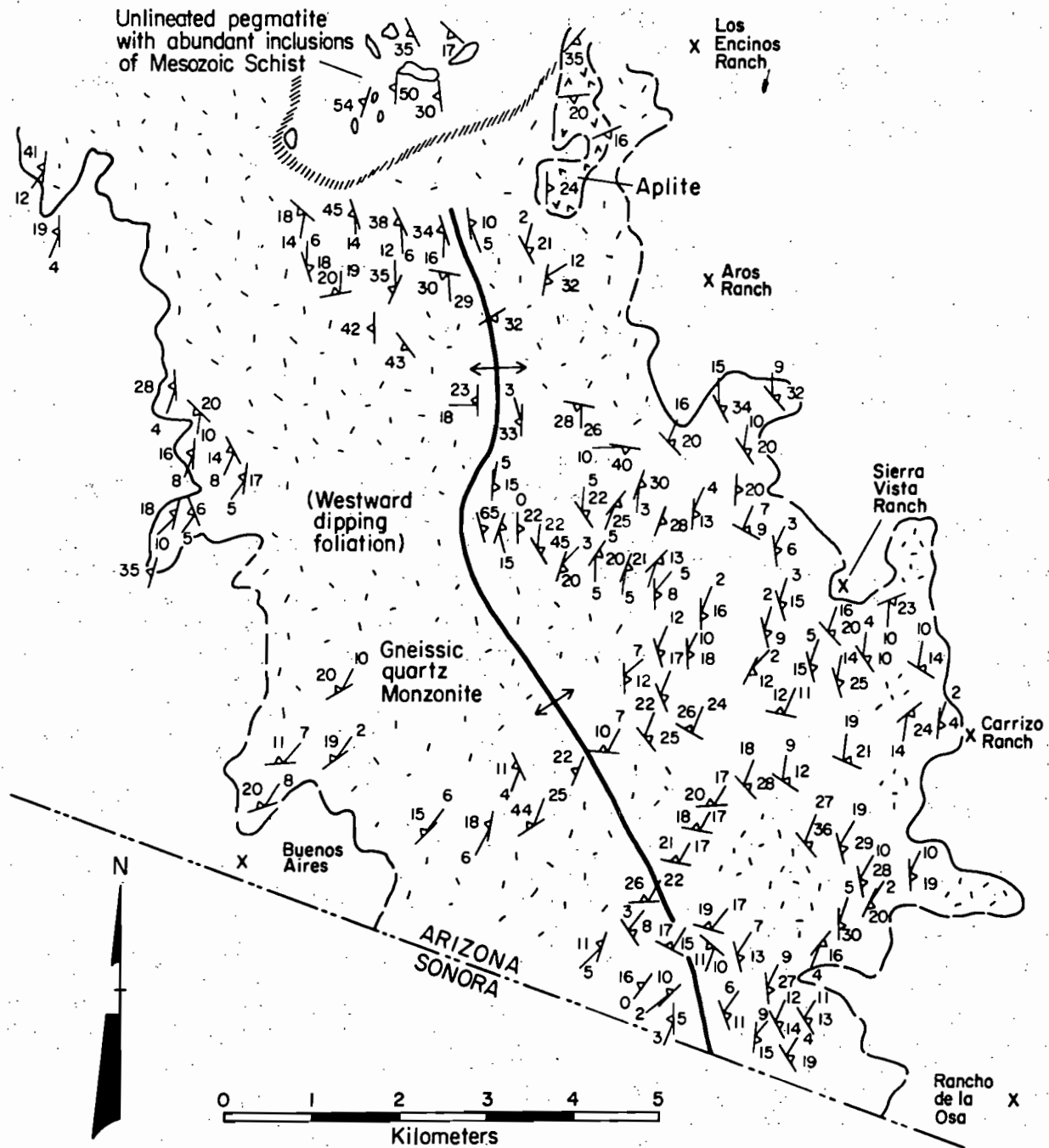


Figure 6. Structure map of Pozo Verde Mountains.

The general geology of the area which Gardulski (1980) studied is shown in Figure 7. The map shows a number of units, ranging in age from lower Paleozoic to Tertiary. The rocks which constitute the main focus of this discussion are Pan Tak granite and Bolsa Quartzite. The Pan Tak granite is the granite of Presumido type for which Wright and Haxel (1980) interpreted a U-Pb zircon age of 58 m.y. Both the granite and the Bolsa Quartzite were converted into mylonitic tectonite. Additionally, rocks of the Cambrian Abrigo and Devonian Martin Formations locally occur within the zone of mylonitic tectonite. These rocks crop out as deformed calc-silicate and marble assemblages.

A décollement zone, the Ajo Road décollement, separates mylonitic tectonite from overlying non-tectonite, unmetamorphosed sedimentary and volcanic rocks (Fig. 7). Closest to the décollement are andesitic to latitic volcanic flows, breccias, and derivative sediments of the Roadside Formation (Gardulski, 1980). Heindl (1965) assigned a Cretaceous age to this lithologic assemblage. Overlying the rocks of the Roadside Formation, still within the upper-plate detachment, are red clastic sedimentary rocks of the Cretaceous (?) Sand Wells Formation (G. Haxel, personal communication, 1980).

Protolith.

The Pan Tak granite has four phases -- an older equigranular granite and three younger phases. These younger phases include equigranular, coarse-grained granite; xenomorphic, medium-grained granite; and swarms of pegmatites. The pegmatites comprise a large part of the northern Coyote Mountains. The phase of the Pan Tak granite which yielded the 58 m.y. age date is the younger equigranular coarse-grained granite. The pegmatites are even younger. According to Wright and Haxel (1980), the isotopic systematics of zircons in the Pan Tak granite demonstrate an anatectic origin for the intrusion. A Pre-Cambrian parent rock is indicated.

The older granitic phase of the Pan Tak granite does not crop out in the study area which Gardulski mapped (1980). The younger phases crop out but are nearly identical compositionally. They are so intimately mixed that they are generally impossible to map as discrete units at a reasonable scale. The pegmatite and granite are commonly gradational, and there are few examples of clear-cut intrusive relationships between the phases.

Mineralogically, the pegmatite and granite consist of quartz, microcline, albitic plagioclase, biotite, and locally-abundant muscovite and garnet. In some sills, a delicate zoning is developed as compositional banding.

The Bolsa Quartzite is mineralogically homogeneous, quartzose rock containing only very minor feldspar, biotite, muscovite, and heavy minerals. It is white, grey, black, or more rarely, maroon in outcrop. The total exposed thickness of the quartzite is estimated at 30 m. Assigning this quartzite to the Bolsa Quartzite is an interpretation which we think is compatible with general outcrop appearance, mineralogy, and structural position underneath recognizable Abrigo and Martin Formations. No conglomeratic layers or lenses occur in the quartzite.

The overall geometry of rock relations in the northern Coyote Mountains has been greatly influenced by the intrusion of the Pan Tak granite suite. Intrusion apparently caused significant dilation of the Paleozoic section. Sills of the Pan Tak granitic system engulfed the sediments. Quartzite lenses are sandwiched between undulatory surfaces of the pegmatite sills. Sharp contacts are maintained in the Abrigo and Martin Formations, with little encroachment of pegmatite into the interiors of these sedimentary rock pods.



EXPLANATION

CENOZOIC - Quaternary Alluvium - Quaternary-Tertiary Gravels - Tertiary Lamprophyre Dikes <i>Pan Tak Granite Phases (58 MY- Date By J.E. WRIGHT for USGS)</i> - Pegmatite - Medium Grained Granite - Coarse Grained Granite MESOZOIC - Cretaceous (?) Sand Wells Formation - Mesozoic Roadside Formation - Unmetamorphosed Limestone <i>Diorite Rocks - Age Unknown</i> - Biotite Quartz Diorite - Hornblende Diorite - Hornblende PALEOZOIC - Paleozoic Abrigo and Martin Formations Undifferentiated - Cambrian Bolsa Quartzite	- Mylonitized Rock - Lithologic contact, with dip where known - Bedding - Primary layering in pegmatite - Foliation - Penetrative foliation and lineation - Discrete slip surfaces or set of slip surfaces and lineation - High angle fault - Joints in prominent pegmatite outcrops - Decollement surface, with sawteeth on upper plate - Lineation - Road - Prospect - Adit - State Route

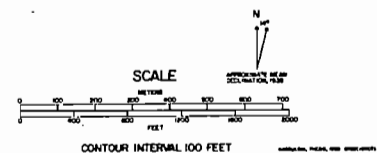


Figure 7. Geologic map of Tohawaw Canyon area, Coyote Mountains (Gardulski, 1980).

Mylonitic Tectonite.

Mylonitic tectonite is exceptionally well exposed in a canyon that Gardulski (1980) informally named Tohawaw Canyon. Tectonite quartzite derived from the Bolsa Quartzite is interlayered with pegmatite sills, the outer margins of which are converted to tectonite (Fig. 8). There are unusually good exposures of the quartzite plated or smeared onto tops of underlying sills (Fig. 9). Abrupt changes in dip of the quartzite, like that shown in Figure 9, reflect radical pinch-and-swell morphology of underlying pegmatite. Some of the sills are like giant augen (Fig. 8).

The exposure at the Tohawaw Canyon study site is mainly a broad expanse of tectonitic pegmatite. Quartzite layers are exposed in the walls, but the undulating floor of the canyon is mostly devoid of quartzite. Much of the floor seems to be a structural surface on the top of a series of pegmatite sills from which the once overlying quartzite has been removed. In effect, the canyon floor is a plan-view rendition of pinch-and-swell morphology of tectonitic pegmatite.

A map of the relationships at the study site is shown in Figure 10. This map was prepared by students in the graduate-level detailed structural analysis class at The University of Arizona (see caption, Fig. 10). The topography and geology of the site was mapped with plane table and alidade at a scale of 1" = 20 feet. The topographic map is an approximation of a structural contour map on the top of one family of pegmatites. Geologic contacts are the same as those shown in Gardulski's map, but in addition the large-scale map shows folded primary compositional layering in pegmatite, mylonitic shear zones which cut the pegmatite, several normal faults, and major through-going joints.

Bolsa Quartzite.

The quartzite is lineated tectonite wherever it crops out. A very strong mineral lineation and foliation penetrate the rock (Fig. 11), and these elements are defined by the orientation of streaked plates of muscovite and biotite as well as by flattened, elongate quartz grains (Fig. 12). Foliation strikes west-northwest and dips gently north-northeast; lineation plunges gently north (Fig. 13).

In thin section, the grains are seen to be undulose and partially to completely recrystallized; flattening has been extreme (Fig. 12). Elongation of quartz parallel to lineation is quite pronounced. Both original and recrystallized grains are deformed. Original grain shape is not preserved. Within individual lenses of quartzite, a great variation in strain is evident. The average long dimension of grains is at least an order of magnitude smaller in the tail of a lens than in the center (e.g., .02 mm vs. .2mm).

Small pinch-and-swell features with wavelengths of 1-10 cm occur locally in the quartzite, and the foliation and lineation have been warped by these small structures (Fig. 14). Where quartzite layers are thin (0.5 cm) and alternating with concordant veins of pegmatite, this pinch-and-swell feature more closely resembles mullion structures whose presence is due to the contrasting ductilities of the two media. They occur where underlying pegmatite has been stepped-down by normal faulting, thus requiring the quartzite to undergo major local stretching.

Small normal faults occur locally in the quartzite, and slip ranges from 1 mm to 20 cm in the direction of plunge of the lineation (Fig. 15).

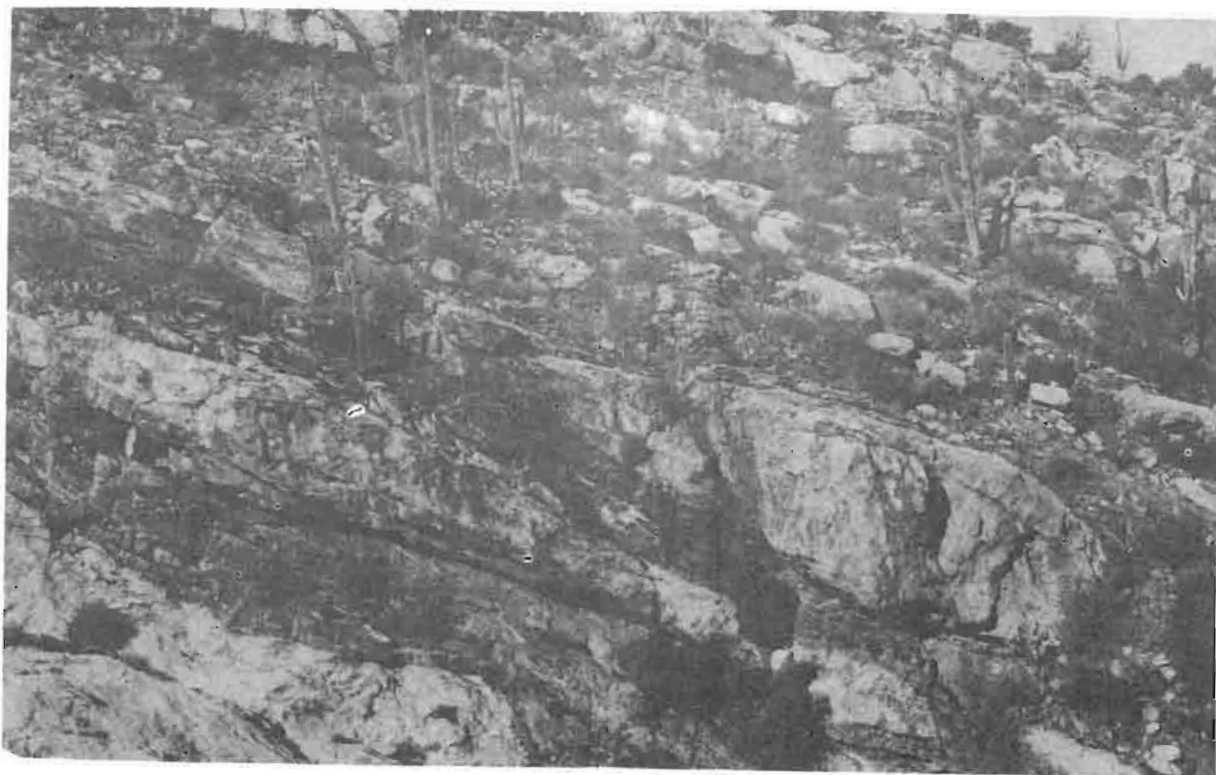


Figure 8. Interlayered quartzite and pegmatite sills, Tohawaw Canyon.



Figure 9. Quartzite plated on top of pegmatite sill, Tohawaw Canyon.

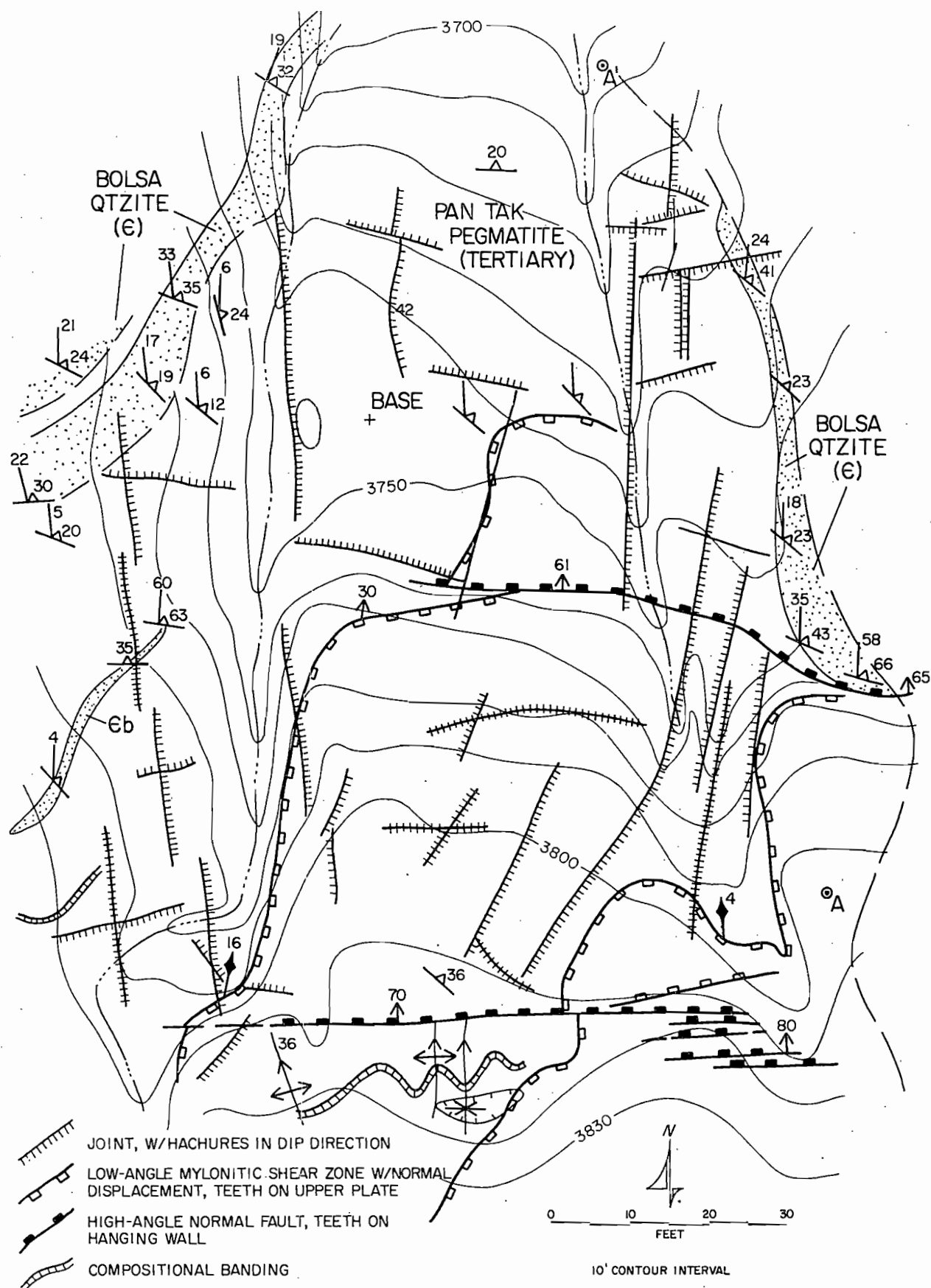


Figure 10. Large-scale structure map of part of Tohawah Canyon.



Figure 11. Lineated tectonite quartzite, Tohawaw Canyon.

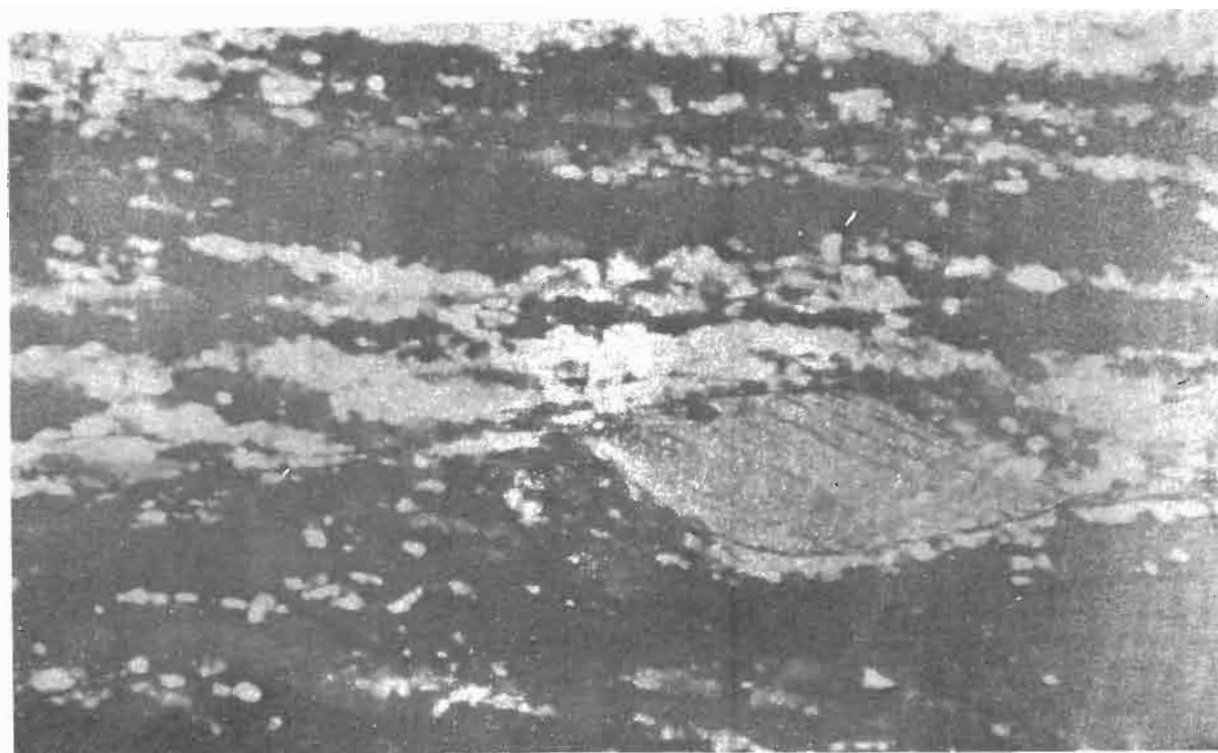


Figure 12. Photomicrograph of tectonite quartzite. 70X.

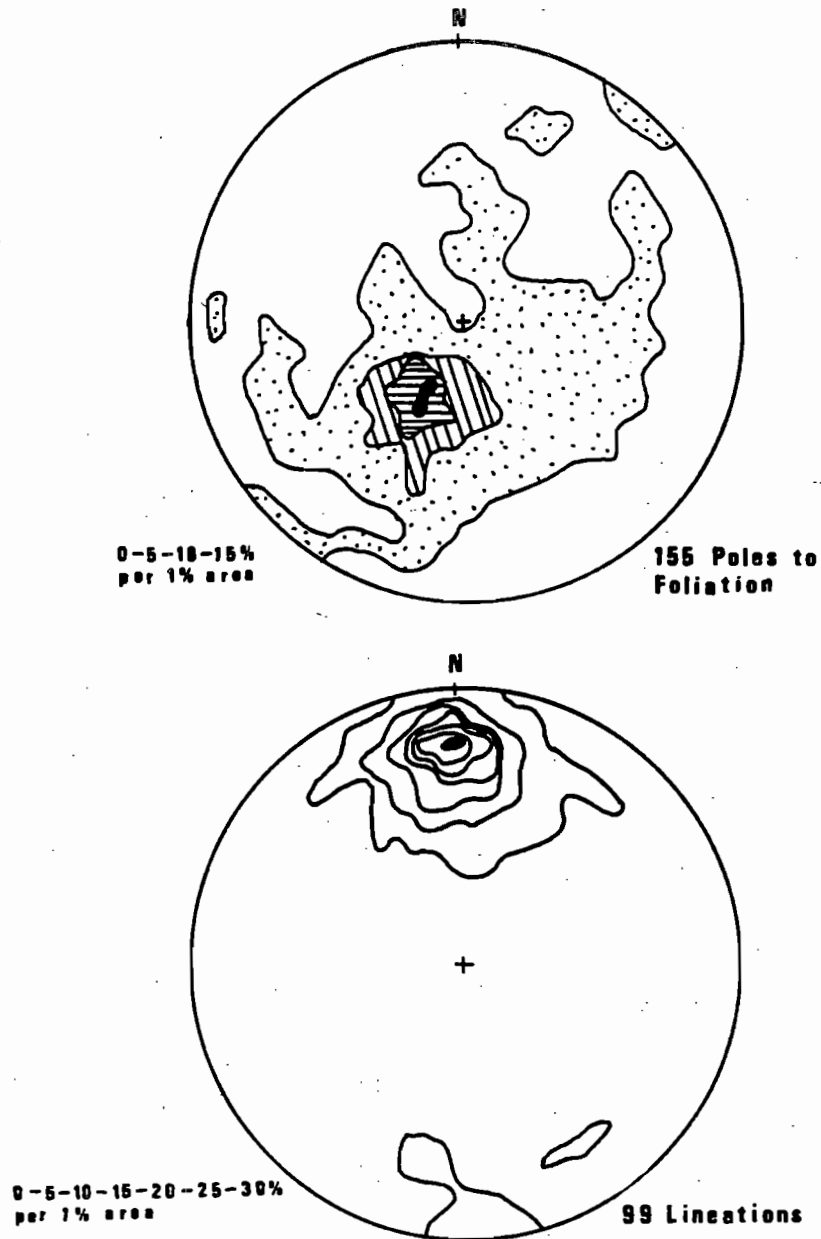


Figure 13. Lower-hemisphere equal-area projection of foliation and lineation, Tohawaw Canyon.

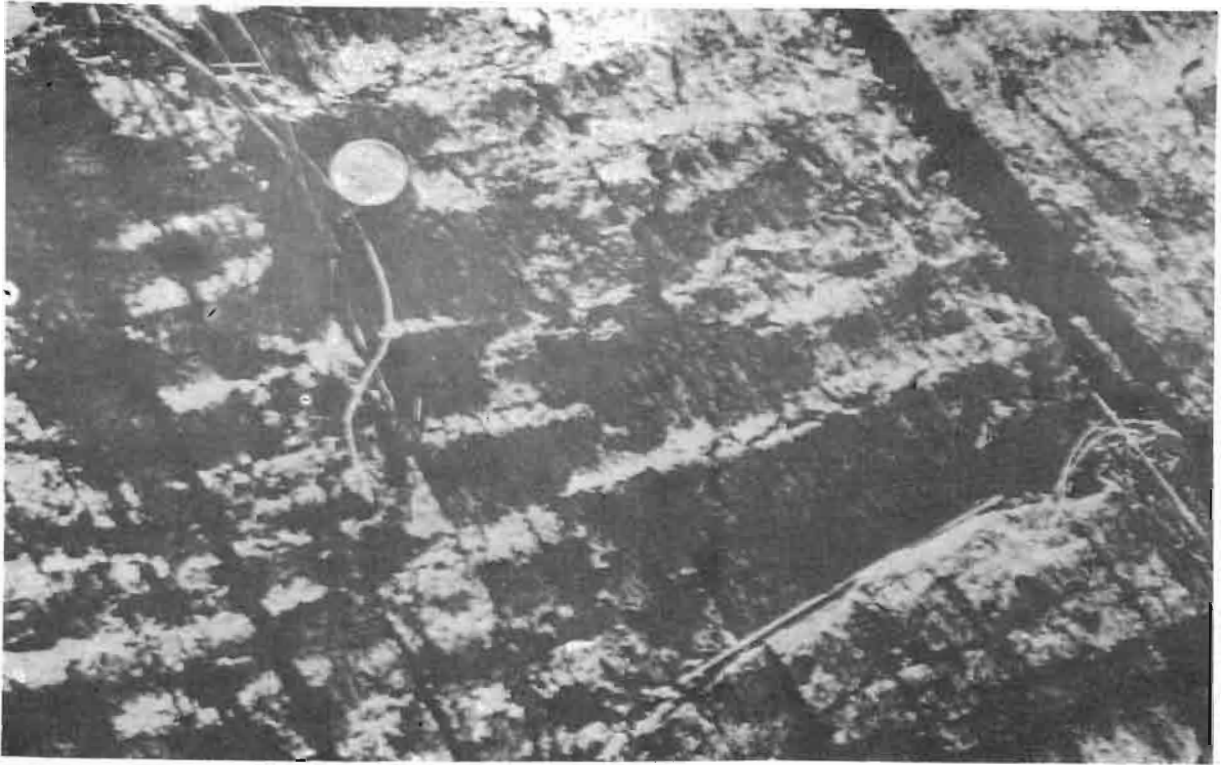


Figure 14. Small crenulations in lineated tectonite quartzite.



Figure 15. Normal faults in tectonite quartzite.

Faults with very small displacements occur within and parallel to the troughs of the pinch-and-swell features. Normal faults with larger amounts of slip are rare, but they affect both quartzite and pegmatite.

Thin section examination reveals the ductile nature of the small normal faults. In the necked regions of the pinch-and-swell structure, north-dipping steep surfaces of shear have developed, but there is no loss of cohesion. Shear zones cut through the necked regions of quartzite, stepping down, across, and through the layers, but without concurrent development of jointing. Large-scale geologic map relationships (Fig. 10) show comparable but larger shear zones in the tectonitic pegmatite. The zones display abrupt shifts in inclination, from subhorizontal to vertical. The mylonitic fabric of the shear zones shows micro-pinch-and-swell and penetrative micro-fractures at a high angle to foliation.

The light and dark banding visible in quartzite outcrops may represent relict bedding. Foliation is generally subparallel to this layering, but the angle between the two may be as great as 30° . There is evidence that bedding may have been isoclinally folded and transposed through passive flow. The strong overprint of the foliation and lineation in this rock has obscured these structures, and the homogeneous nature of the quartzite makes relict bedding hard to discern.

Structural petrological analysis of the quartzite by Gardulski (1980) revealed that the quartzite samples exhibit a strong maximum or split maximum that is located at high angle (60 to 90 degrees) to the lineation and within or very near the plane of the foliation (Fig. 16). The symmetry of the c-axis plots is nearly orthorhombic for several of the specimens, but strictly speaking they are all monoclinic to triclinic. An important feature of all the orientation diagrams for quartzite is the pole-free areas perpendicular to foliation and parallel to lineation. Assuming that the normal to foliation represents an axis of shortening and that the lineation represents an axis of lengthening, the c-axes of quartz cluster about the intermediate strain axis. Such distributions have been produced experimentally as discussed by Wilson (1975), and they arise when prismatic slip predominates over slip on the basal plane of quartz. Bouchez (1977) noted a c-axis maximum at the intermediate axis where prismatic slip in a basal direction was operative at relatively high temperatures. Tullis and others (1973) obtained similar results at high temperatures with slow strain rates and high strains. It is apparent that basal slip was essentially not operative by the time the quartz fabric was "frozen" in the rocks.

Pan Tak Granite.

The coarse-grained young phase of the Pan Tak granite has developed a strong foliation and lineation which may be considered penetrative within the constraints of its grain size and mineralogic heterogeneity. The quartz grains show features typical of plastic deformation--undulatory extinction, grain elongation, and recrystallization. A great deal of the strain in the granite is accommodated in the quartz component of the lithology, while the feldspar fraction has been merely fractured and sheared.

This style of deformation is also characteristic of the pegmatite phase, but it is not as penetrative. Discrete slip surfaces exist in the pegmatite, and these are well-developed only at the margins of the pegmatite bodies. They become very closely spaced as contacts with the quartzite are approached.

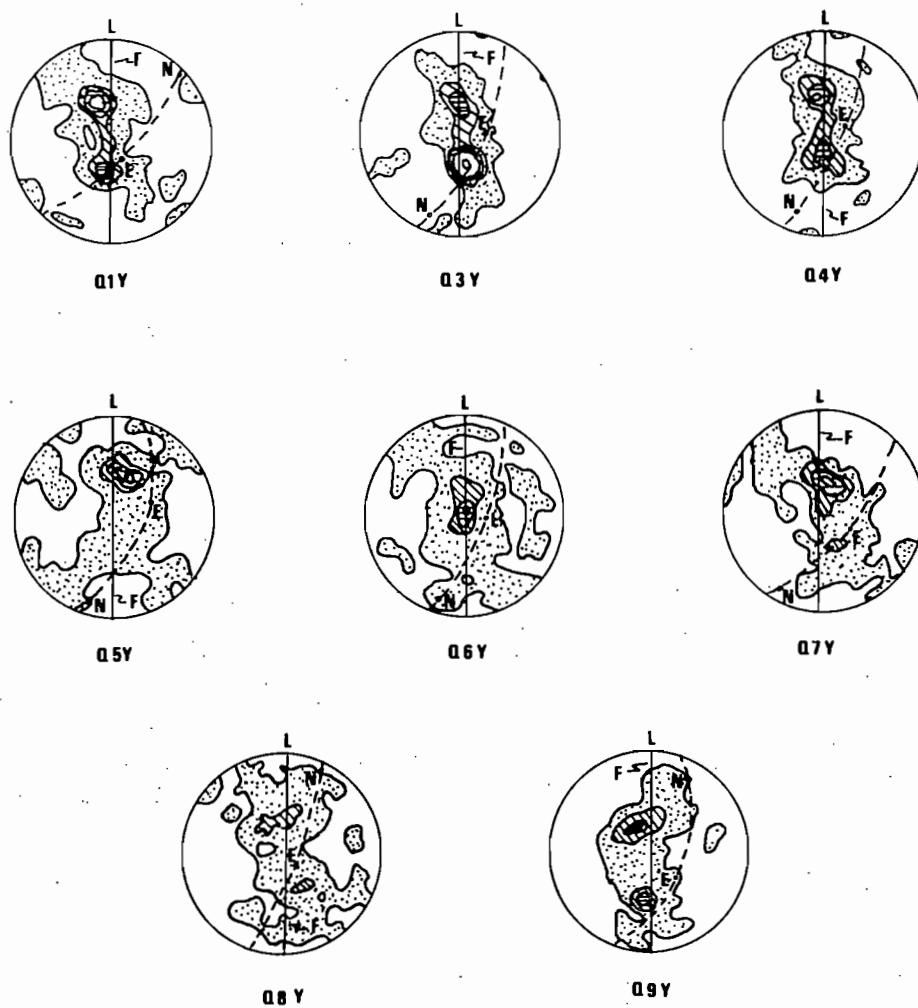


Figure 16. Lower hemisphere pole-density diagrams of c-axis orientations of quartz in tectonite quartzite, Tohawaw Canyon.

Where sedimentary rock packages are insignificant within large volumes of pegmatite, as in the southeastern part of the map area, fairly widely spaced slip surfaces are developed through zones as thick as 100 m. The zones of slip surfaces are separated by undeformed rock, and such alternating bands of deformed and undeformed rock may be traced east-west for hundreds of meters.

Gardulski (1980) examined the structural petrology of the pegmatite and found that there is a much greater scatter of low densities in c-axis distribution as contrasted with diagrams for the quartzite (Fig. 17). She attributed this to the high content of feldspar. As a brittle mineral under the conditions of deformation, the feldspar would protect certain zones of quartz from excessive strain. Attitudes of maxima of c-axes of quartz in the pegmatite are concordant with those for quartzite specimens.

Striated surfaces, spaced centimeters apart, pervade many parts of the tectonic pegmatite exposed in the study site. The surfaces display quite a range of orientations, but mostly they strike east-west and dip at gentle to moderate angles, either south or north (Fig. 18a). The strain significance of these surfaces is an accommodation to extensional stretching displacements and normal-slip. Striae on these surfaces show a strong preferred orientation, defining a great circle which trends north (Fig. 18b). The striae lie within the plane defined by the axes of greatest and least elongational. Some surfaces are spoon-shaped, with dip-slip striae in the depression and strike-slip striae on the walls. S.L. Beard prepared a structure profile (Fig. 19) parallel to the lineation direction along the pegmatite/quartzite contact (see A-A¹, Fig. 10). A main purpose was to evaluate the distribution and dip-directions of striated surface as a function of the pinch-and-swell morphology of tectonic pegmatite. The carefully constructed structural profile shows the stretched nature of the upper surface of the pegmatite(s). Joints in the pegmatite are well developed and systematically aligned. E.Y. Anthony, D.A. Currier, and K.F. Inmann analyzed the jointing and documented the presence of orthogonal sets, one striking N-S, the other striking E-W (Fig. 20). Fracture density in the pegmatite averages about 0.1 cm⁻¹ (data from Anthony, Currier, and Inmann).

Décollement.

The mylonitic terrane of the Coyote Mountains is tectonically juxtaposed against the Roadside Formation and Sand Wells red beds by the Ajo Road décollement (Fig. 7). Slickensides plunge steeply at 40-60° to the north on the fault surface, and normal movement is indicated from the stratigraphic and lithologic relations. The décollement strikes west-northwest and dips north at an average of 45° (Davis, 1980b). In detail it is curvilinear with a sinuous trace along the mountain-pediment interface.

The hanging wall rocks consist of the Roadside Formation overlain by the Sand Wells Formation. These rocks are everywhere moderately fractured, but in the vicinity of the décollement they are shattered and assume a pasty, gougy appearance. At one exceptionally fine exposure, a sliver of intensely sheared Roadside Formation crops out along the fault surface as the lowest member of the upper plate. It is overlain by similarly deformed red beds of the Sand Wells Formation. At all other exposures of the fault in the map area, the Roadside Formation is absent and the red beds occupy the lowest structural position in the hanging wall.

The footwall rocks by the Ajo Road décollement consist primarily of

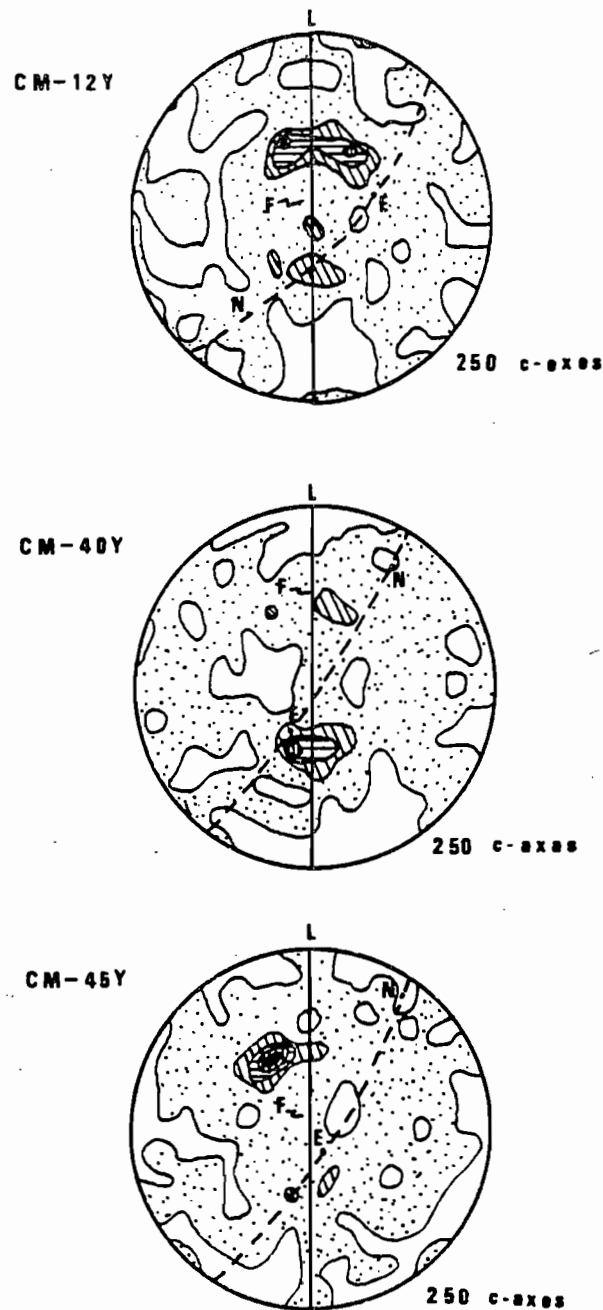


Figure 17. Lower hemisphere pole-density diagrams of c-axis orientations of quartz in pegmatite, Tohawaw Canyon.

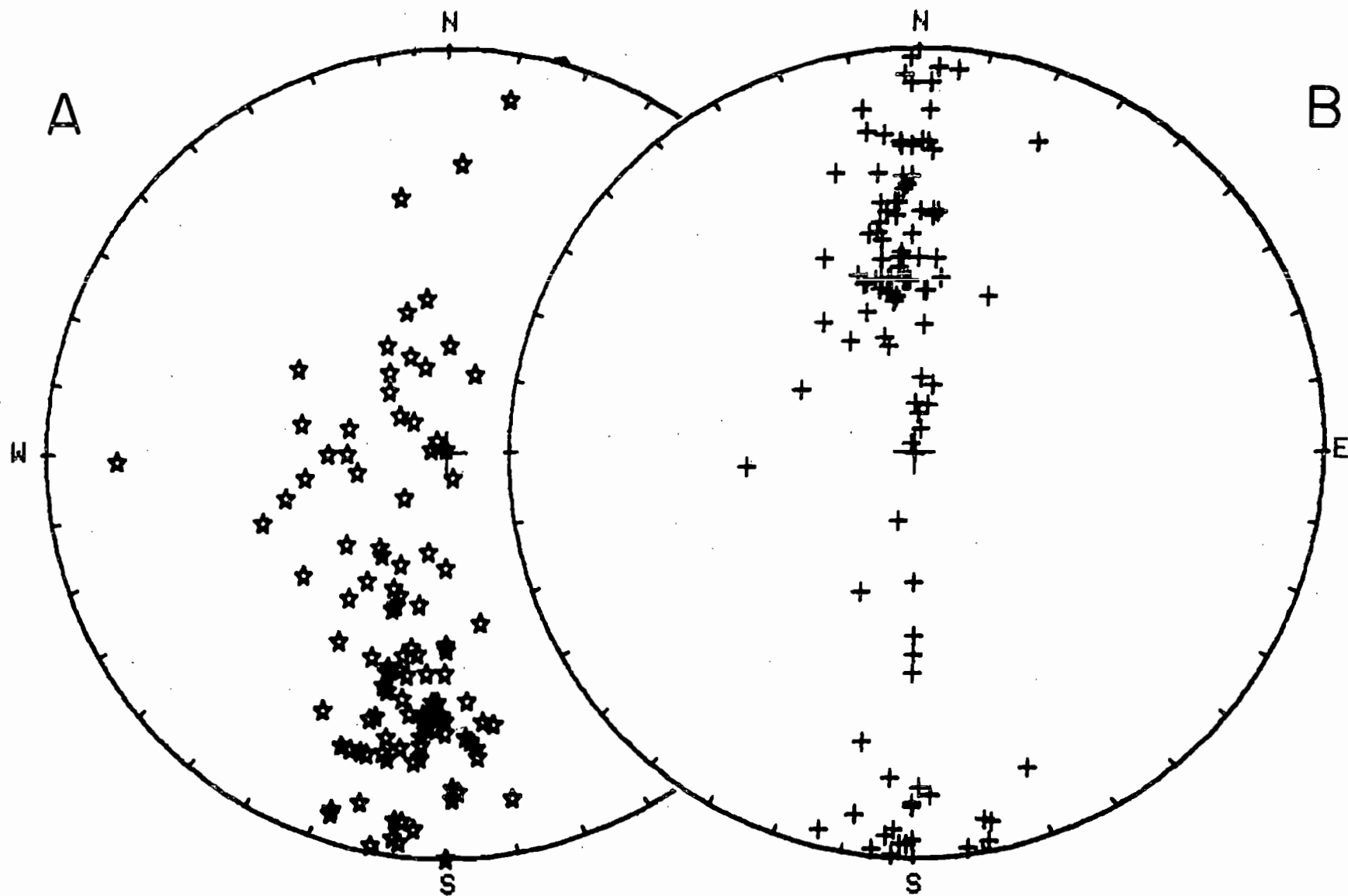


Figure 18. Lower hemisphere equal-area projections of (A) striations and (B) striated surfaces.

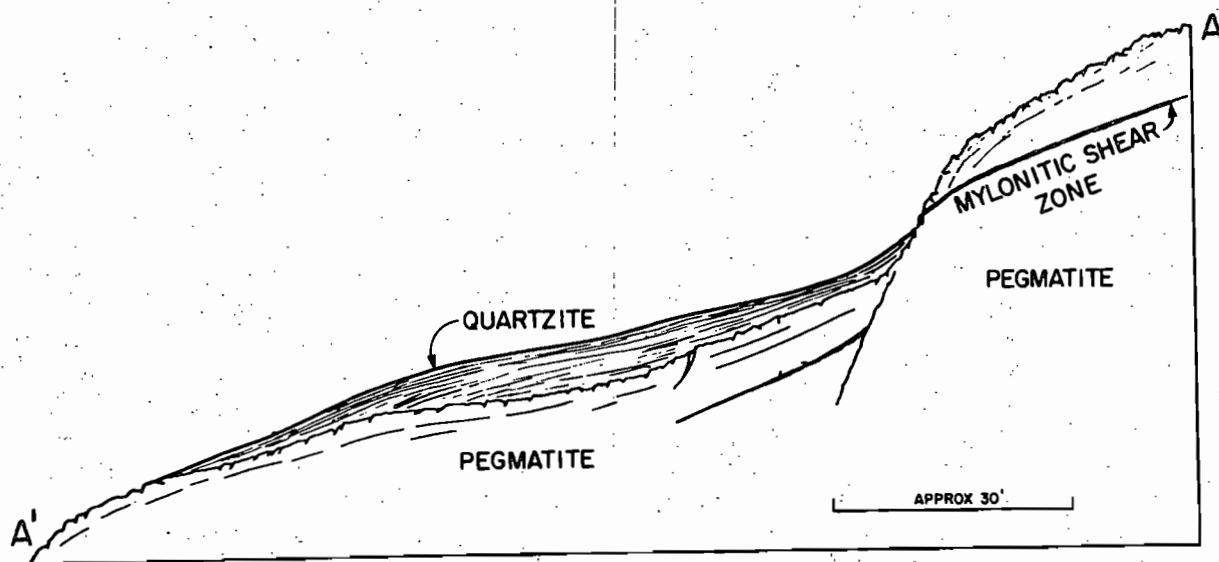


Figure 19. Structure profile showing traces of striated surfaces at outer margin of pegmatite sill(s). Tectonite quartzite is plated on the top of the pegmatite.

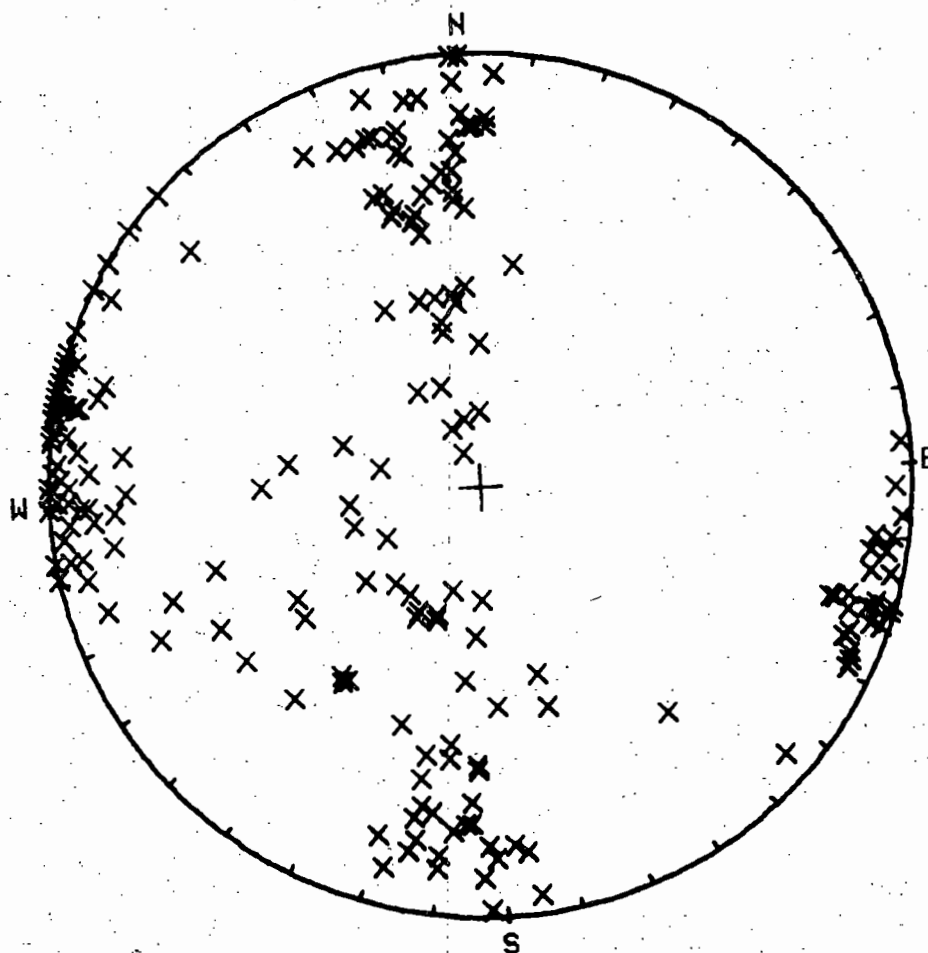


Figure 20. Lower hemisphere equal-area projection of poles to joints, Tohawaw Canyon.

Pan Tak granite and pegmatite, and lamprophyre dikes. Microbrecciation and a brittle, mylonitic foliation marks the décollement zone, and it overprints the earlier, more ductile, mylonitic foliation in the Pak Tak. The décollement-related imprint is characterized by intense brecciation of the footwall rocks, and faint foliation associated with this faulting is measurable in outcrop. The outcrop expression of the footwall near the décollement is that of a brown-to-green, chloritic, tabular mass of rock whose protolith may be determined only from field relations. The mineralogy in hand sample is indeterminate, and alteration and brecciation are so extreme that one of the most notable features of the rock is its fissile but shattered character. The intensity of microbrecciation in these rocks decreases rapidly down and away from the décollement, although a clear cataclastic overprint is visible through at least 100 m of rock.

In thin section, the rocks of the décollement zone include true mylonites, and subangular to subrounded fragments floating in comminuted groundmass attest to intense movement on the fault. Brittle mylonitization and microbrecciation can be seen to overprint previously deformed granite and pegmatite. Locally, quartz veins lace the mylonite, and the quartz is unstrained and unrecrystallized.

TANQUE VERDE MOUNTAINS

Introductory Comments.

The geology of the Tanque Verde Mountains has been mapped and discussed by Drewes (1975, 1978) (Fig. 21). Rocks in the mountain comprise an exceptionally good exposure of metamorphic core complex terrane (Davis, 1980b). The site of interest is located on the northwest flank of the mountain within the general area of the Loop Drive of Saguaro National Monument (East) (Fig. 22). Mylonitic tectonite, décollement zone, and detachment rocks are all well exposed. Analysis of the internal fabric of these bears importantly on kinematic interpretations of the structural evolution of metamorphic core complexes. Especially instructive is the physical and geometric nature of rocks within the décollement zone.

Protolith.

Mylonitic tectonite on the northwest flank of the Tanque Verde Mountains has been derived from two different plutonic rocks (Fig. 21). A coarse-grained mylonitic gneiss, mapped by Drewes (1978) as Precambrian Y Continental Granodiorite, is cataclastically deformed Precambrian quartz monzonite. The mylonitic gneiss is identical texturally and compositionally to a gneissic phase in the foothills of the nearby Santa Catalina Mountains, a rock now recognized as overprinted 1.4 b.y. "Oracle Granite". Creasy and Theodore (1975) originally recognized that the coarse-grained quartz monzonitic gneiss was derived from the Oracle Granite. Shakel and others (1977) extracted zircons from the gneiss and, using Ur-Pb isotopic analysis, found them to yield an age date of 1440 ± 10 m.y. The progressive transformation of the unfoliated Precambrian porphyritic quartz monzonite to coarse-grained mylonitic tectonite can be observed at a number of locations in southern Arizona (Davis and others, 1975; Banks and others, 1977).

The second variety of mylonitic gneiss in the Tanque Verde Mountains is fine- to medium-grained two-mica garnet-bearing quartz monzonitic augen

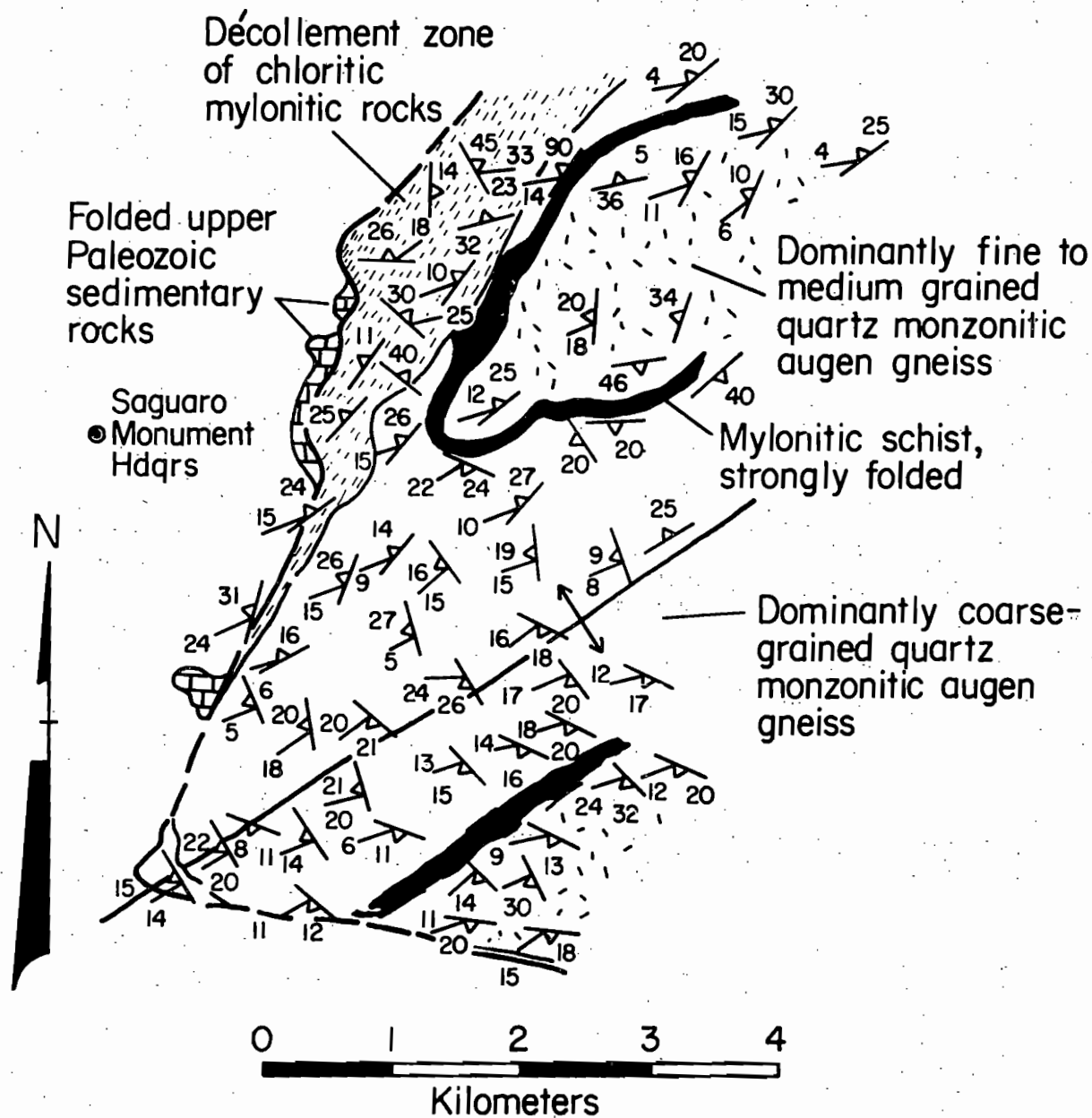


Figure 21. Geologic map of Tanque Verde Mountain area.

gneiss, mapped by Drewes (1978) as Precambrian Y Wrong Mountain Quartz Monzonite. To the north in the Santa Catalina Mountains, Creasey and Theodore (1975) and Banks (1976) mapped this same rock as part of the Quartz Monzonite of Sanmaniego Ridge. They interpreted the two-mica garnet-bearing granite as part of a mid-Tertiary (25 m.y. $\pm$) composite batholith (Creasey and others, 1977; Banks, 1980). Shakel and others (1977) reported that one phase of the two-mica garnet-bearing granitic rock in the Santa Catalina Mountains yielded monzonite crystals which gave U-Th-Ph ages of 44 to 47 m.y. Silver (1978) concluded that the Quartz Monzonite of Wrong Mountain may possibly correlate with the Tertiary granites and gneisses of the Santa Catalina Mountains. Recent isotopic and geochemical investigations by Keith and others (1980) suggest that the two-mica garnet-bearing granites in the Santa Catalina Mountains are about 50 m.y. old, further supporting the probability of the Tertiary age of the texturally and compositionally identical Wrong Mountain Quartz Monzonite.

Mylonitic Tectonite.

The coarse-grained mylonitic tectonite is relatively dark colored in outcrop and cut by penetrative flat to moderately dipping foliation (Fig. 23). Concordant aplite and pegmatite sills enhance the foliation, and these commonly display boudinage and pinch-and-swell. Aplites and pegmatites that cut the foliation at high angles are almost always ptymatically folded. The folds are axial planar with respect to low-dipping penetrative foliation and are typically strongly overturned to recumbent. The mylonitic tectonite everywhere displays low-plunging mineral lineation of a cataclastic nature (Fig. 24). The lineation is penetrative even at the scale of a single hand specimen, with a delicacy of development dependent in large part on the grain size of the host rock. The lineation is expressed by the long directions of inequant feldspar and quartz-feldspar augen, striae and slickensides, elongate aggregates and streaks of crushed minerals, and crenulations on quartz ribbons.

The contact zone between the coarse-grained quartz monzonitic mylonitic tectonite and the medium-grained two-mica garnet-bearing tectonite gneiss is marked by strongly foliated, folded rocks described here as mylonitic schist (Fig. 22). These rocks are fine- to very fine-grained and range in color from brown to steel gray to black. White broken feldspar chips, irregular in size and shape, commonly "float" in the fine-grained, mylonitic matrix. Deformed concordant aplite and pegmatite layers and veins accentuate the foliation. Penetrative folding is a characteristic of these rocks and presumably reflects movements that helped to shape the mylonitic, foliate fabric. The folds are intrafolial, tight to isoclinal, overturned to recumbent structures whose axes lie in the low-dipping foliation. It would seem that the mylonitic schist has been derived from cataclastic reduction of one or both of the mylonitic gneiss phases. Adjacent coarse-grained mylonitic tectonite is locally marked by transposition and folding of the penetrative, low-dipping foliation.

The fine- to medium-grained mylonitic gneisses are light-colored, hypidiomorphic, equigranular, and very homogeneous. They lack the biotite and large porphyroclasts so characteristic of the coarse-grained tectonite gneiss. Foliation is defined by thin laminae of quartz, parallelism of micas, and aligned augen of cataclastically deformed feldspar. The rocks are cut by abundant pegmatite, aplite, and flat joints.

Foliation of mylonitic tectonite on the northwest flank of the Tanque Verde Mountains strikes north-northeast and dips gently northwest. Lineation trends S.65°W., plunging at about 10° to 15°.

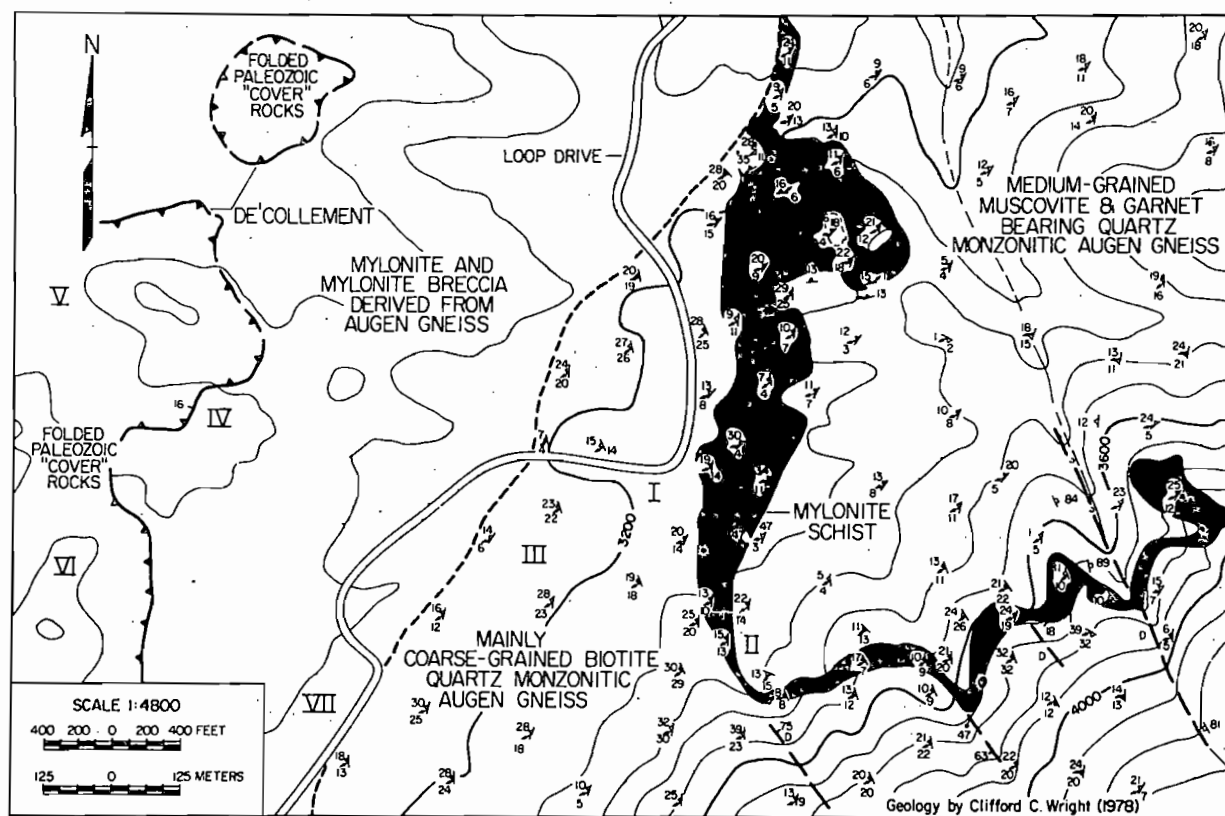


Figure 22. Geologic map of Tanque Verde Mountain area, prepared by C. Wright, senior thesis, Carleton College.

Décollement Zone.

At Saguaro National Monument, in the vicinity of the eastern stretch of the Loop Drive, the contact between mylonitic tectonite and the décollement zone is abrupt. The coarse-grained mylonitic gneiss is overlain in fault contact by microbrecciated and rotated mylonitic gneiss (Fig. 25). The position of the fault boundary is distinguished by (1) abrupt change in orientation of foliation and lineation, and (2) alteration, shattering, and microbrecciation of the mylonitic gneiss. Normal coarse-grained mylonitic augen gneiss below the décollement zone consists of large (up to 4 cm) feldspar porphyroclasts of rounded to elliptical shapes in a matrix of mainly quartz and feldspar. Microscopically, the feldspar porphyroclasts show pull-apart structures with infilling by quartz and chlorite. Quartz occurs in several modes: feldspar-free zones of recrystallization in which quartz is undulatory and very fine grained; zones of flattened, elongate, extremely undulose quartz (L:T = 20:1) adjacent to feldspar porphyroclasts; and zones of polygonal quartz in "pressure shadows" next to feldspar augen. Fracturing is minor and alteration of feldspar is insignificant in the nonmicrobrecciated tectonite gneiss.

By way of contrast, the microbrecciated tectonite gneisses of the décollement zone are marked by extreme fracturing, brecciation, microfaulting, and alteration. The foliation visible in normal gneiss is nearly obliterated by anastomosing bands of microbreccia which cut through the rock. Microfaulting has caused rotation and tilting of micro-fault blocks of up to 60° or 70°. Some parts of this rock are so mylonitized that only a dense aphanitic mass remains. Veinlets of chlorite, epidote, biotite, and opaque minerals are abundant and lace the rock in apparently all orientations. Grain size is greatly reduced.

Different grades of microbrecciated mylonitic tectonite within the décollement zone have been mapped by Davis, and their distribution is shown in Figure 25. The intensities of microbrecciation are described as weak (w), moderate (m), and strong (s). Gardulski studied the petrography of representative examples of each of these field-based intensity classes. Where microbrecciation is weak, feldspar porphyroclasts remain very large (1 to 2 cm) but they are quite fractured. The porphyroclasts are rounded and not uncommonly elliptical. Some show significant pull-apart and chlorite veining. The infilling chlorite displays strongly anomalous interference colors, unlike the finer-grained alteration chlorite found in the matrix. Quartz grains above and below the feldspar porphyroclasts are markedly elongate (10 to 1 ratios) and/or very fine-grained (.001 mm). Extinction in the quartz is much more undulose than for quartz in feldspar-free areas. Away from the feldspar porphyroclasts, the quartz tends to be equidimensional and approximately .01 mm in grain size. Subgrain boundaries are abundant and original grain boundaries are usually vague because of recrystallization. Foliation is deflected around the large feldspar porphyroclasts. Quartz grains in pressure shadows of the feldspar porphyroclasts are relatively coarse. Thin laminae of micas and opaque minerals display anastomosing patterns around feldspar porphyroclasts and lenses of quartz grains.

The moderately microbrecciated rocks are marked by a number of petrographic changes which serve to distinguish them from normal or weakly microbrecciated tectonite gneiss. Quartz is strongly elongate (approximately 10:1) except where shielded by porphyroclasts of feldspar. Grain size of the quartz averages .01 mm. The feldspar porphyroclasts display selvages of re-

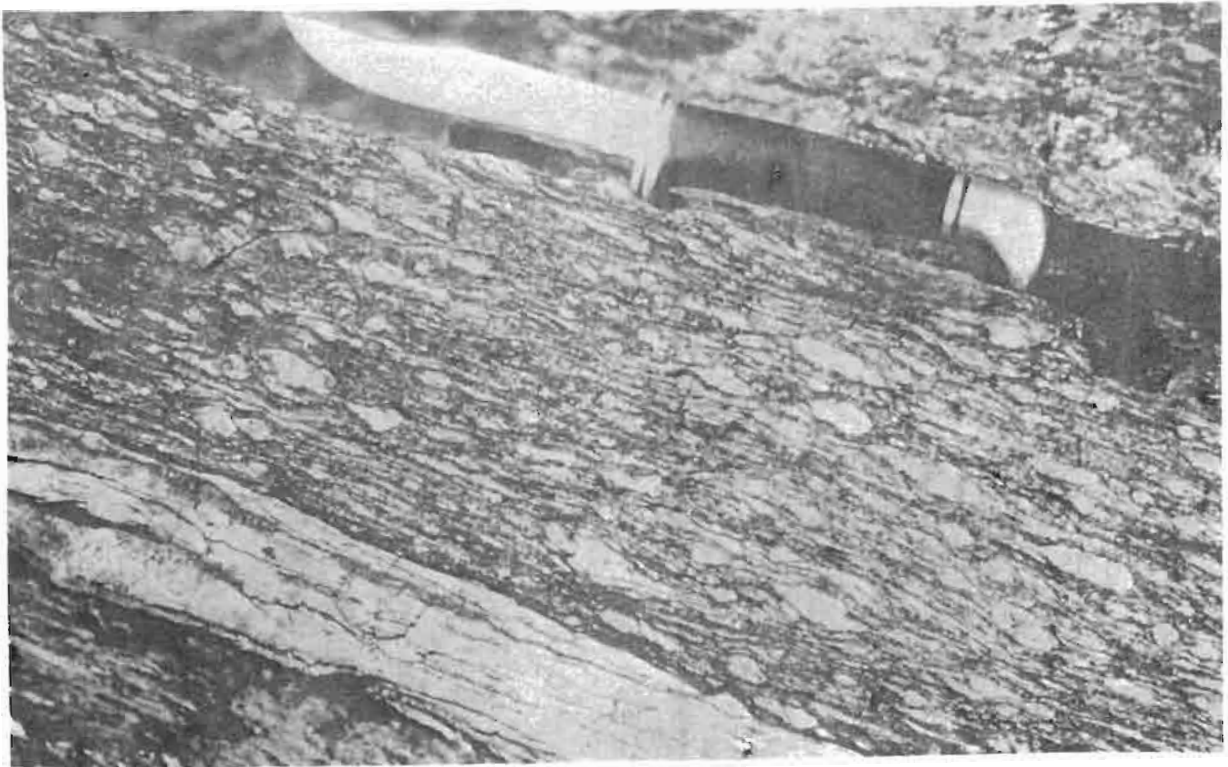


Figure 23. Foliation in tectonite gneiss.

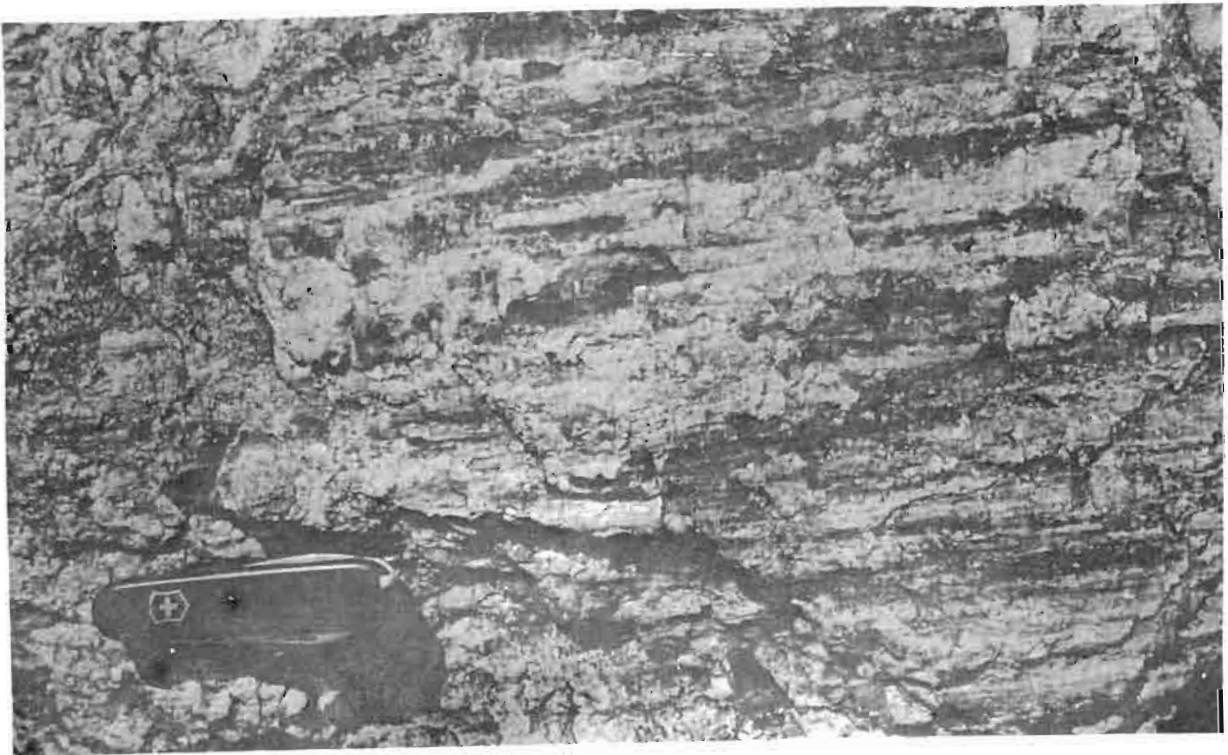
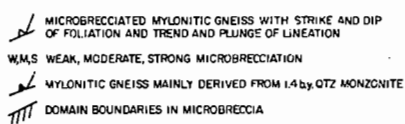


Figure 24. Lineation in tectonite gneiss.



crystallized feldspar. Such feldspar is very fine-grained, approximately .0005 mm. The regions of feldspar recrystallization lie above and below the porphyroclasts, in pull-apart areas, and sometimes in pressure-shadow zones. There is a near total absence of biotite, opaque minerals, epidote, and apatite, and thus the foliation is almost completely defined by elongate quartz grains. Muscovite is coarse grained (.2 to .3 mm) and is recrystallized. The relative abundance of muscovite and the decline in amount of plagioclase feldspar implies that alteration and neomineralization may have been important processes at this stage of deformation.

Strongly microbrecciated tectonite gneiss shows almost complete overprinting of the original foliation. The transition from moderate to strong microbrecciation is marked by decrease in size of feldspar porphyroclasts to an average of .15 to .25 mm. There is a corresponding increase in percentage of recrystallized fine-grained, slightly altered feldspar in the groundmass (up to 1/3 total feldspar). There also appears to be a greater amount of very fine grained white mica, probably muscovite, and this phase helps to define the foliation. The feldspar porphyroclasts which do survive are noticeably more fractured than in the moderately microbrecciated tectonite. Quartz grains are extremely elongate adjacent to the feldspar porphyroclasts. Chlorite content shows an increase and is commonly associated with opaques and epidote. Veinlets of chlorite and epidote plus calcite are moderately abundant.

Where the tectonite gneiss is indeed strongly microbrecciated, the fabric of the samples is one of intense deformation. There are abundant veinlets of chlorite, epidote, and biotite, and these follow innumerable micro-faults which pervade the rock. The average size of feldspar chips is about .1 mm, with the largest feldspars only .3 mm or so. The fine-grained groundmass has an average grain size of only .0005 mm. It consists of quartz and feldspar and comprises an average of 50% of the rock. The larger feldspar chips as well as lenses of coarser quartz (.005 to .03 mm) float in the matrix. Original foliation is absent except where preserved in some micro-fault blocks. Many of the strongly microbrecciated rocks have a 'swirled' appearance due to the abundance of anastomosing fractures and microfaults which lace the rocks. Tiny breccia zones are common, as are veinlets of chlorite and opaques. Although foliation may be preserved in some of the microscopic fault blocks, there are abrupt changes in orientation of the foliation from block to block.

In outcrop, the rocks of the décollement zone are fine-grained and pervasively overprinted by shattering along closely-spaced fractures. Lineation and foliation of the original tectonite gneiss can be recognized, but these are generally masked by the microbrecciation. On fresh surfaces the rocks are sometimes brilliantly blue-green, but their typical weathered color is a drab brown.

Mapping the internal structural fabric of the microbrecciated mylonitic tectonites is very difficult, but at the same time very revealing. Measurements of relict foliation and lineation in the microbrecciated tectonite disclose that the décollement zone is made up of innumerable fault-bounded blocks of a rotational nature. One is struck by the similarity of rotational faulting both on microscopic and macroscopic scales. At the Saguaro National Monument site, relict foliation in the décollement zone generally departs radically from the normal north-northeast strike to an average which is northwest (see Figs. 21 and 25). The typical east-northeast lineation trend for normal tectonite shifts to other orientations, notably southerly. Locally the foliation is vertical, and it is possible that relict foliation in some blocks has been overturned, although this is difficult to prove.

The surprising discovery afforded by detailed structural mapping is that the changes in foliation-lineation attitude are not random, but rather, are very systematic. Domains can be identified within which foliation and lineation are uniform in orientation. These are separated from adjacent domains by inferred or poorly exposed fault boundaries. Where fault displacements are observable they typically show normal separation.

The structural characteristics which make the décollement zone so distinctive formed in the late stages of and/or after the formation of mylonitic tectonite. The microbrecciated mylonitic tectonite which makes up most of the décollement was produced during imbricate faulting and rotation of the mylonitic tectonite. In parts of the décollement zone where mylonitic tectonite is only modestly overprinted by microbrecciation and faulting, there is bimodality of strike orientation of the relict tectonite foliation: some remains in normal attitude, some is rotated to a strike which is nearly perpendicular to original strike.

In effect the décollement zone appears to be a zone of coalescence of gently to moderately dipping faults, some of which may be the lower reaches of listric normal faults. Multiple superposed fault movements locally achieved radical rotation of originally gently dipping foliation. Although the microbrecciation and rotation post-dated the formation of the original mylonitic tectonite, the tectonic movements which produced microbrecciation and rotational faulting seem to be coordinated with those which produced the original lineation. This is shown by the tendency of the strike of rotated foliation to become aligned at right angles to the trend of lineation in underlying mylonitic tectonite. Like that of the mylonitic tectonite, the strain significance of the formation of microbrecciation and rotational faults in the décollement zone is accommodation to near vertical flattening and east-northeast elongation.

Faulting, rotation, and microbrecciation were carried out at a time and at a structural position such that no upper-plate detachment strata became interleaved with microbrecciated mylonitic tectonite of the décollement zone. The very fine-grained microbreccia ledge beneath the décollement surface caps the underlying décollement zone rocks, typically in angular discordance. It too is probably the coalescence of faults (Shackelford, 1977, 1980). Rocks in the capping ledge are so thoroughly microbrecciated that no relict foliation can be seen.

Detachment Rocks.

The décollement surface at the Saguaro National Monument site strikes north-northeast and dips gently northwest. Overlying the décollement are detachments of non-tectonite Precambrian, Paleozoic, Mesozoic, and Tertiary rocks. In the immediate area of interest, the upper-plate detachment rocks are Pennsylvanian(?)–Permian limestones and mudstones, intricately folded and faulted but not generally metamorphosed.

Rocks and structures of the Paleozoic and Mesozoic detachments of the Tanque Verde and Rincon Mountains have been described and discussed by Davis (1973, 1975). Davis proposed a gravitational-tectonic deformational model to explain the formation of the overturned to recumbent fold structures which occupy "cover rocks" exposed from place to place around the base of the gneissic complex. Such folds are very well exposed in Saguaro National Monument (East) and a map revealing some of the complexity of internal deformation appears as Figure 26. The model for gravity-induced folding of such rocks argued against the consensus interpretation that the folds were produced by

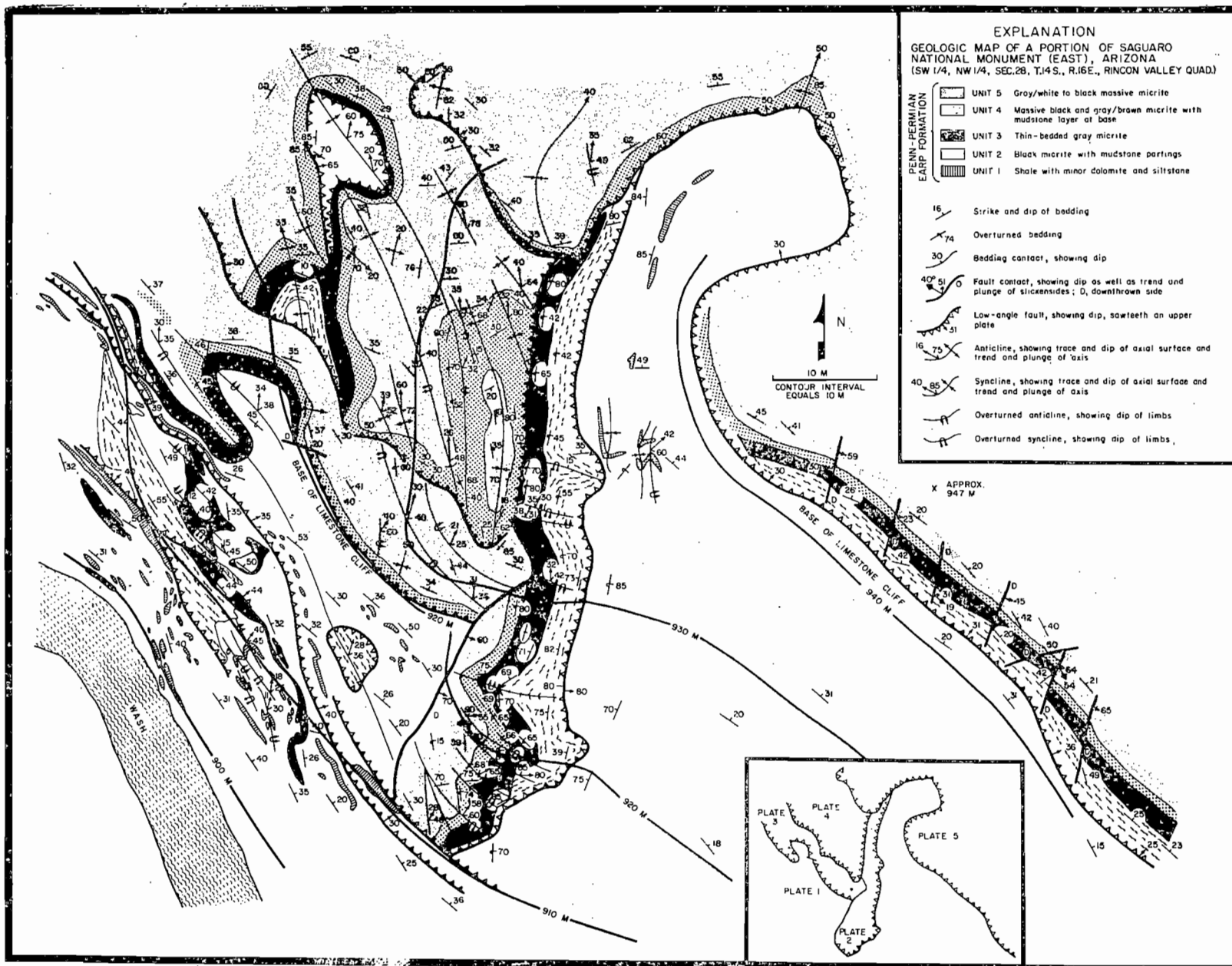


Figure 26. Map of folds in detachment at Saguaro National Monument (Davis and Frost, 1976).

northeast-directed Laramide regional overthrusting.

The gravity-induced folding model of Davis (1975) is here reevaluated in the light of new data amassed over the past five years. Folds in these rocks may well be the result of southwest-directed simple shear, not radially directed gravity gliding. Simple shear and attendant folding accompanied, at least in part, the development of lineated mylonitic tectonite.

In effect, the orientation of all folds in Paleozoic and Mesozoic tectonite and detachment rocks in the Rincon Mountains area are compatible with a model of overall southwestward directed simple shear. This is most keenly reflected in orientations of axial planes of all folds. The modal strike of axial planes is at right angles to penetrative lineation in mylonitic tectonite. Furthermore, the dip of axial planes is consistently northeastward. Such a dip-direction vergence, strongly supports southwest-directed shear. Unlike the axial planes, the axes of the folds show differences from domain to domain, but the axes orientations coupled with style of asymmetry are consistent with southwestward movement. Within the overall southwest-directed movement pattern, there tends to be southwestward divergent flow, with components of movement down the dip of the major antiformal arches like Tanque Verde antiform. This divergence contributed to the "apparent" radial pattern which Davis (1975) emphasized.

Many kinds of relationships can be cited which document the parallelism of movement plan in tectonite and detachments. The décollement zone in the Coyote Mountains contains slickenside striae which are parallel to the mineral lineation in underlying tectonite. Large grooves mapped by Drewes (1978) in the décollement zone of the southern Rincon Mountains near Colossal Cave and on the nose of Tanque Verde Mountain trend $N60^{\circ}-70^{\circ}E$, parallel to mineral lineation in underlying tectonite. The grooves and mullion are likewise parallel to slickenside striae in granite and limestone in the southern Rincon Mountains (Davis and others, 1974). Also, sedimentary strata in detachments, like those in Happy Valley on the eastern side of the Rincon Mountains, strike north-northwest on the average, perpendicular to the east-northeast transport direction.

To sum up, the presence of upper Paleozoic detachment strata atop microbreccias and mylonitic tectonite derived from Precambrian protolith is the net effect of sustained shearing and faulting along a common line of movement. Whether these distinctly different structural assemblages were produced during a continuum of deformation or as a result of discrete superposed events is not presently known. The mylonitic gneisses represent a deep exposure of a regional shear which accommodated normal-slip displacement in a southwestward direction. Not only was Precambrian plutonic rock converted to mylonitic tectonite, but overlying sedimentary rocks of younger Precambrian and Paleozoic age were also converted to tectonites. During prograde metamorphism of the sedimentary rocks, southwestward-directed simple shear resulted in folding and partial transposition of the sequence. The result of the shearing was profound vertical flattening and east-northeast extension (elongation) of all rocks in the zone of tectonite formation. The net effect of the thinning and flattening was a lowering of strata which had been in upper structural positions (e.g., upper Paleozoic strata or Cretaceous strata) to closer proximity with mylonitic tectonite derived from Precambrian protolith (Davis, 1977). At some stage during the change from moderately ductile to moderately brittle deformation, earlier formed mylonitic tectonite was itself disrupted by the shearing process. As a consequence, the décollement zone was formed as a zone of coalescing imbricate faults. The presence of hydrothermal waters during retrograde metamor-

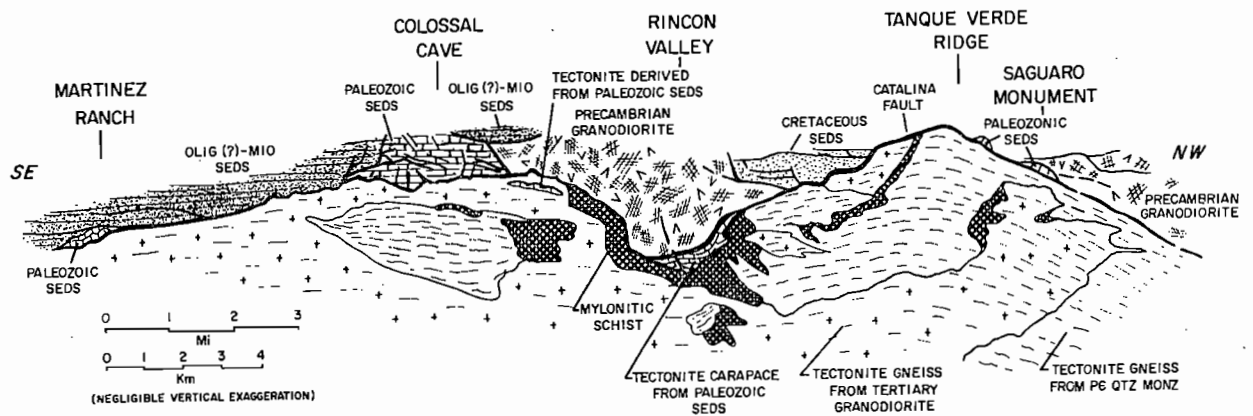


Figure 27. Down-structure view of geology of Rincon Valley and Rincon Mountains.

phism, microbrecciation, and shattering resulted in the thorough, chloritic alteration. The final emplacement of detachment strata onto the top of the décollement zone carried already-folded strata along listric(?) normal faults. Some newly formed folds may have developed as well, but this is difficult to demonstrate. The final emplacement of detachment strata occurred in mid-Miocene time, apparently between 20 and 15 m.y. ago.

Awareness of the kinematics of southwestward-directed movement permits the Rincon Mountains metamorphic core complex to be viewed in an illuminating way. Using J. Hoover Mackin's (1950) down-structure method for viewing geological maps, the map of the Rincon Mountains, prepared by Drewes (1978), can be converted into a structural profile. This is shown in Figure 27. The décollement can be seen to be a broadly arcuate mullion-like fault surface separating mylonitic tectonites below from a variety of upper-plate detachment rocks above. Mylonitic tectonites are seen to be "thin" lens-shaped structural units, some of which are reduced to mylonitic schists at their edges and terminations. On Tanque Verde ridge, the gneisses derived from Tertiary granodiorite climb as a tabular body up-structure, perhaps revealing a distorted vestige of the original intrusive relationship of the granodiorite into Precambrian quartz monzonite. The strong tectonite overprint is such that foliation and lineation cut through the granodioritic gneissic phase discordantly. The deformation thus was passive.

The major sites of formation of mylonitic schist are at contacts between dissimilar gneissic phases and at the deep mullion-like trough in the Rincon Valley re-entrant. The symmetrical distribution of mylonitic schist about the Rincon Valley trough suggests that strain was preferentially concentrated along this "keel" of the detachment system.

Structural units in the upper-plate detachments, whether they be composed of sedimentary or volcanic rocks, are thin and lensoidal, thoroughly pervaded by extensional faults and tear faults. From Tanque Verde ridge southeastward, four detachment domains are recognized: these contain Cretaceous sedimentary rocks, Precambrian Rincon Valley granodiorite, Paleozoic sedimentary rocks, and Oligocene (?)–Miocene sedimentary rocks. Strain styles within the detachments are dissimilar: Paleozoic and Cretaceous strata display exquisite overturned to recumbent folds; Precambrian granodiorite is so shattered that "it would be impossible to find within it a headstone for a grave" (E.B. Mayo, pers. comm., 1972); and the Oligocene (?)–Miocene strata display faulting and rigid-body rotation to northwesterly strikes. In Rincon Valley (Drewes, 1978), the detachment of Rincon Valley granodiorite rests on tectonite derived from Paleozoic rocks, remnants of earlier ductile penetrative simple-shear deformation. Just southeast tectonite gneiss derived from the Tertiary granodiorite contains interlayers of tectonite derived from Paleozoic impure carbonates and quartzites (Drewes, 1978) (see Fig. 27). These probably are overprinted roof pendants, disclosing that the Tertiary granodiorite was originally intruded to structural/stratigraphic levels above the Precambrian basement.

Although the Rincon Valley granodiorite has been interpreted to be a Laramide thrust plate (Drewes, 1978), Figure 27 reveals that it can be viewed simply as one of a variety of country-rock elements that were emplaced by detachment faulting as young as mid-Miocene. Due to frictional shear as the detachments were emplaced, narrow antiformal crests of mullion ridges like Tanque Verde Mountain, and the smaller mullion counterparts in the vicinity of Colossal Cave (Fig. 27), were subjected to high stress and attendant development of cataclastic fabrics, especially microbreccia and "ultramylonite".

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GEOLOGY AND MINERAL DEPOSITS OF SOUTHERN SONORA
AND THE SONORAN SIERRA MADRE OCCIDENTAL

FIELD TRIP No. 11 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 28-30

TRIP LEADERS:

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ROADLOG FOR THE FIELD TRIP:
GEOLOGY AND MINERAL DEPOSITS OF SOUTHERN SONORA
AND THE SONORAN SIERRA MADRE OCCIDENTAL

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GENERAL STATEMENT

The purpose of this field trip is to examine the Paleozoic, Mesozoic and Cenozoic stratigraphy east of Hermosillo. Particular attention will be given to correlation and chronology of the ore deposits and ash flows with discussion of ideas concerning the tectonic framework. The field trip commences in Hermosillo (at an elevation of about 230 meters above sea level), located within the Sonoran Desert part of the southern Basin and Range, and progresses eastward through the Parallel Ranges and Valleys to Yecora. Located within the western flank of Sierra Madre Occidental, Yecora (elevation 1,580 meters) is only eight kilometers from the Chihuahua state line. The land surface gradually rises from west to east and the traverse illustrates how the climate, vegetation, human habitation and industries are related to the geology and physiography of the land over which the road has been built.

In the Sonoran Desert region, the topography is mature where the mountain ranges are small, irregular and separated by relatively broad, gently sloping and extensively alluviated plains in the valleys. At the flank of the Sierra Madre Occidental the ranges are rugged and precipitous, separated by narrow canyons.

In the Sonoran Desert region the Paleozoic carbonate and clastic rocks were deposited on a stable cratonic platform. The Mesozoic was a time of continental sedimentation. Large masses of granitic intrusive rocks invaded parts of the area during the Laramide orogeny and much of the Laramide volcanic cover was removed from western Sonora as a result of epeirogenic uplift and denudation. This was followed by crustal disturbance and the resulting basins were filled with continental clastic sediments. Extensive Mid-Tertiary ash flows were erupted to create the Sierra Madre Occidental. Relicts of some of these flows are found as far west as the Hermosillo area. Late Cenozoic block faulting created most of the present day basins and ranges. Locally derived clastic sediments filled most of the down-dropped basins, while evaporates formed in some of the basins.

The route covered is shown in Figure 1 and the stratigraphic section in Figure 2. Not all of the stratigraphic units occur at the same localities. You will get a glimpse of many wonderful outcrops. Unfortunately, the great distance to be traveled and poor condition of some of the roads restrict the number of stops along the way and also preclude a truly detailed examination of this complex geology.

The distance traversed along the field trip route is given in the left-hand margin as cumulative miles (kilometers in parentheses). The interval between each point of discussion is recorded in the center of the page.

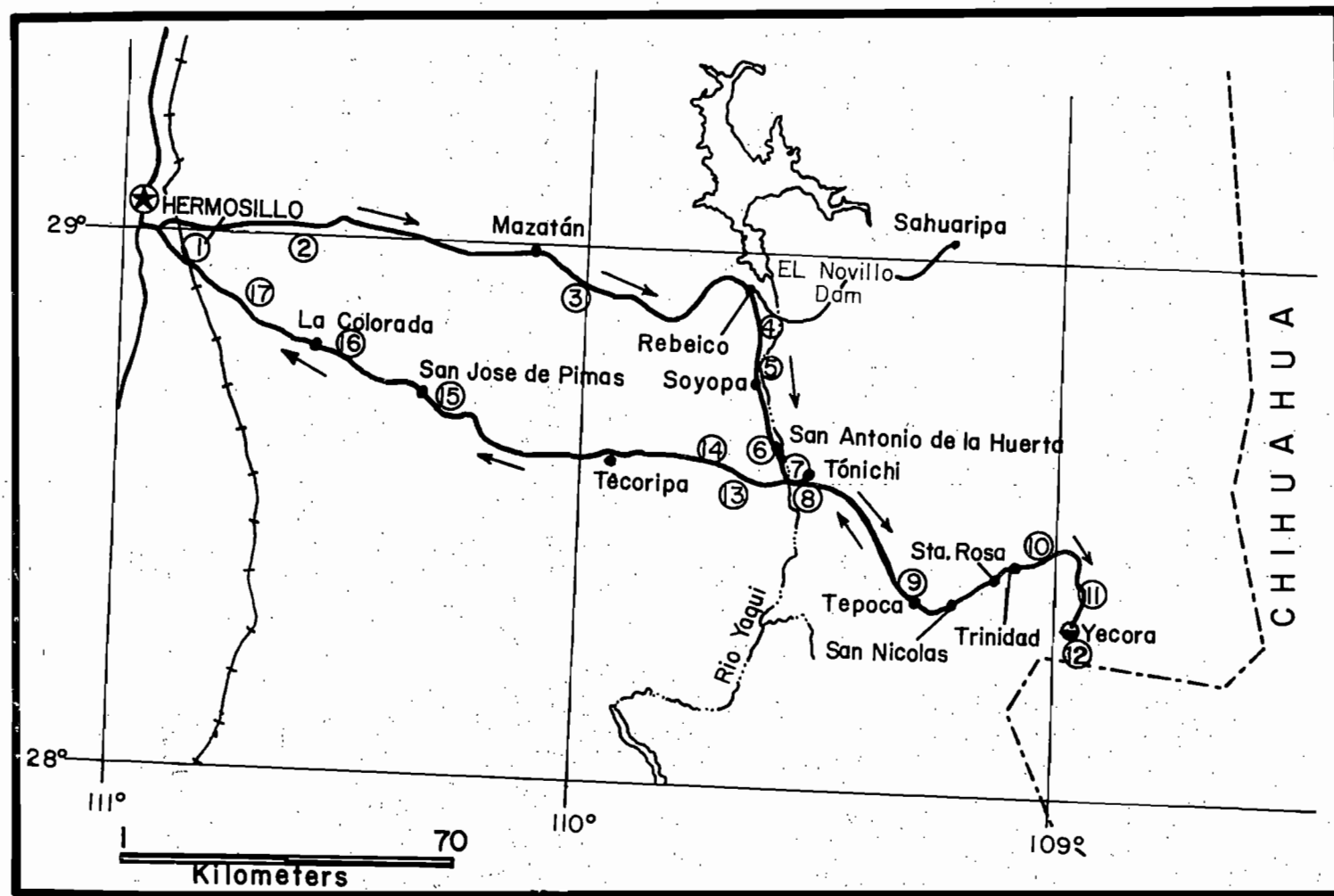
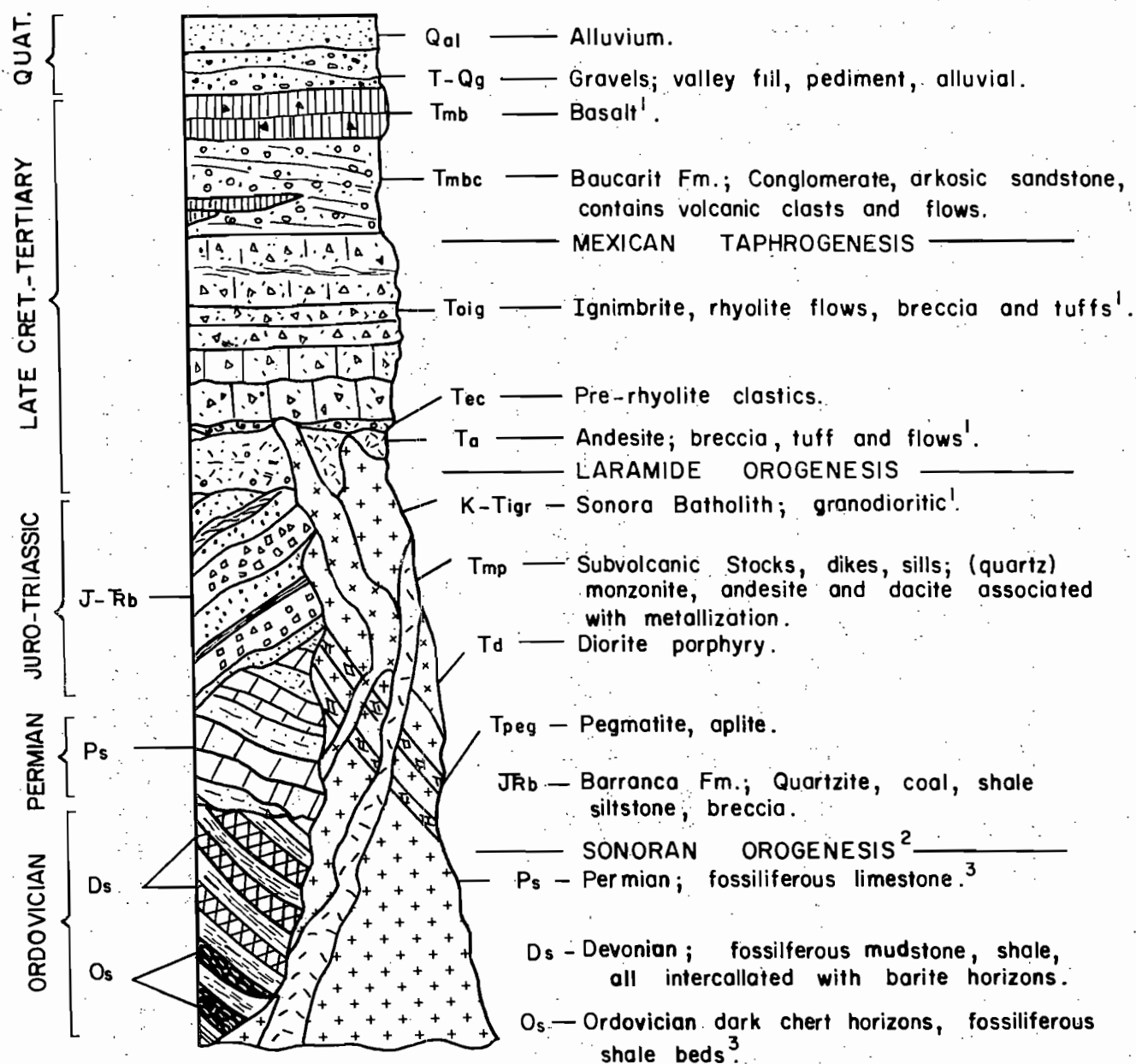


Figure 1. Index map of the field trip route between Hermosillo-Yecora-Hermosillo. Circled numbers indicate stops - discussed in the text.

PERIOD

LITHOLOGY / TECTONISM



1. - See Clark and others (1979) for distribution in Sierra Madre Occidental.

2. - See Fries (1962) for further details.

3. - Pz, Paleozoic undifferentiated on structure section.

Figure 2. Composite stratigraphic column along the excursion route between Hermosillo-Tecoripa-Tonichi-Yecora.

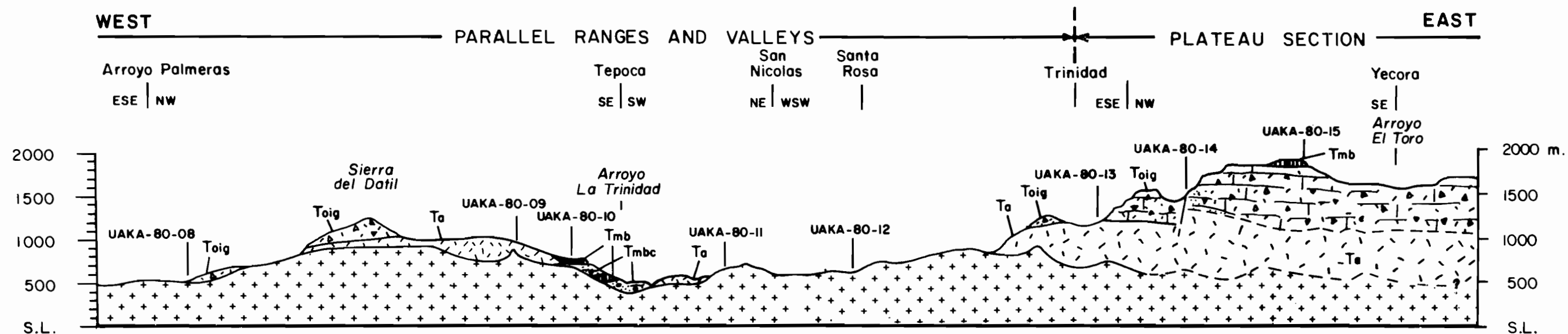
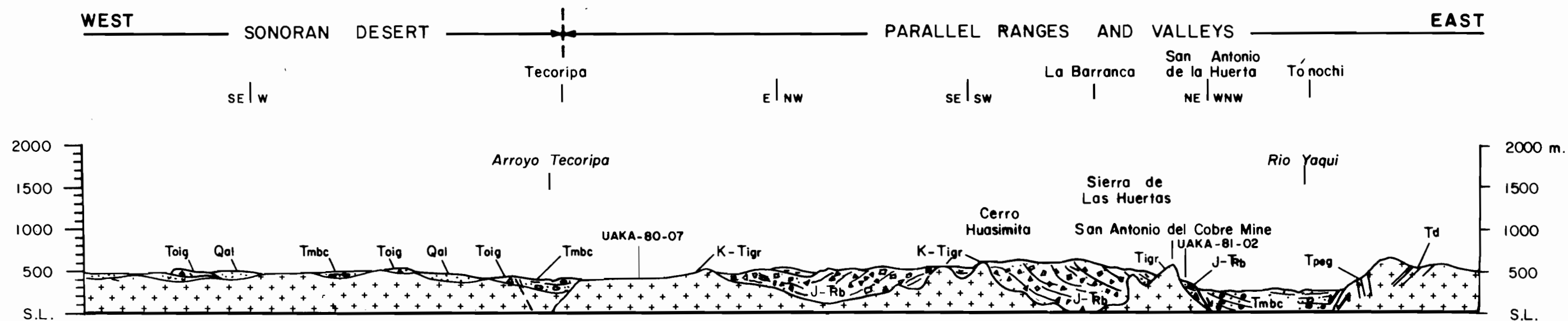
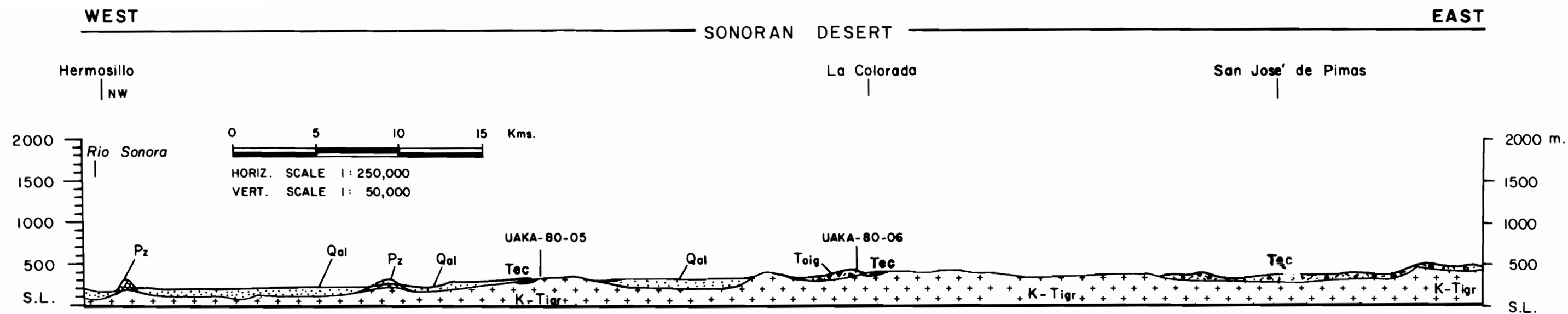


Figure 3. Simplified structural section between Hermosillo and Yecora.

373.
Schedule

Day 1

Stop No.	Field Trip Route Distance from Hermosillo		Arrival	Departure	Highlight
	Mile	Kilometers			
0	0	0	-	8:00	Hermosillo, starting point.
1	19.4	31.2	8:30	10:00	Cement Plant. Paleozoic limestone
2	31.3	50.4	10:30	11:00	Triassic-Jurassic Barranca Formation (?). Deformed sediments
3	67.0	107.8	11:45	13:30	Barison Barite operation Lunch after discussion
4	89.7	144.4	14:30	14:45	Baucarit Formation sediments
5	104.5	168.2	15:00	15:30	Flow in Baucarit Fm. near Soyopa
6	119.2	191.8	16:15	18:30	San Antonio de la Huerta visit Cu-U, Ag-Au operations
Overnight			18:30		San Antonio de la Huerta mine bunk house. Dinner, informal rap session, and then relax until morning.

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Day 2

Stop No.	Field Trip Route Distance from Hermosillo		Arrival	Departure	Highlight
	Mile	Km			
Breakfast			6:30	8:00	Rise, shine and enjoy the breakfast
Start	119.2	191.8		8:00	Leave San Antonio de la Huerta
7	125.3	201.6	8:30	9:30	Complex road cut exposing Triassic to Pleistocene rocks
8	125.8	202.5	9:35	10:00	Type Baucarit Fm. sediments
9	153.5	247.0	10:45	11:15	Baucarit Fm. sediments with overlying younger volcanics and sediments
10	174.0	280.0	11:45	12:15	Eocene sediments and overlying ash flow uff sequence
11	184.0	296.1	12:45	13:00	Mesa capping flow
12	190.4	306.4	13.30	15.30	Lunch, discussion, photography
Over-night	266.7	429.2	??		Return to San Antonio de la Huerta. Dinner and informal discussion

If the weather and road conditions are favorable, we will have a number of quick photo stops during the return trip from Yecora to San Antonio de la Huerta.

375
Day 3

Stop No.	Field Trip Route Distance from Hermosillo		Arrival	Departure	Highlight
	Mile	Kilometers			
Breakfast			6:30	8:30	Rise, shine, enjoy break- fast
	266.7	429.2		8:00	Resume journey
13	272.9	439.2	8:30	9:00	Barranca Fm. good plant fossil area
14	280.7	451.7	10:00	11:00	Landslide breccia in Barranca Fm.
15	314.7	506.5	12:00	12:30	Eocene sediments at road cut near Rancho San Francisco
16	335.5	539.9	13:00	13:30	Tertiary sediments
17	350.4	563.9	14:00	14:30	Granita Hermita-granodiorite phase
18	373.8	601.6	15:00		Libar, comer y charlar

FIRST DAY

Saturday, March 28, 1981

Summary:

We will visit a new cement plant in Permian Limestone and a controversial section of red arkose and shale of possible Triassic-Jurassic Barranca Formation affinity. This will be followed by a lengthy stop at the new Barison deposit with a 3 million ton barite reserve. The barite is reported to be Ordovician in age and appears lithologically identical to Devonian barite deposits in Nevada. Following lunch, we will visit extensive Baucarit Formation sediments. On the approaches of the mining area in San Antonio, we will pass through Barranca shale intruded by diorites. The last visit will be the Mina El Cobre (copper, uranium), Mina Promentorio (gold, silver) and alluvial gold deposits at San Antonio de la Huerta.

Driving distance:	119.2 miles - 191.8 kilometers
Logged distance:	119.2 miles - 191.8 kilometers
Starting time:	8:00 A.M.
Assembly point:	Parking lot, Valle Grande Motel
Number of stops:	6
Overnight accommodation:	Stretch your sleeping bag in mine bunk house or outside, if weather permits.

FIRST DAY

(From Hermosillo)	Interval
Cumulative	Mi. (Km.)
Mi. (Km.)	0.0 (0.0)

Hermosillo is a bustling modern city with a population of about 200,000. It was named after General José M. González de Hermosillo and has been the seat of State government since 1879. It is the center of an agricultural area created by irrigation waters from the nearby Rodriguez Dam on the Sonora River that flows through the city. It houses many State and Federal government and corporate offices, as well as the University of Sonora.

Main rock types that constitute the ranges in the Hermosillo area are Paleozoic limestone with subordinate sandstone intruded by Laramide (?) granitic to dioritic rocks. The elevation of the ranges usually do not exceed over 300 meters above the valley level.

0.0 (0.0) Starting point: Valle Grande Hotel. Proceed east on Kino Blvd. (Carretera Internacional).

0.5 (0.8)

0.5 (0.8) Turn south at Glorieta

0.4 (0.6)

0.9 (1.4) PEMEX storage plant.

0.4 (0.7)

1.3 (2.1) Paleozoic limestone in the hills to the southwest.

0.6 (1.0)

1.9 (3.1) Cross drainage ditch.

0.1 (0.1)

2.0 (3.2) Reservoir Dam at 9 o'clock; Presa Abelardo Lujan Rodriguez.

0.6 (1.0)

2.6 (4.2) Microwave Tower at 3 o'clock on Cerro La Campana, spillway at 9 o'clock.

0.2(0.3)

2.8 (4.5) Laramide (?) granite intruded by aplite and monzonite dikes.

Reservoir at 9 o'clock.

0.7(1.1)

3.5 (5.6) Cement plant at 3 o'clock.

0.2(0.4)

3.7 (6.0) Railroad track crossing

0.2(0.3)

3.9 (6.3) Paleozoic Limestone quarry at 3 o'clock.

0.6(1.1)

4.5 (7.2) State prison at 3 o'clock.

0.6(1.1)

5.1 (8.2) Stockyard at 9 o'clock.

0.5(0.8)

5.6 (9.0) Junction; turn left (to east) on road to Sahuaripa. Stockyard at 9 o'clock.

1.3(2.08)

6.9 (11.1) Animal waste dump at 3 o'clock. Hold your nose!!

1.9(3.1)

8.8 (14.2) Crossing railroad track.

0.3(0.4)

9.1 (14.6) Lime plant at 2 o'clock. Limestone quarried from Paleozoic sections.

2.7(4.4)

11.8 (19.0) Road junction; keep straight.

0.9(1.4)

12.7 (20.4) Roadcut in Paleozoic limestone.

0.9(1.5)

13.6 (21.9) Santa Teresa Mountains at 3 o'clock are composed of Paleozoic limestone.

0.6(1.4)

14.2(22.8) Pig pens at 9 o'clock.

0.5 - 0.7 (0.8 - 0.1)

15.3(24.6) Bedded limestone, dips 70° SE.

STOP 1

4.1(6.6)

19.4(31.2) Entrance to new limestone cement operation

Mr. Anthony Claim, Manager, Cementos Portland Nacional will explain the geologic framework of the limestone quarry located in Sierra de los leyva. The Paleozoic sequence consists of a succession of limestone with minor conglomeratic beds, lenses of dolomite, siliceous limestone and nodular chert. The beds dip 30° to SSE. Exploration and drilling programs indicate that the limestone constitute 89% of the sequence. Reserve at Cerro Chico and El Espolon are 50 and 400 million tons respectively. Exploitation work, in progress, at Cerro Chico has the following over all grades:

56% Limestone, less than 10% silica

32% Limestone, 10 to 30% silica

12% waste

The quarries are in cavernous Paleozoic limestone. Caves encountered during drilling resulted in some loss of the drill steel. It produces one million tons of cement a year. Drive to the cement plant for geologic discussions. Return to the intersection and follow the road log.

3.5(5.6)

22.9(36.8) Passing through hills. Pyroclastic flows cap Paleozoic limestone. Similar flow yields Oligocene K-Ar age in La Colorada area to the south.

1.4(2.3)

24.3(39.1) Road to Las Norias Ranch. Begin ascent through pass in the Sierra San Juan de Dios. Reportedly, in the San Juan de Dios Mountains, Laramide (?) granodiorite intrudes Paleozoic limestone.

2.1(3.4)

26.4(42.5) Outcrop of Paleozoic limestone.

0.7(2.1)

27.1(43.6) Sharp curve to the left.

0.5(0.8)

27.6(44.4) Outcrops of conglomerate rich in granodiorite cobbles and boulders.

1.7(2.8)

29.3(47.2) Tilted conglomerate with limestone cobbles.

1.2(1.5)

30.5(49.1) Clastic sediments in fault contact with underlying fossiliferous limestone (crinoid rich). Dip of limestone 70° SE, strike 50° NE.

STOP 2

0.8(1.3)

31.3(50.4) Highly deformed section of conglomerates and shales, appear to be redbeds but hydrothermally altered. Note limestone cobbles in the conglomerates. According to Radelli (personal communication) these sediments are considered to be Triassic-Jurassic Barranca equivalent and have traditionally been assigned to that group.

Radelli suggested that there may be some confusion between the Permian and Triassic in this area.

1.0(1.6)

32.3(52.0) Turn-off to El Aquarito.

1.0(1.6)

33.3(53.6) Pink and buff-colored agglomerate with sand-size matrix and cobbles and boulders of the volcanic rock overlain by porphyritic dacite.

1.0(1.6)

34.3(55.2) Sierra Mazatán at 11 o'clock. It is a metamorphic core complex, 58 m.y. (Anderson et al., 1980; Coney, 1980), roughly circular with 7.5 mile (12 Km.) diameter.

0.3(0.5)

34.6(55.7) Agglomerate with clasts of vitrophyre.

10.5(16.9)

45.1(72.6) Leucocratic banded gneiss within Sierra Mazatán core complex at 9 to 11 o'clock. The highest point is about 1,091 m. above valley floor. The range is a prominent landmark.

2.7(4.3)

47.8(76.9) Cerro Cobachi at 2 o'clock with about 697 m. relief.

The Cobachi barite property is 4,000 hectares in areal extent, and extends over a region 7 by 2.5 kilometers. According to Dr. Guillermo A. Salas, the barite sequence consists of chert at the base, barite with clay, a sandstone-limestone marker bed, about a meter thick, massive barite with some quartz and baritic sandstone at the top. The sequence overlies Ordovician chert and siltstone with Upper Ordovician graptolites. Devonian brachiopods are replaced by barite. The average grade is about 3.8 gms per cubic centimeter which will require upgrading to 4.3 gms/cc by removal of silica. The reserve is about 5 million tons.

According to Dr. Salas and Mr. Rod Asher, the stratigraphy and fossil assemblage is correlative with some facies of Nevada barite deposits. Hence, they consider it also to be a Devonian marine deposit. However, minor metamorphism, recrystallization, replacement of the original rock and fossils do not preclude a post-Devonian hydrothermal origin for the deposit.

A granodiorite intrusive has thermally metamorphosed a thick carbonate section with identifiable Leonardian fusilinids. Biotite and hornblende from the granodiorite of the eastern slope of a hill west of the south branch of Cobachi to La Vuelta Colorada road yield 66.7 ± 1.4 m.y. and 65.9 ± 1.4 m.y. ages, respectively, indicating Laramide plutonism in the barite district.

1.8(2.9)

49.6(79.8) Turn-off to Rancho Las Calaveras.

0.8 (1.3)

50.4(81.1) Baucarit-like steeply tilted conglomerate.

0.2 (0.3)

50.6(81.4) Vado - Quizuani wash.

3.0 - 4.7 (4.9-)

53.6(86.3) Granite, presumably Laramide in age, is thoroughly weathered.

It is cut by a one meter thick mafic dike.

0.6 (.96)

55.9(90.0) Enter Mazatán. Population 2,500. Keep right on road to Sahuaripa.

2.4 (3.8)

58.3(93.8) Turn-off to Cobachi barite deposit.

5.3 (8.6)

63.6(102.4) Turn-off to Tecoripa.

0.7 (1.1)

64.3(103.5) Finely-bedded gravels in roadcut.

1.5 (2.4)

65.8(105.9) Barison (Barite of Sonora) barite deposit at 1 o'clock.

1.1 (1.8)

66.9(107.7) Redbeds, variegated shale with interbedded limestone lenses.

0.1 (0.1)

67.0(107.8) Entrance to (Barison) barite deposit.

STOP 3

Barite of Sonora

The following description has been written by Ing. Juan Fernando Oliva B., who will also act as guide in the mine area:-

Stratigraphy:

The Ordovician is exposed in an area 7 km in width by 30 km in length, lying below an angular unconformity with Permian strata, but intruded by a granitic pluton. In the central portion of this area there are small structures which, for the most part, are covered by alluvium that masks the contact with the Permian sequence which is found toward the east side. The Ordovician Devonian strata is composed of shales, mudstones, chert and conformable Devonian barite horizons with thicknesses of 0.05 - 15 m and which is light-gray in color. These horizons are exposed in small hills named Saguaro, Las Torres, Las Papas, and Virgen, and others numbered 3, 4, 5, 9 and 10.

The Permian is found toward the northeast of the area in Cerro de Los Rastros as an erosional remnant and consists of reef

limestone with a crinoid, fusilinid and a bivalve fauna.

Tertiary strata are well exposed on the west and southwest part of the area with angular discordance with the Ordovician sediments. The Baucarit Fm. is composed of conglomerates, sandstone and some shale.

Igneous rocks located to the northeast and relatively near the studied area, are found in a pluton which has been described as a biotite granite which may be of Cretaceous age. Generally intruding the Ordovician sequence and found in large exposures to the southeast, outside of the studied area, is a hornblende diorite.

Structure:

The Ordovician exposures in the area are found in beds that strike NE - SW with dips to the southeast, varying to E - W strike with southerly dip. The strata have been subject to great compressive forces from the southeast and south which gave rise to a great number of faults that for the most part have NW - SE strike and also reverse faults that are perpendicular to the others.

Genesis:

Because the barite is found in strata that are concordant with the sedimentary sequence, it is considered to be syngenetic.

The barite nodule area shows nuclei that are possibly clasts which, because of the large amount of sulfate in the basin, became coated and formed barite nodules.

The basin must have been enriched in barium sulfate to the extent that thick strata of barite were deposited due to high sulfate

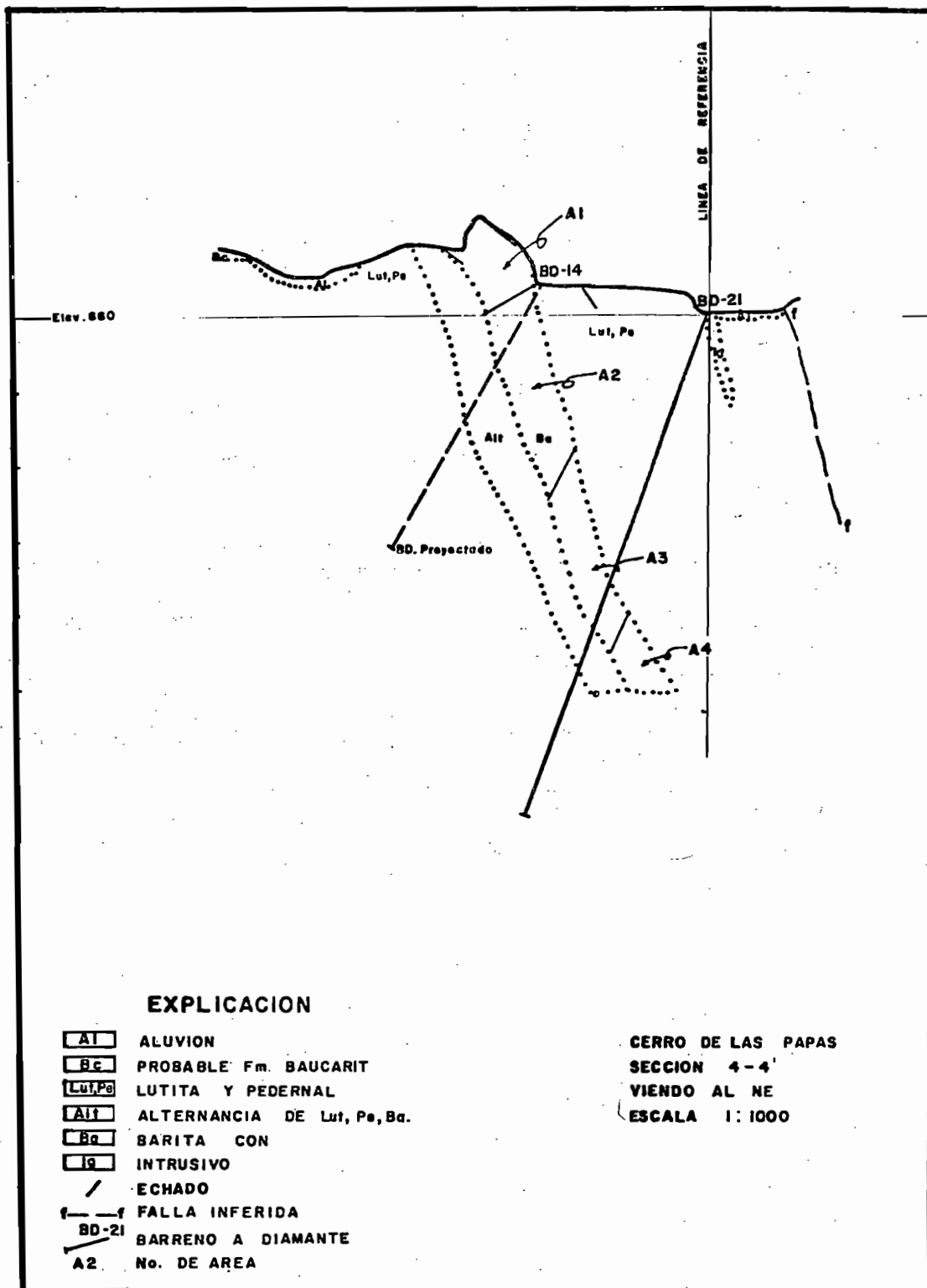


Figure 4. Cross-section of the Barison barite mine.

content without addition of ferrugeneous material nor silica precipitation within the barite layers.

Mining:

Two stages of exploitation are employed:

- (1) Extraction of mineral of low or high quality but always of the massive variety.
- (2) Extraction of similar mineral that is located in alternating sequences with other sediment or in nodules.

Beneficiation:

There are three types of mineral; namely, massive, alternating type and in nodules. The massive variety is nearly pure and is directly treated in the mill. The alternating type, because it contains horizons of shale and chert, will first be treated by jigs and then by flotation with different reagents. The nodular ore will also be treated in jigs and by flotation.

Reserves:

In bulk -

Proven	-	3,007,884 metric tons
Probable	-	875,500 metric tons
Possible	-	500,680 metric tons

By variety -

Massive barite	≥	4.20 (av. 4.2856) - 533,522 m. tons
Massive barite	≤	4.20 (av. 3.91) - 1,259,598 m. tons
Alternating		(av. 3.76) - 1,057,517 m. tons
Nodules		(av. 3.36) - 146,967 m. tons



Figure 5. Ordovician strata. Incompetent shale is deformed while black chert is relatively undeformed.

0.5 (0.8)

67.5(108.6) Medium, black, tilted, bedded limestone in roadcut on south side of road.

0.3 (0.6)

67.8(109.2) Pull-off siding on right-hand side of road. Excellent outcrop of black chert and shale. The shale is mauve, pinky-yellow to grey and incompetent to compression, yielding a fold that overturned to the west. Also, some minor normal faulting with gouge zone. The shale is overlain by a chert bed with thin shale seams varying in thickness from 5 cm to 25 cm. This is part of the Ordovician Devonian strata where the bedded barite is found above.

0.5 (0.7)

68.3(109.9) Arkosic sediments.

0.5 (0.8)

68.8(110.7) Thick sequence of interbedded chert and shale.

0.6 (1.0)

69.4(111.7) Almost continuous outcrop of above sediments to this point.

1.9 (2.1)

70.7(113.8) Arkosic redbeds with oblate-spheroid chert nodules.

0.5 (0.8)

71.2(114.6) Interbedded shale and chert with vertical attitudes.

0.3 (0.5)

71.5(115.1) Thick shale sequence.

0.1 (0.3)

71.7(115.4) Turn off to Las Guasimas.

0.1(0.1)

71.7(115.4) Granitic regolith with clasts of aplitic dikes which resist weathering more effectively than the granites that form a grus.

0.2(0.3)

71.9(115.7) Granite, probably Laramide, with light-colored aplite dikes. Sierra de Martinez at 9 o'clock and Sierra de La Cienega at 3 o'clock. Granite is exposed for the next four miles.

0.1(0.1)

76.2(122.6) Conglomerate with granitic cobbles and pebbles, resulting from denudation of batholith in post Laramide - pre-Oligocene time (Eocene erosion surface).

0.2(0.8)

76.7(123.4) Road to Rancho El Represito.

0.6(1.0)

77.3(124.4) Andesite, fractured and altered with epidote exposed on both sides of the road. It is Laramide (?) in age.

For the next 1.4 miles conglomerate overlies andesite and represents Eocene denudation and deposition.

1.5(2.4)

78.8(126.8) Oligocene pyroclastic ash flow tuffs with large pumice clasts; buff colored, has a northwesterly dip. The tuff overlies a bleached and altered ash bed.

1.0(1.6)

79.8(128.4) Turn off to Llano Colorado on left. Subvolcanic porphyritic monzonite is intruded by breccia dikes with epidotization and quartz veins. Sample has been collected for dating (UAKA-81-01).

0.8 (1.3)

80.6(129.7) Brecciated subvolcanic monzonite with dikes (highly shattered, still in subvolcanic rock with intense oxidation).

0.8 (1.3)

81.4(131.0) Conical hill at 2 o'clock is brick red due to alteration. Subvolcanic monzonite outcrop continues for next half-mile.

0.6 (1.0)

82.0(132.0) Leucocratic pyroclastic flow with large pumice fragments.

0.1 (0.1)

82.1(132.1) Take dirt road on the right (south) to Rebeico, Soyopa and San Antonio de La Huerta.

1.0 (1.6)

83.1(133.7) Pyroxene-andesite, altered and fractured. Road descends to Arroyo Tepustate. Valley fill is typical Baucarit Fm. forming conspicuous bluffs.

0.9 (2.1)

84.4(135.8) Arkosic conglomerate of Baucarit formation overlain by river gravels. The thin beds dip 45° eastward and strike N 10° E. Rows of eucalyptus trees welcome a visitor to Rebeico village. Bear right in Rebeico.



Figure 6. Arkosic conglomerate of Baucarit Fm in the road cut. Note the tilted competent Baucarit beds to the east. Outcrop at mileage 84.4.

0.8 (1.3)

85.2(137.1) Gravel-covered river terrace across the valley at 7 to 9 o'clock.

0.5 (0.8)

85.7(137.9) Arkosic Baucarit Fm. contains clasts of volcanic rocks.

1.7 (1.15)

86.4(139.0) Cattle guard. Sierra Palo Verde to the east (left) and Sierra Las Chicuras to the west. For the next three miles the road is along a ridge top.

1.9 (3.1)

88.3(142.1) Cattle guard.

0.9 (1.4)

89.2(143.6) Baucarit Fm. at 11 o'clock in middle distance.

0.3 (0.4)

89.5(144.0) Baucarit conglomerate and arkosic sandstone. Top of exposure contains boulders of volcanic rocks up to 30 cm in diameter within arkosic sandstone matrix.

STOP 4

0.2 (0.4)

89.7(144.4) Andesite overlain by Baucarit arkosic sandstone. Thick sequence of Baucarit is visible in bluffs up a small valley to the east.

1.0 (1.7)

100.7(162.1) Cattle guard.

2.4 (3.8)

103.1(165.9) Young gravel overlies Baucarit Fm.

0.3 (0.5)

103.4(166.4) Arroyo.

0.2 (0.3)

103.6(166.7) Baucarit Fm. conglomerate takes on distinct red color.

STOP 5

0.9 (1.5)

104.5(168.2) Rhyodacite flow interbedded within Baucarit Fm. Good view of Soyopa in the Rio Yaqui valley which is about 4 km wide. Baucarit dips 16° due east.

0.1 (0.1)

104.6(168.3) Pull-off area. Cerro Picacho to the east.

1.2 (2.0)

105.8(170.3) Baucarit Fm. conglomerate at 2 o'clock.

0.9 (1.4)

106.7(171.7) River gravel overlies Baucarit Fm. Take right fork to San Antonio de la Huerta.

0.8 (1.3)

107.5(173.0) Andesite agglomerate interbedded within tilted Baucarit Fm. sediments.

0.4 (0.6)

107.9(173.6) Pyroclastic ash flow laced with calcite veins and altered to brown clay.

0.5 (0.8)

108.4(174.4) Hydrothermally altered monzonite. Individual grains are difficult to recognize due to alteration.

0.4 (0.7)

108.8(175.1) Brecciated and altered monzonite porphyry with well-developed limonitic soil horizon for 1 mile.

2.0 (3.2)

- 110.8(178.3) Sierra El Aliso to the west and Rio Yaqui to the east. Altered intrusive, dioritic breccia in road cuts.

0.9 (1.5)

- 111.7(179.8) Deformed black shale cut by subvolcanic dikelets altered to clay; limonitic soil. Black shale exposure continues for 0.4 miles.

1.6 (2.5)

- 113.3(182.3) Llanos Plomosas.

0.2 (0.4)

- 113.5(182.7) Altered monzonite breccia.

0.6 (0.9)

- 114.1(183.6) Altered and brecciated diorite porphyry.

0.5 (0.8)

- 114.6(184.4) Brecciated and altered diorite porphyry overlain by poorly sorted conglomerate channeled into the porphyry.

2.2 (3.6)

- 116.8(188.0) Brecciated and altered monzonite porphyry with limonitic stains.

1.4 (2.2)

- 118.2(190.2) Cattle guard.

0.1 (0.2)

- 118.3(190.4) Brecciated monzonite.

0.3 (0.5)

- 118.6(190.9) Take right fork towards Promontorio and Luz del Cobre mines.

The left fork branches to the Pueblo of San Antonio de la Huerta.

0.3 (0.5)

118.9(191.4) Porphyritic diorite breccia.

0.3 (0.4)

119.2(191.8) Mine work shop. Turn right to mine office.

STOP 6

Visit El Cobre Mine and dredging operation.

San Antonio de la Huerta Mining District
(Data furnished by Luz del Cobre, S.A.).

History of Development:

The San Antonio de la Huerta area began to produce gold around 1840, whereas copper was first mined in significant quantities from 1920-1940. According to Lic. Manuel R. Quiroga M. (Personal communication), at the turn of the century some 18,000 inhabitants were living in this locale. Since then the population has been much reduced, although recent developments of the mineralized zones suggest that more people may be attracted to the area in the future as employment opportunities materialize.

In the last few years geophysical surveys included airborne magnetic and ground induced polarization resistivity methods. When geologic data was also considered it was shown that the electrical methods delineated the shapes and positions of the sulfide zones fairly accurately. However, drilling these zones proved that only the main breccia pipe contained values of commercial grade and tonnage on the copper-bearing property. Uranium has also been detected and will be supplied to the federal government (as part of the National Reserve). The copper is recovered by a large-scale in situ leaching process. The ore body indicated in the accompanying diagrams has been prepared for this type of recovery by ALCOA but has not been blasted to date.

In an adjacent arroyo a dredging operation was successful in recovering placer gold from stream gravels that, in places, have a depth to bedrock of 30 m. This operation terminated about three years ago, although Luz del Cobre, S.A. has plans for other types of placer mining.

The nearby Promontorio property contains a disseminated gold-silver ore body which is scheduled to be exploited by open-pit methods. At the time of writing (January, 1981) the new 150 t.p.d. mill was largely complete and processing of the ore should begin later in the year.

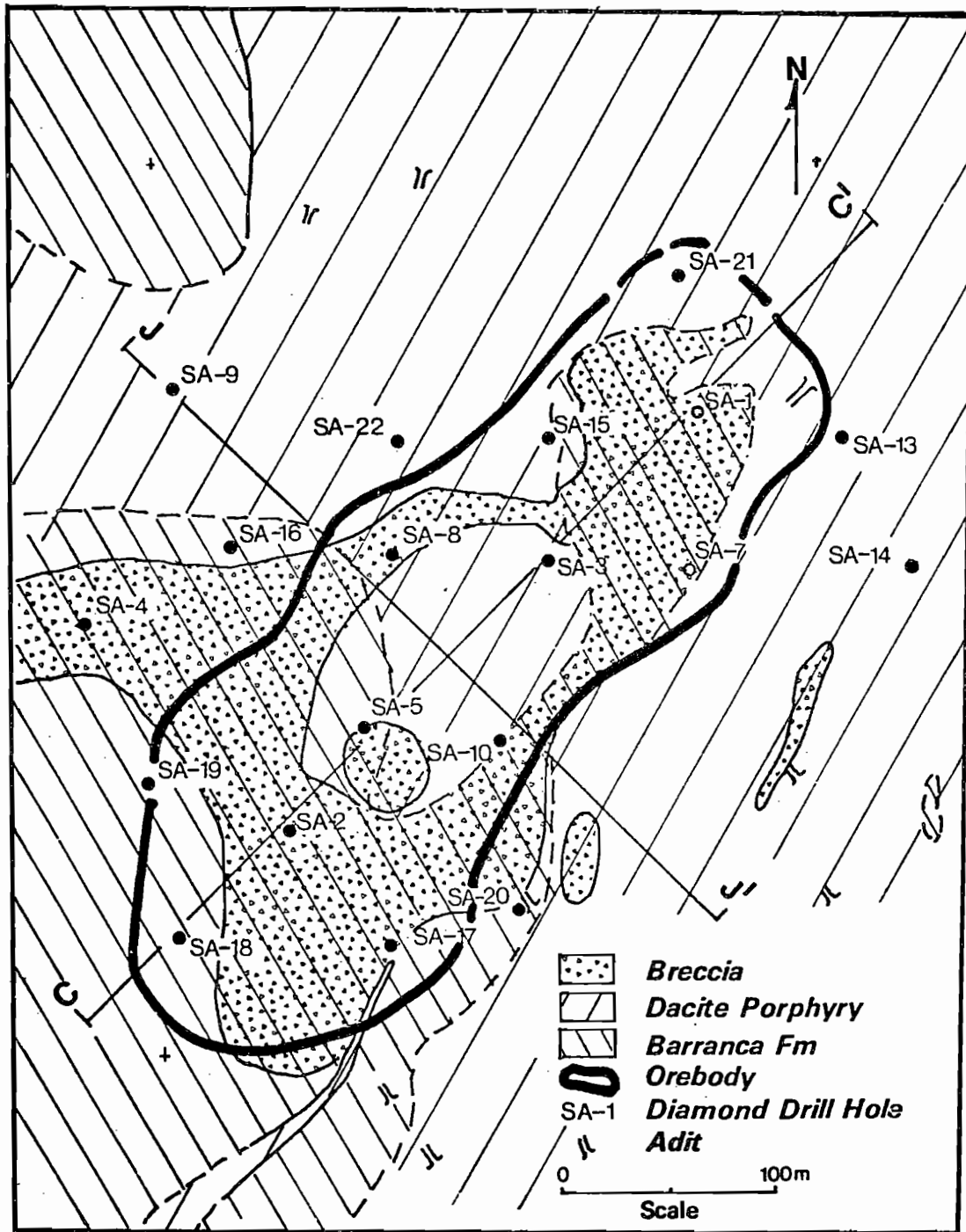
Lithology:

In the immediate area of San Antonio del Cobre Mine, a mineralized breccia pipe was formed during Laramide time. It is probably part of a deeply buried copper porphyry system. The country rocks are mostly argillites and meta-arkoses of the Barranca Formation which are intruded by Laramide age andesite and dacite porphyry dikes and sills. Brecciation of the metasediments and the early intrusives is common and late andesite dikes and sills intrude the periphery of the breccia pipe and follow zones of weakness in the contact rocks (see Figs. 7, 8, and 9).

Mineralization:

There are three items of interest in this district: Mina San Antonio del Cobre (Cu-U), Mina Promontorio (Ag-Au) and also alluvial deposits (Au).

Unfortunately, at this particular time none of these properties is in production. The operating company is Luz del Cobre. Ing. Miguel Angel Ibarra is our guide.



**Geologic map of the
San Antonio Orebody**

Figure 7. Geologic map of the San Antonio ore body.
(Courtesy of Luz del Cobre, S.A.)

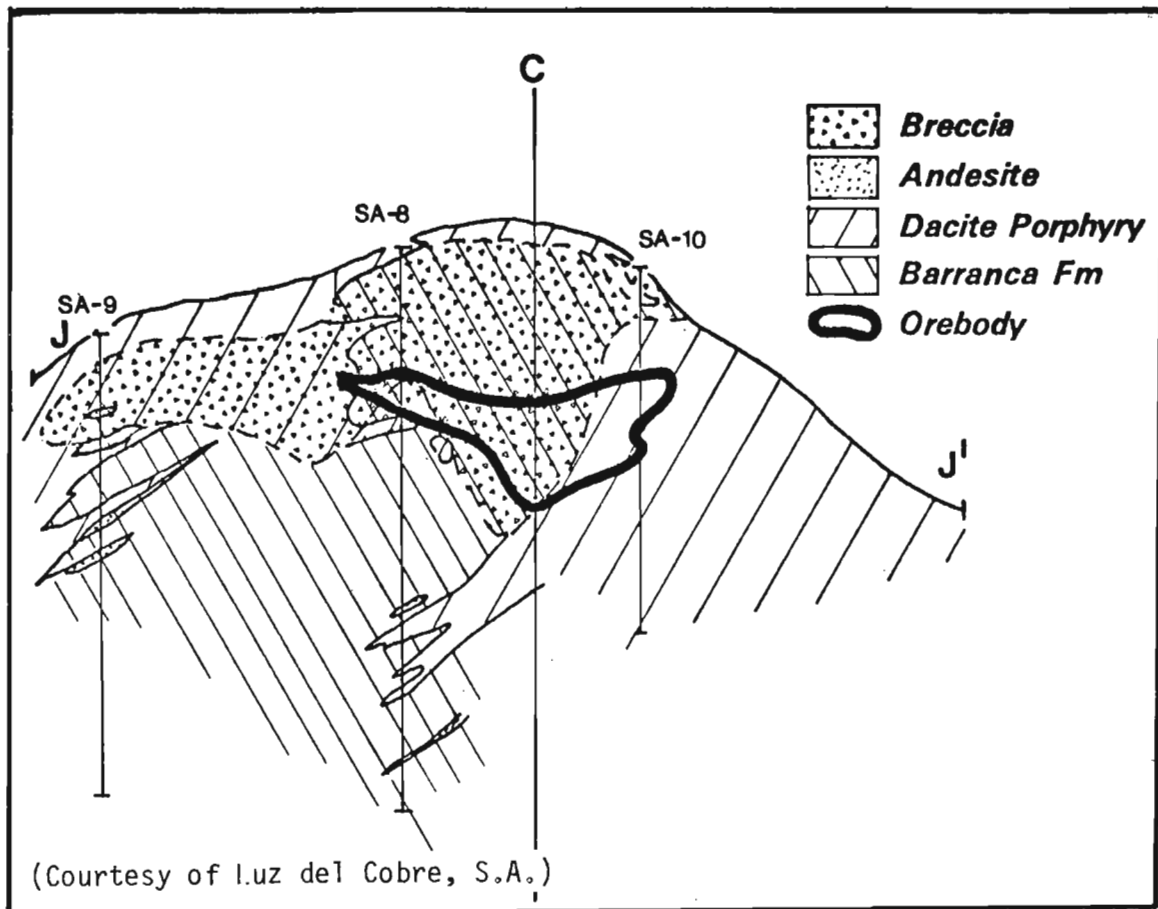


Figure 8. NW-SE cross-section of the ore body.



Figure 9. View of Luz del Cobre Mine area.

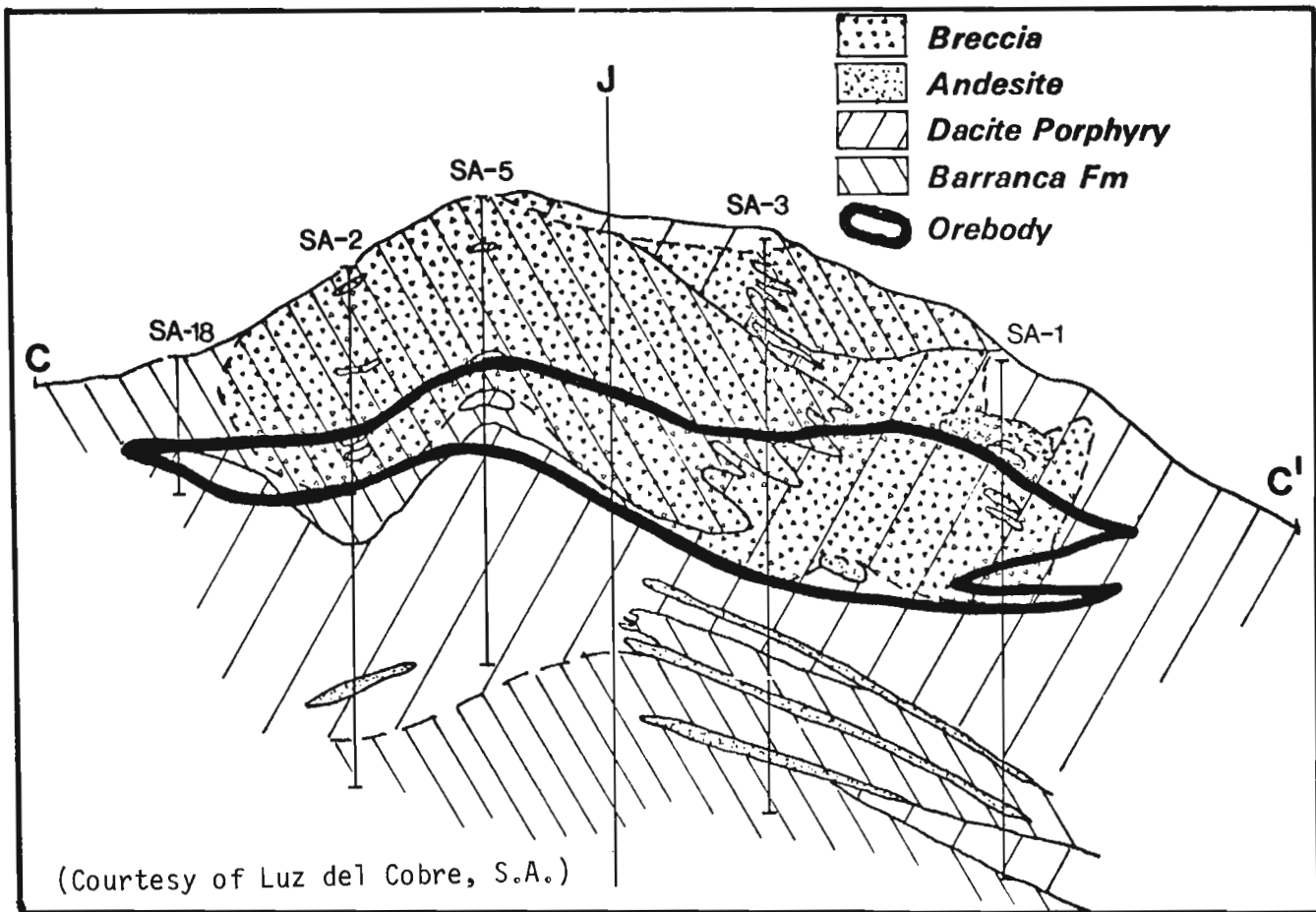


Figure 10. SW-NE Geologic cross-section of ore body.



Figure 11. Dredge used in recent placer mining, San Antonio de la Huerta area.

The road from the mine office and newly constructed mill passes through several exposures of the Barranca Fm. which has been cut by and hydrothermally altered along the road above the mine office and mill.

There are numerous adits in the oxidized mountain side where the breccia and ore bodies are located. Hydrothermal alteration is pervasive, as seen on the outcrops and within the mine workings and consists of a peripheral chlorite-epidote-calcite-pyrite-specularite assemblage (Sillitoe, 1976). In addition, there is considerable development of clay and silicification within the pipe and rudimentary stockwork. The mineralization consists of collapse breccia cavity filling by the copper minerals shown in Table 1 plus small amounts of molybdenite. Uranium occurs in the form of meta-torbernite.

Oxidation is evident by the presence of goethite, malachite, chrysocolla, hematite, limonite pitch and neotocite.

Most of the hypogene mineralization is associated with the breccia pipe, however, some disseminated mineralization is found in the intrusives. Hypogene minerals in the ore zone, in order of abundance are: specularite, pyrite, quartz, siderite, magnetite, chalcopryite, sericite, barite and orthoclase.

Supergene processes developed a secondary chalcocite zone and several iron minerals. The copper mineralogy, Table 1, indicates the degree of supergene activity within the breccia pipe. About 15% of the copper in the San Antonio deposit is chalcopryite. Most of the chalcopryite occurs in the brecciated metasediment matrix or disseminated through the intrusives.

Within the supergene zone, the better ore grades occur in the brecciated metasediments and in the upper portions of the brecciated intrusives. Much of the supergene copper occurs as chalcocite and covellite. During erosion, which removed much of the mineralized breccia pipe, oxidation of copper sulphides formed the copper silicate, carbonate and oxide minerals. These are found near the top of the supergene zone except for small scattered pods of chrysocolla found near the surface.

The breccia pipe is a very near surface orebody. Drilling has proved that the pipe roots are located immediately below the 547 level and that the contact is relatively sharp.

TABLE 1. Copper Mineralogy of the San Antonio Orebody (Based on Diamond Drilling Data, courtesy Luz del Cobre, S.A.)

<u>Mineralogy</u>	<u>% of Net Copper Content</u>
<u>Copper Silicate and Carbonate Minerals</u> (Chrysocolla, Malachite, Azurite)	10
<u>Copper Oxide Minerals</u> (Cuprite, Tenorite)	10
<u>Copper Sulfide Minerals</u> (Chalcocite, Covellite)	65
<u>Copper Iron Sulfide Minerals</u> (Chalcopyrite)	15
TOTAL	100

Return to mine work shop. Drive 0.2 mile and take left fork to bunk house for overnight stop and relaxation.

END FIRST DAY

Sunday, March 29, 1981

Summary:

We will visit a complex road cut with Barranca Fm., breccia, pyroclastic flow and Baucarit at the first stop. We will have ample opportunity to study the type Oligocene-early Miocene Baucarit Fm sediments followed by Pre-Oligocene arkosic volcani-clastics that may represent denudation of the Laramide Sonoran batholith. We will then stop at high mesas enroute to Yecora to see mid-Tertiary ash flows and basaltic andesites. We might have a photographic stop if time permits. Overnight stop will be at San Antonio de la Huerta.

Driving distance: 147.5 miles - 237.4 kilometers

Logged distance: 71.2 miles - 114.6 kilometers

Starting time: 8:00 A.M.

Assembly point: Mine work shop, San Antonio de la Huerta

Number of stops: 6

Overnight accommodation: Stretch your sleeping bag in mine bunk house or under open sky, if weather permits and relax until morning.

From Hermosillo

Cumulative Mi.(Km.)	Interval Mi.(Km.) 0.2 (.3)

119.4 (192.2) From bunk house return to previous intersection and take
left fork toward Ténichi.

0.1(.2)

119.5 (192.3) Cross Arroyo.

0.1(.1)

119.6 (192.5) Young colluvium deposit in contact with altered diorite
breccia.

0.2(.3)

119.8 (192.8) Arroyo.

0.1(.2)

119.9 (192.9) Altered diorite breccia laced with dikelets. Continued out-
crops of breccia.

0.8(1.2)

120.7 (194.2) Cross arroyo.

0.7(1.1)

121.4 (195.4) Hydrothermally altered breccia on both sides of arroyo
crossing.

(0.2(0.3)

121.6 (195.7) Greywacke composed almost entirely of pyroxene and plagio-
clase; dip 25°, strike N 70°E. It is probably part of
Barranca Fm. However, it could also be post-porphyry sediment.

403.

0.6(1.0)

122.2 (196.7) Altered breccia.

0.1(.1)

122.3 (196.8) Arroyo.

0.2(.1)

122.5 (197.1) Fairly fresh, brecciated microdiorite just before crossing arroyo. Sample collected for dating (UAKA-81-02).

0.4(0.7)

122.9 (197.8) Continue passing through subvolcanic intrusive complex to arroyo, crossing at next mileage.

1.2(1.9)

124.1 (199.7) Arroyo.

0.4(0.7)

124.5 (200.4) Turn left toward Tónichi on Sonora highway 16. Sierra Los Bledos is to the southeast; Sierra El Encinal to the east-northeast; Sierra de Las Huertas to the northwest; and Cerro Obispo to the south.

0.5(0.8)

125.0 (201.2) Bedded Baucarit Fm. conglomerates dip east towards Rio Yaqui. Highway marker 165.

STOP 7

0.3(0.5)

125.3 (201.7) Complex road cut with Barranca Fm., breccia, pyroclastic flow, Baucarit Fm. and colluvium. Folded Barranca Fm. is at the base of the cuts on both sides of the road and is unconformably overlain by andesite breccia. On the east end of the cut facing north a pyroclastic rhyodacite flow is faulted down against the breccia. The pyroclastic flow is in turn



Figure 12. Baucarit Fm in the road cut and in bluffs on both sides of Rio Yaqui Valley.

unconformably overlain by Late Tertiary - Quaternary gravels and colluvium. The west end of the roadcut shows Baucarit Fm. faulted down against the breccia.

STOP 8

0.5(0.8)

- 125.8 (202.5) Good exposure of Baucarit Fm. conglomerate from here to the bridge on the Rio Yaqui. It is a good vista point to photograph the Baucarit Fm. Competent beds display a unique landscape. Clasts are composed of diverse lithologies of the various preceding geologic ages.

0.3(0.4)

- 126.1 (202.9) Bridge on Rio Yaqui. This river flows all year round. It drains the controlled overflow from El Novillo Reservoir, built to control flood and supply power and irrigation water

to the Ciudad Obregon area. El Yaqui bridge was built in 1978 with equal proportions of state and federal money during the governorship of Lic. Alejandro Carrillo Marcor.

1.2(2.0)

- 127.3 (204.9) Dirt road to the left (north) leads to Tónichi - a small pueblo with a 10 ton per day concentrator mill. Sphalerite, scheelite, chalcopyrite, bornite and molybdenite are concentrated to copper, tungsten and molybdenum. The ore appears to be a skarn and the host rock is rather fine-grained. In the arroyo near the mill, Baucarit Fm. conglomerate overlies a granodioritic basement intruded by aplites. Keep ahead on paved road.

0.4(0.6)

- 127.7 (205.5) Baucarit Fm. continues in road cuts.

0.3(0.4)

- 128.4 (206.6) Weathered granodiorite cut by aplite dikes.

0.7(1.3)

- 129.1 (207.9) Granodiorite intruded by pegmatite. The pegmatite is in turn cut by fine-grained diorite that, in turn, is intruded by an even finer-grained rock, possibly an andesite dike. The hornblende crystals in the granodiorite are fairly large and fresh. The pegmatite dikes are up to 2 m wide.

0.3(0.4)

- 129.4 (208.2) Turn off to Onabas and Rio Chico to the south.

3.2(5.2)

- 132.6 (213.4) Dark-colored mafic dikes crisscross the granodiorite. We have been passing through batholithic granodiorite for the last 3.5 miles.

0.8(1.0)

- 133.4 (214.7) The number of diorite dikes has increased and the granodiorite becomes injected, in parts, by numerous dikes.

2.3(3.7)

- 135.7 (218.4) The batholith has become dioritic in composition at this mileage.

0.7(1.1)

- 136.4 (219.5) Brick-red diorite porphyry results from weathering to limonite.

1.4(2.3)

- 137.8 (221.8) Granodioritic phases, in places, are intruded by dikes, and
to
139.8 (225.0) are brecciated, hydrothermally altered and stained with
limonite. Some of the units are probably subvolcanic
andesite porphyry.

0.1(0.2)

- 139.9 (225.2) Arroyo Palmeras highway construction camp. The arroyo is well named because it has many palm trees growing within it.

0.2(0.3)

- 140.1 (225.5) Ashflow unit with eutaxitic texture crops out above the road cuts. The ash flows are tilted from 20 to 35° and the unit is at least 300 m thick.

0.8(1.3)

- 140.9 (226.8) Bridge on arroyo. Highway marker 201 km. Sample collected for dating at 201 km plus 400 meters (UAKA-80-8).

1.0(1.6)

141.9 (228.4) Base of volcanic section with volcanogenic sediments containing angular clasts. Sediments are brick red due to iron stains.

0.5(0.8)

142.4 (229.2) Agglomerate with monolithologic volcanic clasts.

to

142.8 (229.8)

0.7(1.1)

143.5 (230.9) Pyroclastic flows and some volcanogenic sediments.

0.9(1.5)

144.4 (232.4) Subvolcanic porphyry system with andesitic dikes and disseminated sulfides including chalcopyrite. It is fractured and

to

145.9 (234.8)

hydrothermally altered. Limonitic soil caps the hills.

0.1(0.2)

146.0 (235.0) Porphyry is brecciated and is not limonitic. Sample UAKA-80-9 is a fine-grained purple andesite and yields 31 m.y. age. This andesite represents the first phase of a porphyry system.

0.1(0.1)

146.1 (235.1) Bridge over arroyo. Shattered andesites.

0.4(0.7)

146.5 (235.8) One run-down hut is the evidence of Mina La Olga. The area is conspicuously hydrothermally altered with bright yellow, red and orange oxidation colors. Limonitic soil is distinctive in the hills.

2.1(3.3)

148.6 (239.1) For the next 5.5 miles we are in a Baucarit Sequence with interbedded andesites and basalts and conspicuous baked



Figure 13. Baucarit Formation sediments overlain by dark volcanic rocks.

basal rubble zones and weathering horizons. The sequence gently dip in a westerly direction. It is capped by a 14 m.y. basalt (UAKA-80-10). The 31 m.y. date on older andesite (UAKA-80-09) brackets the Baucarit between the Mid-Oligocene and the Mid-Miocene in this area. In other areas in Sonora, we have dated the Baucarit at lower Miocene to Mid-Miocene in age.

4.9(7.9)

STOP 9

153.5 (247.0) We have been steeply descending through the Baucarit Fm. to the valley of Arroyo La Trinidad. At this stop we will see a section of Baucarit including a basalt flow.

0.6(1.0)

154.1 (248.0) Bridge over Arroyo La Trinidad along which is located the town of Tepoca with its orange groves and many well-kept

houses with flower gardens. The Baucarit Fm. is also exposed on the east side of the arroyo.

NOTE: The distance between Tepoca and San Nicolas is 8.1 miles (13 km) according to the map, and 9.9 miles (16 km.) according to our odometer. Part of the log is for the new highway, part on a detour road and part on the old road. Hence, the log will not be precise for the new highway.

0.3(0.5)

154.4 (248.5) Non-stratified fanglomerate.

1.9(3.0)

156.3 (251.5) Fractured, faulted and hydrothermally altered sequence of volcanics underlying the Baucarit Fm. The conglomeratic horizons have fine-grained tuffaceous layers. The beds are gently folded.

0.6(1.0)

156.9 (252.5) Red clay sediments, steeply dipping to the east and overlain by altered basalt or basaltic andesite. Both units have been deformed and broken by a fault.

0.9(1.5)

157.8 (254.0) Granodiorite outcrops in the roadcuts, 49.6 m.y. age biotite (UAKA-80-11). Highway marker 231.54 Km. The granodiorite is exposed over a wide region from here to Santa Rosa.

3.7(5.9)

161.5 (259.9) San Nicolas. This pueblo will presumably be an important area in the future because the Hermosillo - Chihuahua and Ciudad Obregon - Agua Prieta highways under construction cross here. According to local people, two small mines

are now abandoned in this area. La Sesentona was worked by the Spaniards who produced 60 kilos of silver while Mina La Cruz produced scheelite for tungsten metal.

The 29 mile (47 km.) stretch between San Nicolas and Yecora was logged on the old winding dirt road. The new highway at San Nicolas runs about one km. south of the old road. The new highway follows the old road in broad terms but different in details. For example, large stretches of the old road were in the arroyos and along winding topography. This part of the log is hence, simplified and generalized.

1.2(1.9)

162.7 (261.8) Granodiorite along Arroyo La Trinidad.

3.7(6.0)

166.4 (267.8) Biotite hornblende granodiorite (UAKA-80-12) with concordant 49.5 m.y. age was collected along the arroyo.

3.1(5.0)

169.5 (272.8) Santa Rosa - a small pueblo. Granodiorite is exposed nearby. Molybdenum ore is mined from Tres Piedras mine nearby.

1.2(1.9)

170.7 (274.7) Porphyritic andesite outcrops along a small arroyo. This porphyry is probably associated with the granodiorite as a subvolcanic phase. About 200 meters up the mountain side, the rock becomes finer-grained and grades to bedded purple andesite, about 20 m thick. 20 meters higher up a rhyodacite

flow with eutaxitic texture crops out. Hence, the ash flow caps andesite and granodiorite. Further up the hill slope we pass along alternating outcrops of andesite and ash flow at various levels. They are apparently not interbedded. At the ridge top, thin to medium-bedded andesite tuff beds strike 332° with a dip of 32° towards the southwest.

1.9(3.1)

- 172.6 (277.8) Left fork leads to La Trinidad - a fissure-vein silver mining camp. Take the right fork.

STOP 10

1.4(2.2)

- 174.0 (280.0) An ash flow unit, presumably a basal unit of the upper volcanic series, is in fault contact with a volcanic-clastic section. The high-angle fault strikes 130° and the beds dip 54° to the south. Drag folds indicate that the volcanic-clastic section is down-faulted. It consists of 3 - 4 meters of arkosic sediments and 4 - 5 meters of brick-red clastic rocks followed by three basaltic andesite flows, and capped by an ash flow unit. The middle andesite (UAKA-80-14) yields a 27 m.y. age. The flows are vesicular and filled with zeolite.

1.7(2.8)

- 175.7 (282.8) Cream-colored ash flow. Good view of La Trinidad to the west. For the next 7.4 miles the road traverses through pine forest.

STOP 11

7.4(11.9)

183.1 (294.7) Mesa capping basaltic andesite (UAKA-80-15) with a 23.2 m.y. age. Elevation 1,788 meters. Good view of the parallel ranges flanking the Sierra Madre Occidental. The road descends towards the Yecora valley.

4.6(7.4)

187.7 (302.1) Good view of Yecora and the high plains of the valley. The pyroclastic unit contains rock fragments and is highly jointed, giving an impression of a vent area.

STOP 12

2.7(4.3)

190.4 (306.4) Yecora, the gateway to western Chihuahua, is on the bank of Arroyo El Toro. Elevation 1,542 m. The valley is a cattle ranching area. Landscape is typical of the Sierra Madre Occidental mesetas.

76.3(122.8)

266.7 (429.2) San Antonio de la Huerta mine. Back track to the mine bunk house for dinner, informal discussion and overnight relaxation.

END OF THE SECOND DAY

THIRD DAY

Monday, March 30, 1981

Summary:

We will have ample time to visit the type Barranca Formation, made famous by King (1939) after his one-man expedition in the thirties. Then we will view Eocene sandstone and conglomerate resulting from denudation of Sonoran batholith. This great denudation took place on a continental scale all along the western cordillera of the United States of America and Mexico. Then we will view a typical granodiorite facies of the Sonoran batholith. The last stop will be "PARA LIBAR, COMER y CHARLAR ANTES VAMOS A REGRESSOR A NUESTROS HOGARES, DULCE HOGARES; QUE LES VAYAN BIEN! ADIOS!!!" However, participants wishing to leave early for travel connections may do so.

Driving distance: 107.1 miles - 172.4 kilometers

Logged distance: 100.9 miles - 162.4 kilometers

Starting time: 8:00 A.M.

Assembly point: Mine work shop, San Antonio de La Huerta

Number of stops: 6

THIRD DAY

Interval
Mi.(Km.)

Cumulative
Mi.(Km.)

The starting point for the remainder of the log is at the intersection of the paved highway and the road to San Antonio de la Huerta. Begin log at 266.7 miles (429.2 Km.).

0.6(1.0)

267.3 (430.2) Basaltic outcrop with red-baked base.

0.5(0.8)

267.8 (431.0) Light-colored rhyodacite on northside of road.

0.4(0.6)

268.2 (431.6) Very fine-grained granodiorite.

0.4(0.7)

268.6 (432.3) Fine-grained shattered granodiorite porphyry.

0.5(0.8)

269.1 (433.1) Shattered subvolcanic porphyry.

0.3(0.5)

269.4 (433.6) Protrusive units of variegated colors are cut by a leucocratic dike. Km. 160 highway marker around here.

0.3(0.4)

269.7 (434.0) Rancho Flores turn-off to the north.

2.2(3.6)

271.9 (437.6) Fine-grained diorite with spheroidal weathering. Winding road.

0.6(0.9)

272.5 (438.5) Light-colored fine diorite with fine-grained dikes. Winding road. Km. 155 highway marker around here.



Figure 14. Barranca formation at mileage 272.9.

STOP 13

0.4(0.7)

272.9 (439.2) Barranca Fm.; cross-bedded arkosic sandstone, siltstone and shale (black, grey and reddish brown colors) are slightly metamorphosed. The beds dip 40° E and strike 345°. This is a good area for collecting plant fossils. Park in pull-off area on right side. See King's (1939) paper for details.

0.5 (0.8)

273.4 (440.0) Exposure of fine-grained granodiorite intrusion. Barranca Fm. may be large roof pendant surrounded by granodiorite batholith.

0.2(0.3)

273.6 (440.3) A small sliver of Barranca Fm. within the fine-grained granodiorite.

0.1 (0.2)

273.7 (440.5) Barranca Fm. on both sides of the arroyo.

0.9(1.4)

274.6 (441.9) Shale units of Barranca Fm. are folded, fractured, deformed, and hydrothermally altered. Quartzite layers are competent. Some beds are vertical.

0.4 (0.7)

275.0 (442.6) Rancho La Barranca. The Formation is named after this Ranch. Km 151 marker is around here.

0.5(0.8)

275.5 (443.4) Border phase of fractured and deformed Barranca Fm. intruded by granodiorite including a vertical dioritic dike which is

three meters wide.

0.3(0.5)

275.8 (443.9) Conglomeratic phase of Barranca Fm., hydrothermally altered and deformed.

0.1(0.1)

275.9 (444.0) Well-bedded, folded Barranca Fm. with cross bedding.

1.0(1.6)

276.9 (445.6) Subvolcanic, epidotized, dominantly fine-grained granodioritic breccia, but not monolithic (grey-purple in color). Buff-colored cliffs to the east may be ash flows.

1.3(2.1)

278.2 (447.7) Breccia with predominantly Barranca Fm. clasts. Shales seem to form slip planes. Some large segments of shale are almost intact. Some clasts are highly metamorphosed.

0.6(1.0)

278.8 (448.7) Cross Arroyo.

STOP 14

1.9(3.0)

280.7 (451.7) Park vehicle at turn-off by arroyo. We will walk back through the section for a distance of one mile. At the bottom of the section, the sediments are arkosic sandstones and conglomerates. Going up the section, we begin to get stringers and boulders of breccia. The clasts must have been washed in from adjacent areas. The stringers increase going up section until we are in a section of landslide breccia at least 100 meters thick.

0.9(1.5)

281.6 (453.2) Dirt road to the north goes to San Xavier.

0.4(0.6)

282.0 (453.8) Km. 140 marker.

0.2(0.4)

282.2 (454.2) Small outcrop of Barranca Fm. landslide breccia.

0.5(0.8)

282.7 (455.8) Barranca Fm. Most of the section here is composed of greywacke with some breccia and conglomerate with shaley partings.

0.2(0.3)

282.9 (455.3) Greywacke facies of Barranca Fm.

0.5(0.8)

283.4 (456.1) Limonitic weathering and hydrothermally-altered Barranca Fm.

0.2(0.3)

283.6 (456.4) Metasomatized and pyritized Barranca Fm. border facies with batholith.

0.5(0.8)

284.1 (457.2) Km. 135 highway marker. Metasomatized and pyritized Barranca Fm.

1.4(2.3)

285.5 (459.5) Barranca Fm. greywacke. Highly oxidized but not pyritized

1.1(1.7)

286.6 (261.2) Bedded Barranca Fm.

0.7(1.12)

287.3 (462.4) Granodiorite - Barranca Fm. contact.

0.9(1.4)

288.2 (463.8) Granodiorite in the wash.

1.0(1.6)

289.2 (465.4) Apophyses of granodiorite in Barranca (?) Fm.

1.4(2.3)

290.6 (467.7) Km. 125 marker.

1.9(3.0)

292.5 (470.7) Granodiorite in small hills to the north. UAKA-80-07 yields
62.2 m.y. hornblende and 60.7 m.y. biotite age.

2.5(4.1)

295.0 (474.8) Bridge.

0.6(0.9)

295.6 (475.7) Entering Tecoripa. Two large hills to the south are part
of the Sierra Verde.

0.5(0.8)

296.1 (476.5) Baucarit conglomerate exposed in the river.

0.8(1.3)

296.9 (477.8) Baucarit conglomerate.

0.7(1.1)

297.6 (478.9) Rhyodacite flow.

0.8(1.3)

298.4 (480.2) Flat stretch ahead without roadcuts exposing bed rocks.
we are passing from the parallel ranges to the buried ranges
of the Sonoran Desert sub-provinces.

8.0(12.9)

306.4 (493.1) Km. 100 marker. Continue west on Sonora Highway 16. Upper
Sonoran grassland region where vegetation changes from

mountain type to grassland. It consists of mesquite, creosote, organ pipe-like cactus, palo verde, prickly pear (nopal) and abundant grass.

0.7(1.1)

- 307.1 (494.2) Baucarit Fm. containing cobbles and boulders of primarily volcanic and subvolcanic rocks. This seems to overlie a breccia composed of large and small fragments of subvolcanic rocks including an olivine-bearing basalt or basaltic andesite.

0.5(0.8)

- 307.6 (495.0) Bridge.

0.4(0.6)

- 308.0 (495.2) Pyroclastic flow of probable Oligocene age.

0.4(0.7)

- 308.4 (495.7) Highly weathered granodiorite intruded by small felsic dikes.

0.7(1.7)

- 309.1 (497.4) Bridge. Probable Pliocene colluvium just east of the bridge. Notice thick grassland soils.

1.9(3.1)

- 311.0 (500.5) Bridge. Highly weathered Baucarit or younger gravels outcrop for the next 0.8 miles.

0.8(1.3)

- 311.8 (501.8) Thick grassland soils for the next 0.2 miles.

1.2(1.9)

- 313.0 (503.7) Bridge. Thick colluvial soil for the next 1.4 miles.

0.9(1.5)

- 313.9 (505.2) Sanidine-bearing pyroclastic flow with oxidized biotite. Sampled for K-Ar dating. Km. 85 marker here.

0.5(0.8)

- 314.4 (506.0) Well-bedded arkosic sandstone, siltstone and conglomerate containing cobbles and pebbles primarily of quartzite, granodiorite, jasper and chert. This unit is conspicuous by the absence of volcanic cobbles. It is overlain by a boulder-cobble regolith of the ash flow. Hence, this sedimentary unit predates the ash flow. It dips 20° towards S 40° E.

0.2(0.3)

- 314.6 (506.3) Pre-ash flow sediment, about 70 meters thick, is overlain by boulders of the ash flow.

STOP 15.

0.1(0.2)

- 314.7 (506.5) Rancho San Francisco: we will walk back to the outcrop. Pre-ash flow sediments exposed in small outcrops for the next half-mile.

2.0(3.2)

- 316.7 (509.7) Sediment with cobbles are cemented with carbonate and overlain by ash flow near Km. 80 marker.

3.9 (6.2)

- 320.6 (516.0) Lower pre-ash flow carbonate cemented unit. Cobbles are mostly quartzite with minor hornblende dacite and subvolcanic porphyry.

0.1(0.1)

- 320.7 (516.1) Bridge. Broad valley.

0.3(0.5)

- 321.0 (516.6) Carbonate cemented unit.

0.5(0.8)

- 321.5 (517.4) Entering San José de Pimas.

- 0.7(1.1)
- 322.2 (518.5) Paved stream crossing.
- 2.5(4.1)
- 324.7 (522.6) Road to Cobachi on right.
- 0.9(1.4)
- 325.6 (524.0) Lower carbonate cemented unit with andesites to rhyolite
flow-banded volcanic clasts.
- 1.1(1.8)
- 326.7 (525.8) Bridge near Km. 65 marker.
- 0.5(0.8)
- 327.2 (526.6) Granite outcrop west of carbonate cemented sediment.
- 0.3(0.4)
- 327.5 (527.0) Granodiorite outcrop with diorite dikes.
- 0.2(0.4)
- 327.7 (527.4) Diorite dike intruded into granite. Granite weathers away
leaving diorite outcrop.
- 0.3 (3.7)
- 330.0 (531.1) Granite.
- 0.3(0.5)
- 330.3 (531.6) Granite, deeply weathered.
- 0.7(1.1)
- 331.0 (532.7) Bridge.
- 0.2(0.3)
- 331.2 (533.0) Road to San Feliciano Ranch.
- 0.8(1.3)
- 332.0 (534.3) Chicken farms. The three chicken farms belong to Grupo
A.I.R.R.E.

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0.3(0.5)

332.3 (534.8) Bridge.

1.1(1.8)

333.4 (536.6) Second Chicken Farm.

0.9(1.4)

334.3 (538.0) Third Chicken Farm.

STOP 16.

1.2(1.9)

335.5 (539.9) Carbonate sediments with granite regolith; sediments dip
15° E.

0.6(1.0)

336.1 (540.9) Granite.

0.1(0.2)

336.2 (541.1) Granite.

0.4(0.6)

336.6 (541.7) Granite.

0.3(0.5)

336.9 (542.2) Granite.

0.8(1.3)

337.7 (543.5) Granite with diorite dikes for next 0.3 miles.

0.5(0.8)

338.2 (544.3) Bridge.

0.4(0.6)

338.6 (544.9) Granodiorite.

0.3(0.5)

338.9 (545.4) Granodiorite just before entrance to La Colorada.

423.

0.3(0.5)

339.2 (545.9) Abandoned mine building at 9 o'clock.

0.7(1.1)

339.9 (547.0) Enter La Colorada.

0.5(0.8)

340.4 (547.8) Bridge.

0.6(0.9)

341.0 (548.7) Bridge. Ignimbrite in hill at 12 o'clock sampled (UAKA-80-06) for dating.

0.5(0.8)

341.5 (549.6) "Shrine" in pyroclastic flow at 9 o'clock.

0.5(0.9)

342.0 (550.4) Ash flow boulders along road.

0.8(1.3)

342.8 (551.7) Ejido on left.

0.1(0.1)

342.9 (551.8) Bridge.

3.4(5.5)

346.3 (557.3) Beehives.

0.6(1.0)

346.9 (558.3) Two bridges in 0.2 miles.

2.1(3.4)

349.0 (561.7) Begin ascent in granodiorite boulders on both sides of road.

0.9(1.4)

349.9 (563.1) Granodiorite. Continue in granodiorite.

STOP 17.

0.5(0.8)

350.4 (563.9) "Granita Hermita". Hornblende from here yields 62.9 m.y.
(UAKA-80-05) age.

1.6(2.6)

352.0 (566.5) Road curves right.

0.6(1.0)

352.6 (567.5) Carbonate cemented gravels contain andesite, diorite and granodiorite cobbles.

J.3(0.4)

352.9 (567.9) Road ascends. Begin long straight stretch without outcrops.

5.6(9.0)

358.5 (576.9) Calera Willard "Cal del Pacifico Productos y Derivados S.A."

0.7(1.2)

359.2 (578.1) Cross railroad track.

0.2(0.3)

359.4 (578.4) Stockyard to right.

0.7(1.1)

360.1 (579.5) Granja Porcicola "La Mision" entrance.

2.2(3.5)

362.3 (583.1) Banded carbonate deposit with angular carbonate clast. Rubbly outcrop.

1.2(1.9)

363.5 (585.0) Bridge.

2.1(3.4)

365.6 (588.4) Airstrip and stockyard on right.

1.8(2.9)

367.4 (591.3) Intersection. Turn left, then turn right on to periferico.

5.2(8.3)

372.6 (599.6) Take left around glorieta onto Blvd. Kino. Statue is of Kino.

373.8 (601.6) Valle Grande Hotel.

- END EXCURSION -

ACKNOWLEDGEMENT

We wish to express our profuse thanks to Barite of Sonora and Luz del Cobre for making available their unpublished records and to many individuals who helped in numerous ways in our efforts with the road log and to Patricia Clark who typed the manuscript. Particular thanks to the taxpayers who supported us through the National Science Foundation Grants EAR-78-11535 and INT-78-21876.

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MESOZOIC THROUGH EARLY TERTIARY SEDIMENTATIONAL
AND TECTONIC PATTERNS
OF NORTHEAST SONORA AND SOUTHEAST ARIZONA

(Roadlog not available at press time)

FIELD TRIP No. 12 PREPARED FOR THE
GEOLOGICAL SOCIETY OF AMERICA
CORDILLERAN SECTION
1981 ANNUAL MEETING

MARCH 28-30

TRIP LEADERS:

S. KEITH, BUREAU OF GEOLOGY AND MINERAL TECHNOLOGY, GEOLOGICAL
SURVEY BRANCH, 845 N. PARK AVE., TUCSON, ARIZONA 85719
U.S.A.

W. L. BILODEAU, DEPARTMENT OF GEOLOGY, SCHOOL OF EARTH SCIENCE,
STANFORD UNIVERSITY, STANFORD CA 94305, U.S.A.

J. L. LONGORIA (FORMERLY AT:) DEPARTAMENTO DE GEOLOGÍA,
FACULTAD DE INGENIERIA, UNIVERSIDAD DE SONORA, HERMOSI-
LLO, SONORA, MEXICO.

Late manuscript corresponding to Field trip 2,
Geology of coastal Sonora between Puerto Lobos
and Bahía Kino: see p. 57-149.

Reconnaissance geology of Cerro Pozo Serna, western Sonora, Mexico.

(R.J. DOWLEN: R.G. GASTIL)

RECONNAISSANCE GEOLOGY OF CERRO POZO SERNA, WESTERN SONORA, MEXICO

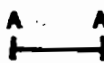
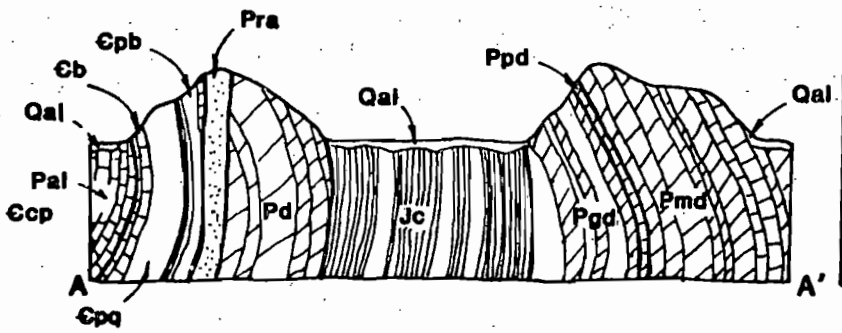
ROBERT J. DOWLEN, Woodward-Clyde Consultants, San Diego, Ca 92110
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Cerro Pozo Serna, located approximately 60 Km south of Caborca, Sonora, Mexico (Lat. $30^{\circ} 10'$, Long. $112^{\circ} 12'$), contains at least 1900m of Late Jurassic intraformational breccias, sandstones, and shales. These sediments, previously considered Paleozoic (Dumble, 1900; King, 1939), comprise a north-striking overturned fault block situated between apparently unnamed early Paleozoic/Pre-Cambrian dolomites. The entire sequence is overturned towards the west between Tertiary (?) age northwest-striking faults. Overturned Early Cambrian beds of the Puerto Blanco Formation, Provodora Quartzite, Buelna Formation, and Cerro Prieto Formation, which are correlated to the Early Cambrian beds reported by Eells (1972), are in fault contact with the dolomites on the west flank of Cerro Pozo Serna (Fig. 2).

The lower one-third of the exposed Jurassic section contains an extremely fossiliferous, taxonomically diverse, invertebrate fauna containing internal molds, casts, and impressions of ammonites, belemnites, bivalves, gastropods, scaphopods, anthozoans and crinoids. Unidentifiable crinoid and mollusk fragments occur within a 50m thick biosparite at the top of the exposed section; no fossils have been found in the remainder of the section. A Late Oxfordian-Lower Kimmeridgian age was tentatively assigned the lower one-third of the section due to the presence of the ammonoid genera Amoeboceras and Idoceras (Dowlen and Stump, 1973; Beauvias and Stump, in prep). A columnar section is presented in Fig. 2.

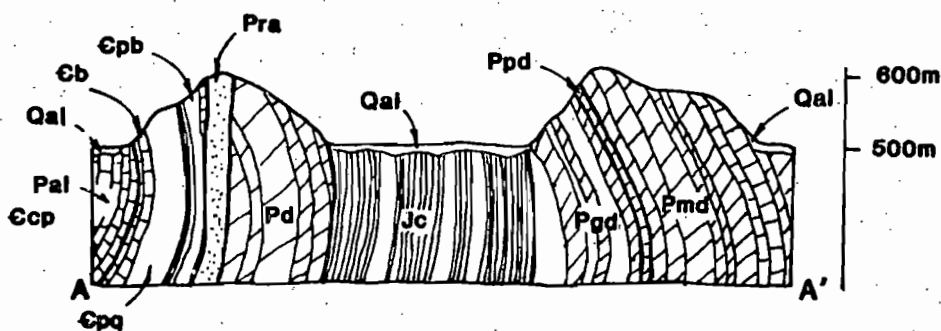
El Cerro Pozo Serna localizado aproximadamente a 60 Km al sur de Caborca, Sonora (Latitud $30^{\circ}10'$ y longitud $112^{\circ}12'$) contiene por lo menos 1900 m de brechas intraformacionales, areniscas y lutitas del Jurásico Superior. Estos sedimentos previamente considerados Paleozoicos (Dumble, 1900; King, 1939), comprenden un block fallado invertido con rumbo de estratos al norte. Situado entre dolomitas aparentemente del Paleozoico Inferior- Precámbrico no nombradas. La secuencia entera está invertida hacia el oeste entre fallas de rumbo noroeste de edad Terciaria (?). Capas invertidas del Cámbrico Temprano de la Formación Puerto Blanco, Cuarcita Proveedora, Formación Buelna y Formación Cerro Prieto, que son correlacionadas con capas del Cámbrico Temprano reportadas por Eells (1972), están en contacto con las dolomitas por una falla en el flanco oeste del Cerro Pozo Serna. (Figura 2).

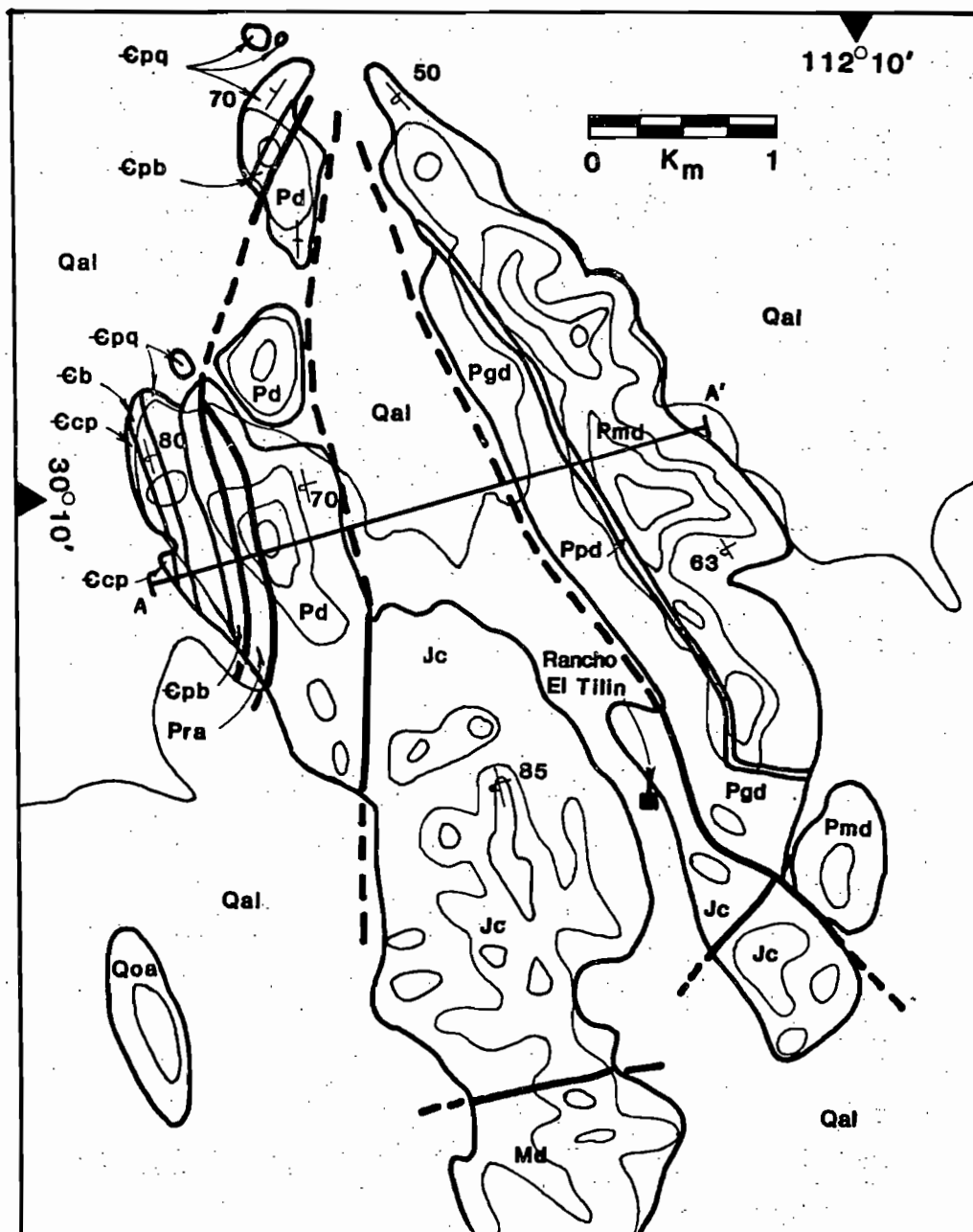
LEGEND:

Cenozoic	Qal	Alluvium		Fault
	Qoa	Older Alluvium		Geologic Contact
Mesozoic	Md	Dacite		Dip and Strike
	Jc	Sandstones and Shales		Overtured Bed
Paleozoic	Ccp	Cerro Prieto Formation		Vertical Bed
	Cb	Buelna Formation		Geologic Cross Section
	Cpq	Provodora Quartzite		
	Cpb	Puerto Blanco Formation		
Paleozoic/Pre-Cambrian	Pra	Red Arkose		
	Pd	Tan and Gray Dolomite		
	Pgd	Gray Dolomite and Sandstone		
	Ppd	Pink Dolomite		
	Pmd	Mottled Dolomite		

A A'

Geologic Cross Section

**FIGURE 1**



RECONNAISSANCE GEOLOGIC MAP
CERRO POZO SERNA
SONORA, MEXICO

FIGURE 1

COLUMNAR SECTION OF THE CERRO POZO SERNA REGION, SONORA, MEXICO

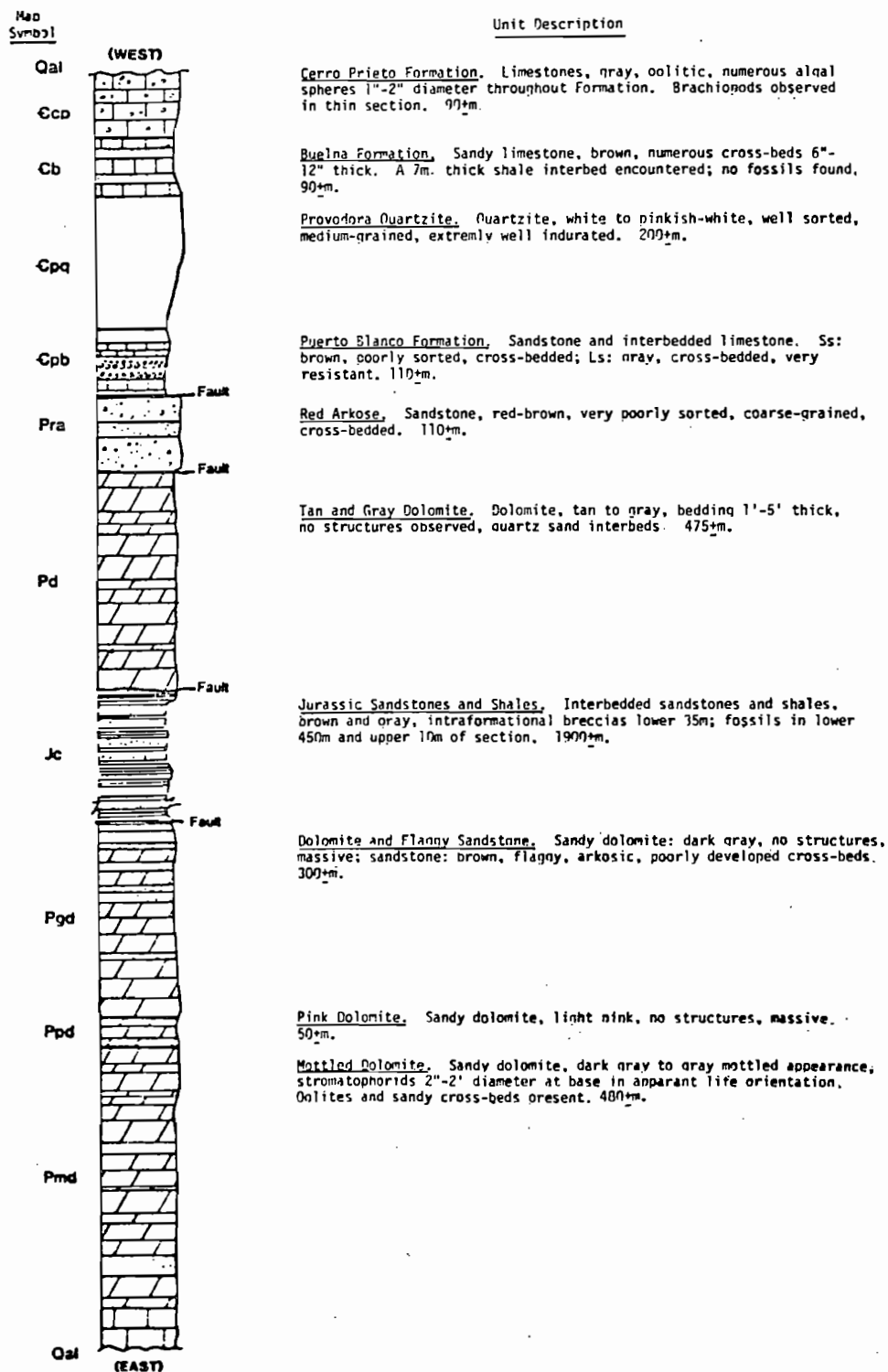


FIGURE 2

Un tercio de la sección expuesta del Jurásico Inferior es extremadamente fosilífera de diversa Taxonomía, contiene moldes internos, moldes externos de invertebrados e impresiones de amonitas, belemnites, bivalvos, gasteropodos, escafopodos, antozoarios y crinoideos. Fragmentos de crinoideos y moluscos no identificados aparecen en una capa de biosparita de 50 m de espesor, en la parte superior de la sección expuesta. No se han encontrado fósiles en el resto de la sección. Una edad Oxfordiano Superior- Kimmeridgiano Inferior fué tentativamente asignada a un tercio de la sección debido a la presencia de amonoideos del género Amoeboceras e Idoceras (Dowlen y Stump, 1973; Beauvias y Stump, en prep.). Una columna estratigráfica se presenta en la figura 2.

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