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ORIGIN OF MARBLE BY REPLACEMENT OF GYPSUM IN CARBONATE
BRECCIA NAPPEs, CARSON SINK REGION, NEVADA¹

Carson Desert
Churchill Co

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ABSTRACT

Sheets of carbonate breccia and unbrecciated calcitic carbonate rocks, called carbonate breccia nappes, occur as tectonic slices in nappe piles dominated by pelitic rocks of Early Mesozoic age in the Carson Sink region of Nevada. The most extensive carbonate breccia nappe, the Muttelbury nappe, crops out over 100 km². Such sheets are composites of a number of different carbonate rock types, including widespread calcite marble. The coarse grain size and granoblastic texture of the marble imply an origin by metamorphism of at least moderate grade. The degree of apparent recrystallization of marble, however, contrasts markedly with that of most of the limestone with which it is intercalated and is independent of degree of metamorphism, locally nil, of pelitic rocks in adjacent nappes. If the marble is a product of thermal metamorphism, its grade is seemingly incompatible with that of associated rocks. On the basis of field and petrographic studies, the origin of the marble is enigmatic. Carbon and oxygen isotopic ratios of calcites from various carbonate rocks of the Muttelbury Formation and the probably correlative Jurassic Lovelock Formation fall into two distinct trends. One trend has nearly constant δC^{13} and variable δO^{18} , and the other has nearly constant δO^{18} and δC^{18} that varies between -7 and -25 ‰ (PDB). The first includes all the fine-grained carbonate rocks and calcarenites whereas the second comprises all of the analyses of the problematic marble. The interpretation of the isotopic trend of marble is that rocks precursory to the marble were coarse-grained gypsum such as those which are abundant in the Lovelock Formation and in one small body in the Muttelbury Formation. The gypsum rocks were calcitized by biogenic carbon derived from sources outside the sulfate beds, probably from euxinic limestones stratigraphically below the sulfate deposits, or perhaps, from a widespread pelite terrane. Biogenic carbon could have been transported as hydrocarbons in an aqueous medium or as dissolved carbon species oxidized in the source beds. Intraformational breccia of the Muttelbury Formation is a solution breccia created by aqueous dissolution of sulfate rocks remaining after calcitization was complete. Calcite-quartz sand, ubiquitous in the breccia, was derived from sand originally disseminated throughout such sulfate beds.

INTRODUCTION

An extensive carbonate nappe in the West Humboldt Range, Nevada, contains coarse-grained marble as its predominant lithic constituent. The origin of the marble on the basis of field and petrographic studies is enigmatic because, on one hand, its texture is suggestive of substantial metamorphism, but on the other, many of the rocks contacting marble are little metamorphosed. Analyses of the carbon and oxygen isotopic compositions of rocks of the carbonate nappe, however, indicate disparate isotopic histories between marble and associated carbonate rocks. The isotopic studies lead to the proposal that

marble originated by replacement of coarse-grained gypsum. Replacement proceeded either by oxidation by sulfate of biogenic carbon moving through the gypsum unit or by transfer of oxidized biogenic carbon species to gypsum beds undergoing aqueous solution, followed in both cases by calcite precipitation. The first reaction is similar to that proposed by Thode et al. (1954) and Feely and Kulp (1957) for the origin of cap rock in salt domes. There is no evident role of elevated temperature in the replacement of gypsum by marble. The possibility emerges that stratiform carbonate rocks generated by calcitization of sulfate beds might exist in terranes other than the Carson region. By visual inspection, the rocks so produced are not easily distinguished from metamorphosed carbonate rocks of depositional origin.

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GEOLOGIC SETTING

Thrust-bounded tabular bodies (nappes) of carbonate rocks occur at different structural levels in several nappe piles of Jurassic age in the Carson Sink region of Nevada (Speed 1974a). Mesozoic pelitic rocks are the major constituent of the nappe piles, and the carbonate bodies represent no more than 1% or so of the Mesozoic terrane in the Carson region. The carbonate bodies are called carbonate breccia nappes because they contain breccia in generally greater abundance than unbrecciated rocks.

The most extensively exposed carbonate breccia nappe of the Carson region is the Muttelbury nappe whose rocks constitute the Muttelbury Formation (fig. 1). The Muttelbury Formation crops out in the West Humboldt Range over a strike length of 22 km. Its exposure area is 100 km², and its original extent could have been a factor of 5 to 10 greater (Speed 1974a). The maximum thickness of the Muttelbury Formation is about 70 m. Geologic and isotopic studies of carbonate breccia nappes in the Carson region have centered on the Muttelbury nappe, but reconnaissance of the other nappes tentatively indicates that hypotheses of origin of the Muttelbury Formation are applicable to the others. Further description of the Muttelbury Formation is given in the next section.

Thrust nappes of pelite with which the carbonate breccia nappes are intercalated represent slices from a thick and widespread terrane in the Carson region of Lower Mesozoic silicate mudstone, minor sandstone, and generally sparse dark clastic and algal limestone. No recognized carbonate units within positionally continuous sections of pelite can be correlated with rocks of the carbonate breccia nappes. Thus, the source of rocks in the carbonate breccia nappes must have been a unit (or units) other than the pelite succession.

The only nonpelitic Mesozoic formation of the western Carson region, the Lower

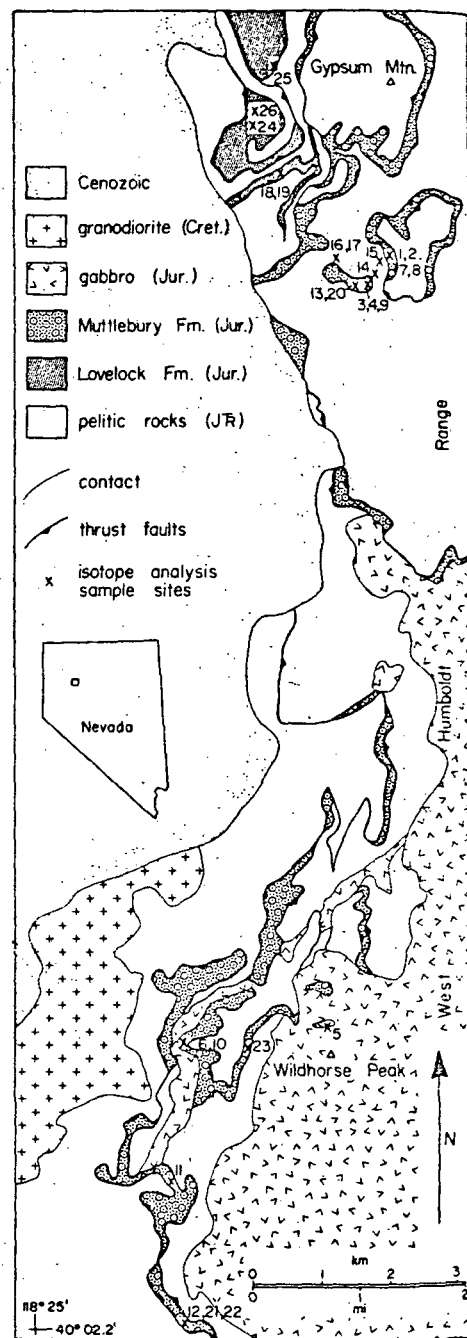


FIG. 1.—Geologic map of part of the West Humboldt Range, Nevada, showing outcrop distribution of the Muttelbury and Lovelock Formations and locations of specimens whose analyses are reported in table 1.

of Middle Jurassic Lovelock Formation (Speed 1974b), lies concordantly above the pelitic rocks and plays an important role in the origin of the Muttelbury Formation. The Lovelock Formation crops out only in a pair of nappes deep within the nappe pile of the northern West Humboldt Range (fig. 1). The Lovelock Formation includes a lower member of about 30 m of dark micrite that was probably deposited in lagoonal or subtidal euxinic conditions and a thin (5 m) middle member of calcarenite and calcirudite. The upper member is composed of at least 100 m of interbedded calcitic gypsum and calcarenite. Such gypsum constitutes the only evaporite rocks of known stratigraphic position within the Carson region. Also within the upper member are several lensoid bodies, up to 100 m long, of coarse-grained marble much like that of the Muttelbury Formation. These bodies occupy no particular stratigraphic position in the upper member. In fact, the largest body of marble in the Lovelock Formation occurs at the structural base of the formation where the upper member intersects the nappe bottom. Exposed contacts of marble are with surficially recrystallized gypsum, and primary relationships of the two rock types are uncertain. The distribution of marble in the upper member suggests generation of the marble may have been related to thrusting, but no more specific petrogenetic process emerges from field observations. It is noteworthy that in the gypsum-free lower and middle members of the Lovelock Formation, coarse-grained granoblastic marble is absent. Limestones of the lower two members are locally bleached and recrystallized, especially near calcite veins, but such metalimestones are texturally (and isotopically) distinct from marble of the upper member.

MUTTELBURY FORMATION

The Muttelbury Formation consists almost entirely of calcitic carbonate rocks in a melange of lensoid and tabular bodies

embedded in intraformational breccia. Individual bodies of unbrecciated rock are elongate parallel to formation strike and can be traced over distances from a few meters to more than 10 km. Most of the bodies comprise tightly interfolded, concordant layers of marble, graphitic crystalline limestone, and calcarenite. Two bodies are composed exclusively of little recrystallized micrite, and another contains only gypsum and interbedded calcarenite. Contact relationships and similarities of internal deformation of the unbrecciated rocks of the Muttelbury Formation suggest that such rocks had similar tectonic histories before breccia formation and could all have been initially in depositional continuity (Speed 1974a). Petrographic characteristics of each of the carbonate rock types are discussed below.

Marble: Marble consists of colorless, coarse-grained rocks characterized by smooth-walled mosaiced calcite. At places, marble has a well-defined grain shape foliation. Homogeneous marble contains calcite grains between 0.5 and 5 mm diam. Layered marble consists of homogeneous marble with intercalations 5-10 mm thick of darker, finer-grained marble containing up to 10% discrete sand-sized grains of quartz and albite. The thin layers contain calcite of 0.3-0.5 mm diam. Pyrite is prevalent along calcite grain boundaries, especially in the finer-grained layers, and opaque dust exists within and between calcite grains. The regularity of the layering and the concentration of quartz and albite in the fine-grained layers indicate initial bedding by composition. Petrographic examination provides an immediate but ultimately erroneous interpretation that the marble is a strongly recrystallized limestone locally with thin interbeds of sandy limestone.

Crystalline limestone: Crystalline limestone consists of dark grey, variably recrystallized, bedded carbonate rocks. Most of the crystalline limestone is comprised of highly intersutured calcite grains between 0.07 and 0.5 mm diam, together with scattered irregular calcite porphyroblasts up to 1.5 mm long. Moderately

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recrystallized calcarenite and calcirudite constitute scattered thin beds in the unit. The absence of relict skeletal, accretionary, or other sand-sized particles in much of the crystalline limestone, however, suggests the unit was chiefly derived from very fine-grained rocks.

Micrite: Micrite consists of dark grey, thin-bedded limestone of grain size 10 μ m or less, except for patchy zones of microsparite and sparite. Thin interbeds of calcarenite and calcirudite occur with micrite.

Calcarenite: Thin-bedded, fine-grained quartzose calcarenite occurs uniformly in sections up to 10 m thick. The texture of the calcarenite indicates probable slight recrystallization and strong compaction (Speed 1974a).

Gypsum: Gypsum is largely coarse-grained and granoblastic and commonly possesses grain-shape foliation. Diameters of gypsum grains in massive gypsum are up to 15 mm, but generally, 0.5–8 mm. Laminated gypsum consists of layers 1–2 cm thick of pure coarse-grained gypsum alternating with 2–10 mm thick laminae of finer-grained gypsum and up to 50% sand-sized calcite and quartz grains. Calcarenite interbedded with gypsum is fine-grained and thin-bedded and texturally resembles calcarenite in homogeneous bodies elsewhere in the formation.

Marble constitutes roughly 50% of the unbrecciated rocks and breccia fragments of the Muttlebury Formation. Marble occurs throughout the formation, and no gradations in its textural properties are recognized. In contrast, micrite occurs in the Muttlebury Formation only at the distal ends of the formation exposure area (fig. 1), and crystalline limestone only in the more central area. The occurrence of interbeds of similar distinctive calcarenite and calcirudite in both micrite and crystalline limestone and the proposition that crystalline limestone had a very fine-grained limestone parent suggest that the two limestone types are correlative. Further, the restricted distribution of the two limestone types is related to the metamorphic aureole of a Cretaceous

granodiorite (micrite outside, crystalline limestone inside), implying that the crystalline limestone is the product of Cretaceous heating of micrite, at least in part.

Breccia of the Muttlebury Formation: The breccia consists largely of unsorted polymictic fragments of the carbonate rock types described above. Gypsum clasts are absent. The breccia was created after the unbrecciated rocks had been folded and had been emplaced as the Muttlebury nappe (Speed 1974a). The matrix to the lithic clasts in the breccia is composed of sand-sized monocrystalline calcite and quartz grains. At places, the calcite-quartz sand increases in proportion to the degree that it is the framework constituent and lithic fragments float in it. The grain-size distribution of the calcite-quartz sand is apparently discrete from that of associated lithic clasts. Thus, the sand was not created by the same fragmentation that produced the lithic clasts. The rock fragments are clearly from intraformational sources, but there is no evident source for the calcite-quartz sand. An origin of the breccia of the Muttlebury Formation must account for the source and transport of calcite-quartz sand. This paper presents isotopic data which combine with other evidence to indicate an origin of the sand.

PROBLEM OF ORIGIN OF MARBLE

The texture of marble in the Muttlebury Formation suggests high-grade metamorphism of a limestone precursor. There is no evidence to support a metamorphic origin of the marble, however, in terms of metamorphic minerals. Unlike the relationship of micrite and crystalline limestone to granodiorite, the texture of marble is independent of proximity to intrusions. Moreover, marble is uniformly coarse-grained whether it contacts unmetamorphosed pelite or hornfels in adjacent nappes. Thus, if marble is a metamorphic product, the recrystallization occurred before the Muttlebury nappe arrived at its current position in Jurassic

time. Further, the degree of metamorphism between micrite and crystalline limestone was not uniform. The micrite before the Cretaceous was texturally distinct from the adjacent crystalline limestone. Petrographic evidence from the Muttlebury Formation indicates incompatibility of metamorphism.

Each of the marble in the Muttlebury Formation is represented by rocks of the Muttlebury Formation. The rocks of the Muttlebury Formation are correctly identified as initially in the Muttlebury Formation. Lovelock and Speed (1974) support the Muttlebury Formation as the predominant source of the Muttlebury Formation. The associated limestones of the Muttlebury Formation seem to be related to the Muttlebury hypothesis.

Data.— Rb-Sr and oxygen isotopic data from the Muttlebury Formation, and the results of the table 1 are indicated. The $\delta^{18}O$ values are usual for formation of the Muttlebury Formation (1.0‰). The $\delta^{18}O$ values are between 1.0 and 1.5‰. On a plot of $\delta^{18}O$ versus $\delta^{13}C$, the pairs fall in a line defined by a bound-

time. Further, there is strong contrast in the degree of apparent recrystallization between marble and associated carbonate rocks, and the evidence that crystalline limestone was probably unmetamorphosed micrite before contact metamorphism in the Cretaceous magnifies the original textural difference between marble and adjacent limestones. Thus, field and petrographic studies imply that the Muttletbury Formation contains carbonate rocks incompatible in their grades of metamorphism.

Each of the rock types other than marble in the Muttletbury Formation is represented among strata of the Lovelock Formation. Thus, if the unbrecciated rocks of the Muttletbury Formation are correctly interpreted as having been initially in depositional continuity, they are likely stratigraphic affiliates of the Lovelock Formation. The occurrence of gypsum in both formations strongly supports the correlation. Yet, marble is the predominant rock type in the Muttletbury Formation whereas in the Lovelock Formation, marble is minor and is associated with gypsum and not with limestones. Thus, marble in both formations seems to be an aberrant lithology, and progress in understanding the origin of the Muttletbury Formation demands an hypothesis of the origin of the marble.

ISOTOPE STUDIES

Data.—Ratios of stable carbon and oxygen isotopes in calcites from the Muttletbury Formation, Lovelock Formation, and pelitic rocks are presented in table 1 and figure 2. Specimen locations are indicated on figure 1. Ratios (R) of O^{18}/O^{16} and C^{13}/C^{12} are expressed in the usual form, $\delta X = (R_{\text{sample}} - R_{\text{std}}) R_{\text{std}}^{-1} (1,000)$, where X is the heavier isotope. Standards are the Pee Dee belemnite (PDB) for carbon and standard mean ocean water (SMOW) for oxygen.

On a plot of δC^{13} versus δO^{18} (fig. 2), pairs fall in two distinct groups identified by a boundary in figure 2. Group 1 ranges

in δO^{18} by 18‰ but only about 9‰ in δC^{13} . Group 2 consists of pairs with a clearly different trend, nearly constant δO^{18} but variable δC^{13} which ranges over about 16‰. Group 1 comprises analyses from a wide variety of carbonate rocks: micrite (21) and calcarenite (7, 12), calcite sand in the breccia (8-11), and calcarenite in the gypsum subunit (13, 20) of the Muttletbury Formation; micrite of the Lovelock Formation (18) and its recrystallized equivalent (19); and algal limestones (16, 17) within pelite below the Muttletbury nappe. Significantly, all of the analyses of marble (1-6, 22, 24-26) occur in group 2 together with those from a crystalline limestone (23) from the Muttletbury Formation and two clastic limestones (14, 15) from the same pelite nappe as the algal limestones of group 1 (16, 17).

Rocks from both the Lovelock and Muttletbury Formations occur in the two distinctive isotopic trends shown in figure 2. In particular, marble in both formations occupies the group 2 light carbon trend. The apparently parallel evolutions of C and O isotopes in the two formations strengthens the earlier proposition that the Muttletbury Formation is a correlative with the Lovelock Formation.

Isotopic trends in limestone and meta-limestones.—Several reference fields of isotope pairs are shown in figure 2 with which groups 1 and 2 may be compared. The field of modern marine calcite comprises isotopic compositions of calcites in equilibrium with HCO_3^- of shallow modern ocean water, modern shells from animals and plants that introduce no nonequilibrium effects during precipitation of carbonate (Weber and Raup 1966), and some Quaternary marine limestones (Keith and Weber 1964). This field approximates the initial composition of all marine carbonates, assuming the temperature and isotopic composition of sea water have not varied drastically through time. Fresh-water calcite, on the other hand, contains lighter and more variable ratios owing to the mixing of light oxygen rainwater and

TABLE 1

CARBON AND OXYGEN ISOTOPIC RATIOS OF CALCITES FROM THE MUTTLEBURY FORMATION, LOVELOCK FORMATION, AND LIMESTONE IN PELITE NAPPE SUBJACENT TO THE MUTTLEBURY NAPPE

Field No.	Analysis No.	Petrographic Description	$\delta^{18}\text{O}$ (SMOW) (‰)	$\delta^{13}\text{C}$ (PDB) (‰)
Muttlebury Formation:				
McB1a ...	1	Marble: coarse-grained, homogeneous	17.7	-23.2
McB1b ...	2	Marble: dark, fine-grained lamination	18.4	-21.5
McA3a ...	3	Marble: coarse-grained, homogeneous	19.0	-17.4
McA3b ...	4	Marble: dark, fine-grained lamination	18.9	-14.8
102	5	Marble: coarse-grained, homogeneous	14.4	-14.4
147	6	Marble: coarse-grained, homogeneous	17.1	-13.8
McB2b ...	7	Micrite	15.2	-1.0
McB2a ...	8	Calcite sand in breccia	15.2	-2.7
McA2a ...	9	Calcite sand in breccia	16.8	-0.5
148	10	Calcite sand in breccia	14.9	+0.1
9-127-20..	11	Calcite sand in breccia	10.3	-7.0
9-125-79..	12	Calcarenite	14.9	+0.2
McA1 ...	13	Calcite sand disseminated in gypsum	16.5	-1.2
10-24-22..	20	Calcarenite bed in gypsum	18.0	+1.7
9-125-79c..	21	Micrite	16.4	+0.4
9-125-96a..	22	Marble: coarse-grained, homogeneous	17.7	-24.0
9-131-31..	23	Crystalline limestone	19.5	-7.0
Lovelock Formation:				
9-145-6 ...	24	Marble of upper member	17.5	-10.0
9-145-11b	25	Marble of upper member at nappe base	18.6	-14.2
9-145-7b	26	Marble of upper member	16.9	-11.4
McDA ...	18	Micrite of lower member (microsparite)	18.2	+0.6
McDB ...	19	Recrystallized micrite of lower member	7.2	-5.3
Limestone in pelitic rocks:				
McAr ...	14	Grey calcarenite	17.0	-12.1
McAB1 ...	15	Grey calcarenite	20.0	-12.7
McC ...	16	Stromatolitic limestone; fine-grained	23.8	-0.2
McCA ...	17	Stromatolitic limestone; partly recrystallized	25.0	+0.4

NOTE.—Samples 8-11 are from secondary calcarenite and sample 12 from primary calcarenite in terminology of Speed (1974a).

light biogenic carbon in water saturated in atmospheric CO_2 .

The patterned field in figure 2 shows the range of 500 analyses from Phanerozoic marine and fresh-water limestones from Keith and Weber (1964). The line $\delta^{13}\text{C} = -2\text{‰}$ discriminates marine ($> -2\text{‰}$) from freshwater ($< -2\text{‰}$) limestones at about the 85% level. Thus, the trend of changing isotopic compositions of marine limestones through Phanerozoic time from their initial composition (modern marine calcite) is toward lighter oxygen over a range of about 15‰ (Keith and Weber 1964), but with little change in $\delta^{13}\text{C}$. A mode in the isotopic compositions of ancient marine limestones, recognized by

Clayton et al. (1966) and Sheppard and Schwarz (1971), is shown on figure 2 as normal marine limestone. The isotopic evolution of most limestone thus follows the path from modern marine calcite to normal marine limestone. There appears to be a wholesale exchange of oxygen isotopes in marine carbonate deposits during and after diagenesis with isotopes whose activities are externally controlled, presumably in flowing pore water. Such waters, however, apparently do not carry enough biogenic carbon to provide significant exchange of carbon isotopes with marine calcite (Clayton and Degens 1959). Elevation of temperature of carbonate rocks produces a variety of potential

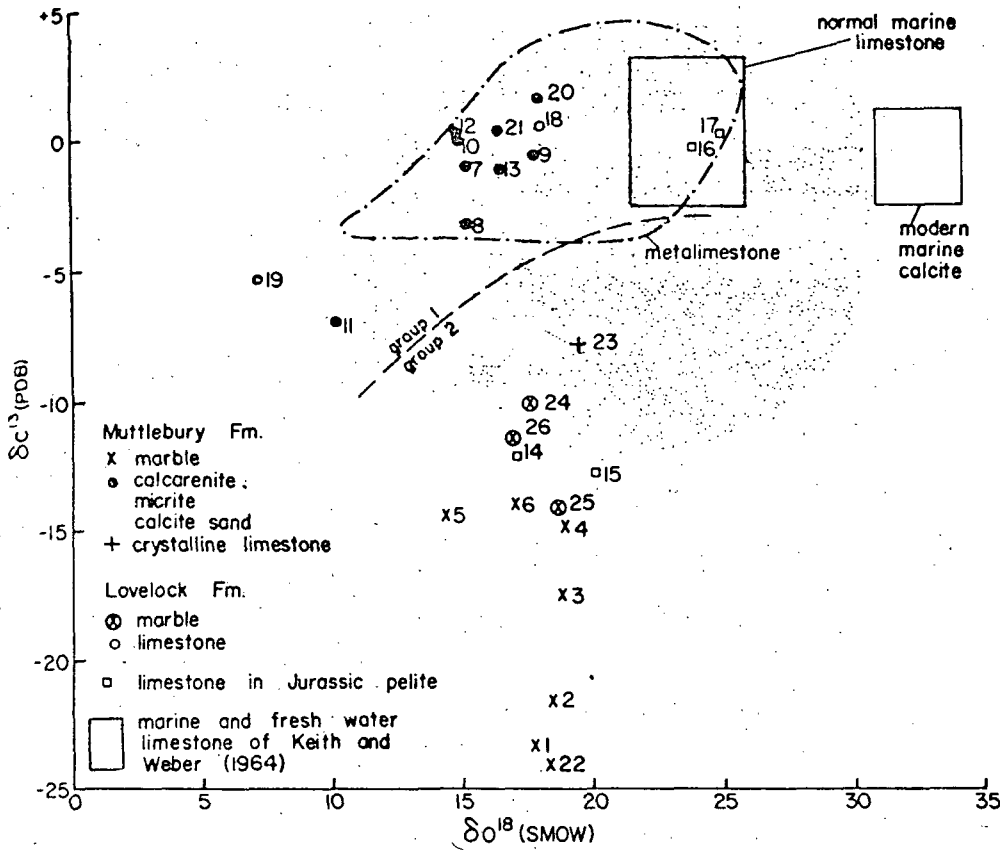


FIG. 2.—Plot of δC^{13} versus δO^{18} for specimens from the Muttlebury and Lovelock Formations and from limestones in pelitic rocks below the Muttlebury Formation. See table 1 for data and text for explanation of reference fields on figure.

isotopic exchanges. At constant composition, pure limestone will be metamorphosed with unchanged isotopic composition, whereas in impure limestone, silicate-carbonate oxygen exchange should lower δO^{18} of carbonate and graphite-carbonate exchange may lower δC^{13} of carbonate in limestone initially rich in organic substances. Metamorphism of carbonate rocks open to fluid transfer may yield isotopic changes owing to decarbonation and to isotopic exchange with through-moving fluid whose isotopic composition is not at equilibrium with the carbonate system. Decarbonation during metamorphism of impure limestone leaves the residual carbonate depleted in both C^{13} and O^{18} as

demonstrated by Shieh and Taylor (1969) and Sheppard and Schwarcz (1971). Hydrothermal metamorphism of limestone will yield δO^{18} of calcite that depends on the isotopic composition of the water and temperature, assuming the mass of fluid is large with respect to that of carbonate. At a given $\delta O_{H_2O}^{18}$, increasing temperature will progressively lower δO^{18} of calcite.

The field of isotopic compositions of metallimestones that contain minor silicate minerals is shown on figure 2; the data are from Engel et al. (1958), Shieh and Taylor (1969), and Sheppard and Schwarcz (1971). The metallimestone field essentially lies within that of unmetamorphosed limestone suggesting that such carbonate rocks are

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relatively closed systems during metamorphism. In most metamorphic carbonates, however, δO^{18} is shifted to the light side of that of normal marine limestone. Thus, in many marine carbonate rocks, trends in δO^{18} owing to exchange during diagenesis and burial are apparently continued during metamorphism, presumably due to exchange with water at elevated temperature. Trends in δC^{13} in metalimestones are vague except at the light δC^{13} , δO^{18} end of the field. There, decreases in δC^{13} and δO^{18} are concomitant, and such pairs are chiefly from metalimestones that underwent decarbonation.

Interpretation of group 1.—Isotopic pairs in group 1 trend from normal marine limestone through the metalimestone field to values of both δO^{18} and δC^{13} lower than those of the reference metalimestones. The majority of group 1 pairs cluster at $\delta O^{18} \sim 15\%$ and $\delta C^{13} \sim 0$. Such analyses (7, 12, 13, 16, 17, 18, 20, 21) are of limestones certainly or probably marine from direct sedimentary evidence, lithic associations, or correlations and of the problematic calcite sand (8, 9, 10) in the breccia of the Muttelbury Formation. The δC^{13} value in such rocks supports a marine rather than fresh-water origin. None of the rocks in this part of the trend has textures indicating significant thermal recrystallization. Thus, the isotopic changes in such rocks can be interpreted simply as due to oxygen exchange with meteoric water that is cold or perhaps at slightly elevated temperature. Two analyses (16, 17) of algal limestone in the pelite nappe subjacent to the Muttelbury nappe indicate the algal limestones have been lightened in δO^{18} less than other analyzed limestones. An explanation is that the flux of meteoric water through the algal limestone was less than in the other limestones owing to the relative impermeability of the pelitic rocks in which they occur as lensoid bodies.

Two points in group 1 (11, 19) represent the compositions of strongly recrystallized

rocks owing to local hydrothermal metamorphism. Analysis 19 is from a metalimestone that grades in a few tens of meters on strike to micrite (18) in the lower member of the Lovelock Formation. Analysis 11 is of recrystallized calcite sand in the Muttelbury Formation whose composition was presumably similar to analyses 8, 9, and 10 before hydrothermal activity. Thus, isotopic changes in the local recrystallization of limestones and calcite sand involve concomitant decrease in δO^{18} and δC^{13} and are probably due to decarbonation and oxygen exchange with hot water.

Interpretation of group 2.—Owing to light carbon, group 2 lies largely outside the isotopic field of marine and freshwater limestone. The isotopic composition of marble, among rocks of group 2, thus cannot have been inherited from a limestone precursor, and a metamorphic origin of the marble must have involved significant chemical transfer, at least of carbon. However, group 2 lies outside the field of metalimestone compositions (fig. 2). Marble of the Muttelbury and Lovelock Formations has therefore been generated under different conditions from marble that evolves from normal contact and regional metamorphism. Thermal decarbonation as a cause of the light carbon in group 2 can be dismissed because decarbonation should have produced concomitant reduction in δC^{13} and δO^{18} . Moreover, the constancy of δC^{13} in most group 1 analyses is hard to explain by any theory of preferential removal of C^{13} by metamorphism. Thus, the carbon isotopic composition of group 2 rocks must have arisen by the incorporation of light carbon from a source that was not initially the carbonate itself.

Biogenic carbon isotopic composition.—The reservoir of light carbon in the earth is chiefly organic material (petroleum, coal, dispersed matter in sediments) whose composition range is $-35 \leq \delta C^{13} \leq -22$ (Cheney and Jensen 1965). Therefore, calcite whose δC^{13} is lighter than that of

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freshwater limestone likely incorporated carbon oxidized from material of biogenic origin.

The subsurface evolution of biogenic carbon may follow a number of possible paths. Wholesale oxidation of organic substances at their source would yield CO_2 of bulk δC^{13} equal to that of the initial material. Hydrolysis and fluid transport of such CO_2 would provide light carbon for carbonate precipitation. Alternatively, organic compounds themselves may be transported in solution and suspension in subsurface fluids under reducing conditions and provide light carbon for calcite deposition when sufficiently high activities of O_2 and Ca^{+2} are encountered. Further, bacterial and thermal decomposition of organic matter produces gases, chiefly CH_4 and CO_2 , that could transport light carbon to a site of calcite precipitation. Between CH_4 and CO_2 at equilibrium, however, there is a large temperature-dependent fractionation of δC^{13} : ($\delta\text{C}_{\text{CO}_2}^{13} - \delta\text{C}_{\text{CH}_4}^{13}$) \approx 68‰ at 25°C, 49‰ at 100°C and 19‰ at 400°C according to calculations by Bottinga (1969). Thus, carbon in CH_4 may be far lighter than that of the source material, and oxidation of such methane could provide carbon for calcite as light as $\delta\text{C}^{13} = -80$ ‰, as proposed by Jensen (1968). Observations of the carbon compositions of coexisting CH_4 and CO_2 in gas in oil fields, bottom muds, and fumaroles indicate that very light carbon methane in fact exists, and that there is an approach to an equilibrium fractionation of δC^{13} in such gases as a function of temperature (Craig 1953; Murata et al. 1969; Bottinga 1969).

Light carbon carbonates.—Massive homogeneous calcite caprock of Gulf Coast salt domes contains carbon with $-48 \leq \delta\text{C}^{13} \leq -19$ (Thode et al. 1954; Feely and Kulp 1957). Light carbon calcite rock ($-43 \leq \delta\text{C}^{13} \leq +3$) is intercalated with sulfur in layers that are interpreted as beds in Sicily (Dessau et al. 1962; Jensen 1968). In both cases, light carbon calcite is thought to result from the oxidation of

organic carbon by sulfate under bacterial activation. Diagenetic light carbon carbonate minerals have been identified in noncarbonate rocks (Hodgson 1966; Hathaway and Degens 1968; Murata et al. 1969; Deuser 1970). Authors believe that the carbon in such minerals is biogenic and was transported in methane or in HCO_3^- that underwent isotopic exchange with methane. The deposition of calcite with biogenic carbon thus seems to be an established process, although it apparently occurs only under special conditions.

ORIGIN OF THE MARBLE

Precursor rocks.—The carbon composition of marble in the Muttelbury and Lovelock Formations requires that much of the carbon in some samples and nearly all in others is of biogenic origin. This conclusion invites consideration of origins of the marble in terms of precursor rocks and source and incorporation process of the biogenic carbon.

We submit that the marble is unlikely to have been derived from marine limestone because of the vast quantity of biogenic carbon required to change the carbon composition to the observed values. Conceptual carbon exchange between marine calcite and codepositional organic matter would require an unreasonably large proportion (>50%) of organic material in the original rock. Moreover, the marble contains no widespread graphite which would presumably result from copious organic matter. An exchange hypothesis also has trouble explaining the occurrence of bodies of massive marble in gypsum of the upper member of the Lovelock Formation and the absence of marble in the dark, presumably carbonaceous, limestones of the lower and middle members.

Isotopic exchange between marine limestone and biogenic carbon in a moving fluid would have required a large and unlikely throughput of carbon. The conceptual fluid presumably was chiefly water such that an enormous volume of

TABLE 2
REACTIONS REPRESENTING ALTERNATE PROCESSES OF CALCITIZATION OF SULFATE

	Standard Free-Energy Change (Kcal/mole CaCO ₃)
(1) CH ₄ + CaSO ₄ → CaCO ₃ + H ₂ O + H ₂ S	-12.5
(2a) 4C ₆ H ₁₄ + 19CaSO ₄ + H ₂ O → 19CaCO ₃ + 19H ₂ S + 5CO ₂	-18.7
(2b) 4C ₆ H ₆ + 15CaSO ₄ + 3H ₂ O → 15CaCO ₃ + 15H ₂ S + 9CO ₂	-16.2
(3) CaSO ₄ + 2HCO ₃ ⁻ → CaCO ₃ + 2HSO ₄ ⁻	-33.5

fluid would have passed through the carbonate body to provide the required mass of biogenic carbon. It is hard to see how the fluid could have been so channeled as to exchange carbon only with the marble precursor but allow the intercalated limestones of group 1 to maintain their original carbon compositions.

Because the texture of the marble implies far more complete recrystallization than in associated rocks, however, it is possible to appeal to selective exchange between carbon in a fluid and calcite undergoing solution and reprecipitation. If exchange was proportional to degree of recrystallization, an inverse correlation between δC^{13} and grain size might be predicted. However, no such relationship emerges from analyses of marble of different grain size. Fine-grained marble (2 and 4, table 1) has δC^{13} values close to those of adjacent masses of homogeneous marble and heavier than those of some coarser marbles at other locations. Moreover, under this hypothesis, one might expect δC^{13} values in group 1 to be generally lower and more variable.

The discussion above casts doubt on an origin of the carbon isotopic composition of the marble by incorporation of biogenic material in a marine limestone. The alternatives are to consider precursor rocks with light carbon or no carbon, either of which dismisses limestone from further attention.

Noncarbonate sedimentary rocks that conceivably could have given rise to the marble are silicate rocks, coal, and sulfate

rocks. Silicate predecessors can be eliminated, because quartz and feldspar grains in the fine-grained marble layers exhibit no reaction relations with calcite. Coal could have constituted beds that were transformed into light carbon calcite under appropriate externally maintained oxygen and calcium activities. The former existence of coal in the Muttletbury Formation has little geological support. Freshwater limestone is unrecognized and clays of a lagoonal or swamp origin do not occur. The former occurrence of coal in sulfate beds of the Lovelock Formation is unlikely. There is no significant amount of graphite in rocks of either formation, as might be produced by incomplete oxidation of coal. Generally taken, sulfate rocks constitute the only reasonable precursor to the marble.

Calcitization of sulfate.—We propose that bedded sulfate deposits were the forerunners of marble of the Muttletbury and Lovelock Formations and that gypsum and/or anhydrite were replaced by calcite with light carbon. Later considerations suggest that the organic matter was not initially codistributed with the sulfate but that biogenic carbon was transported into the sulfate bed. Calcitization of sulfate involving biogenic carbon from an external source could have been achieved by different processes that can be represented schematically by reactions 1 to 3 of table 2.

Reaction 1 represents the oxidation of methane from decomposing organic matter by dissolved sulfate when methane enters

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the sulfate bed. The calcium activity causes precipitation of calcite with carbon isotope composition derived from methane. Sulfur is lost as gaseous and dissolved H_2S , or if iron exists, pyrite will precipitate. Reaction 1 with such methane at $T < 50^\circ C$ would have produced calcite with lighter carbon than that observed in marble. Compositions of methane observed in geothermal discharges, however, are in an appropriate range ($\delta C^{13} = -16$ to -39% ; Bottinga 1969, table 10) for creation of the marble by reaction 1. Carbon dioxide associated with natural methane will be involved in calcitization by reaction 1 to the extent of the dissolution of CO_2 in an aqueous medium and would raise δC^{13} of calcite over that produced by methane alone.

Reactions 2 symbolize the bulk transfer of large hydrocarbon molecules to the sulfate rocks from decomposing organic matter. The hydrocarbons in 2 are selected among a myriad of possible molecules because hexane (2a) is abundant in petroleum and benzene (2b) is apparently particularly fugitive in fractionating organic substances. Large hydrocarbon molecules ($\geq C_6$) have carbon isotopic compositions similar to bulk organic matter (Silverman 1964) such that reactions 2 should produce calcite with δC^{13} between -22 and -35% . The highest carbon in our modest sampling is -24% .

The feasibility of reactions 2 is demonstrated by the emergence of oil and sulfur with variably cold and boiling water and steam from springs in volcanic rocks in northwestern Wyoming (Love and Good 1970). In two samples of such oil, largely alkanes, δC^{13} are -28.9 and -26.4% . Love and Good (1970) believed that oil in the springs is extracted by igneous waters from subvolcanic sedimentary rocks.

Reaction 3 represents precipitation in a sulfate bed owing to influx of a fluid bearing dissolved light carbon species derived from wholly oxidized organic material. Calcite precipitated according to reaction 3 will have a carbon composition similar to that of the organic matter. The

feasibility of reaction 3 is questioned chiefly in the unknown, and perhaps unlikely, conditions required to generate a fluid of appropriate composition. It would appear generally that rapid oxidation of organic material is necessary to produce high concentrations of dissolved carbonate species. Qualitatively, it is not clear that any subsurface fluid is sufficiently oxidizing. Moreover, the composition of the fluid will depend strongly on the extent of reaction of each unit of fluid with a given unit of organic material, and thus depend on the flux.

In summary, reaction 1 with methane of compositions found in geothermal discharges and reactions 2 and 3 all represent feasible mechanisms for calcitization of gypsum. Reactions 2 are less conservative of biogenic carbon than reactions 1 and 3, but there is little basis for choosing one mechanism as more likely than others. Except for reaction 1, elevated temperature is not apparently required for calcitization. Indeed, petrographic evidence that marble was initially associated with largely unmetamorphosed rocks implies that calcitization occurred at low temperatures.

Evidence for sulfate precursor rocks.—The existence of gypsum in both the Lovelock and Muttelbury Formations and the restriction of marble in the Lovelock Formation to the gypsum-bearing member make it highly plausible that sulfate beds were precursors of the light carbon marble. Both gypsum and marble have coarse-grained granoblastic thick homogeneous layers and thinner, quartzose fine-grained beds. Both rock types sharply contact limestones with smaller grain sizes and less thorough apparent recrystallization. Thin interbeds of calcarenite and isoclinal folds of beds occur in both gypsum and marble. In summary, the macroscopic evidence amply supports the hypothesis of a sulfate precursor. No alternate hypothesis can explain the textural contrasts and sharp contacts between marble and associated carbonate

UNIVERSITY OF MICHIGAN

rocks and the restriction in the Lovelock Formation of marble to the gypsiferous upper member.

On the microscopic scale, however, we have found no gypsum relics in marble or partially calcitized gypsum grains that would substantiate the proposed process. The existence of fine pyrite lining calcite grains in the marble, however, supports the idea that sulfur was produced during calcitization, as expressed by reactions 1 and 2 (table 2). The apparent absence of mixed gypsum-light carbon calcite rocks implies that calcitization proceeded by local complete replacement at a moving boundary rather than by simultaneous partial replacement, say along grain margins, throughout large tracts of gypsum. Exposed contacts between gypsum rocks and marble in the Lovelock Formation provide little evidence because they are products of modern surficial recrystallization of the gypsum rocks.

Other aspects of calcitization and a model.—The variation in δC^{13} in marble has no recognized correlation with petrographic properties. There is, however, an apparent relation between thickness and δC^{13} of marble: lightest carbon occurs where marble is thickest in the Muttelbury Formation and the heaviest, in the relatively thin marble bodies of the Lovelock Formation. On this basis, we propose that variations in δC^{13} in the marble resulted from different degrees of mixing of fluids bearing biogenic carbon with normal meteoric water carrying inorganic carbon. Where relatively large volumes of sulfate were calcitized, the input of fluids bearing biogenic carbon was greatest and there was minimum contamination by ubiquitous meteoric waters.

The constancy of δO^{18} in marble (fig. 2) indicates that the oxygen composition of the marble is independent of the causes of the large variability of δC^{13} in marble. Conceivable oxygen sources were an aqueous fluid, hydration water in precursor gypsum, and sulfate as in reaction

1a. The value of δO^{18} in marble, however, is similar to that of the main cluster of limestone analyses in group 1 (fig. 2), suggesting that the source of oxygen in the marble was more ubiquitous than the sulfate rocks precursor to the marble. Thus, an aqueous fluid is the most likely oxygen source, and it may be concluded that marble underwent oxygen exchange with water under similar conditions as the associated limestones.

The homogeneity and lensoid shape of the bodies of marble in the Lovelock Formation suggest that calcitization progressed out from fractures rather than occurring uniformly throughout the sulfate rocks. Furthermore, the occurrence of marble at the structural base of the Lovelock Formation where the gypsiferous member contacts the basal thrust supports the idea that calcitization was related to fracture zones. The implication is that fluids transporting biogenic carbon flowed through cracks.

A conceptual model of the process starts with initial pervasion of the sulfate beds by meteoric water carrying inorganic carbon. Such waters were subsaturated with respect to calcite. Then, creation of fractures allowed local ingress of waters containing abundant biogenic carbon in dissolved and particulate species from a different source. Mixing of the two fluids provided sufficient activity of oxidized carbonate species to precipitate calcite. Both the volume and carbon isotopic composition of calcite precipitated would have been proportional to the influx of fluid with biogenic carbon. Extensive calcitization occurred out from conduits carrying large volumes of fluid with biogenic carbon, and the calcite produced at such places has δC^{13} approaching that of the organic source.

Problem of detrital calcite in gypsum.—Detrital fine-grained calcite-quartz sand constitutes up to 50% of some laminae and thin beds of fine-grained gypsum in both the Muttelbury and Lovelock Formations

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isotopic analysis indicate the carbonate. Calcitization 1, 2 or 3 sl detrital calcite grains would quartzose thin identify detrital over, either to Judging by calc beds, a marble could ostensibly fine) calcite. T such marble wo between that of marine calcite analyses of adjacent, thin marble are given. In the fine somewhat heavier than the pre however, that isotopic samples demonstrate the ions of calcite.

SOURCE OF

Abundant of evaporites would produced light in gypsum. The disseminated moreover, of a sum beds imply probably derived deposits.

The euxinic the Lovelock F elements in the M possible sources accretion and they environment would have been present stones were in gypsum beds of carbon would estimate that

Isotopic analyses (table 1; 13 and 20) indicate the carbon of such calcite sand is marine. Calcitization of gypsum by reactions 1, 2 or 3 should not have affected the detrital calcite, and relict calcite sand grains would be predicted to exist in quartzose thin marble layers. We cannot identify detrital calcite in marble, however, either texturally or isotopically. Judging by calcite/quartz ratios in gypsum beds, a marble layer with 5% quartz could ostensibly have 50% detrital (marine) calcite. The carbon composition of such marble would reasonably be midway between that of homogeneous marble and marine calcite of group 1. Two paired analyses of adjacent fine-grained, quartz-bearing, thin marble layers and coarse marble are given in table 1 (1 and 2; 3 and 4). In the fine-grained marble δC^{13} is somewhat heavier (by 1.7 and 2.6‰), but less than the predicted value. It is possible, however, that both petrographic and isotopic sampling are not sufficient to demonstrate the existence of two generations of calcite in the marble.

SOURCE OF BIOGENIC CARBON IN MARBLE

Abundant organic matter within the evaporites would seem likely to have produced light carbon calcite disseminated in gypsum. Thus, the apparent absence of disseminated replacement calcite and, moreover, of abundant graphite, in gypsum beds implies that biogenic carbon was probably derived outside the evaporite deposits.

The euxinic micrite and calcarenite of the Lovelock Formation and their equivalents in the Muttletbury Formation were possible sources because they are carbonaceous and they accumulated partly in an environment where organic matter could have been preserved. Further, such limestones were initially coextensive with the gypsum beds so that only local transport of carbon would have been required. We estimate that the limestones of the

Lovelock Formation would have had to contain an average of 3 wt % organic matter to supply enough biogenic carbon to form all the marble of the Muttletbury Formation. Replacement was not likely to have been perfectly efficient, and organic contents required of the limestones would have exceeded 3%. Such proportions of organic matter, however, are apparently rare in limestone (Ronov 1958; Graf 1960; Hunt 1961), and it is doubtful that the limestone of the Lovelock Formation could have been the total source of biogenic carbon.

The enormous volume of Jurassic and Triassic pelitic rocks of the Carson region forms yet another potential source of biogenic carbon. The carbon isotopic compositions in the pelite of the Carson region are presumably in the same range as those in analyzed pelitic rocks from other areas ($-23 > \delta C^{13} > -32$, Schwarcz 1969; Barker and Friedman 1969). The pelites may have released a large volume of carbon as they dewatered, compacted, and locally developed slaty cleavage during Middle Jurassic folding and nappe transport (Speed 1974b). The numerous thrust faults within the Mesozoic terrane of the Carson region may have acted as conduits collecting and transporting biogenic carbon moving out of adjacent tracts of pelite.

We envision that substances bearing biogenic carbon were flushed out of limestones associated with the gypsum beds and/or the pelite terrane during deformation and nappe translation. Faults penetrating the gypsum unit provided the fluid conduits from which calcitization of gypsum proceeded outward.

CARBON IN LIMESTONES OF GROUP 2

The crystalline limestone of group 2 is a metamorphosed micrite that was probably equivalent to micrites in the Lovelock and Muttletbury Formations (table 1: 7, 18, 21). Thus, carbon in the crystalline limestone has been reduced about 7‰ relative to its

presumed initial value. The change in δC^{13} was probably not due to thermal decarbonation because, considering concomitant lowering of δO^{18} in such decarbonation, the initial composition would have been near modern marine calcite (fig. 2), a value considerably different from that of the analyzed micrites. It seems more likely that during metamorphism, calcite precipitated with lighter carbon than in the original micrite because of mixing of HCO_3^- in meteoric waters, local oxidation of organic matter in the micrite, or conceivably, by equilibration of carbon isotopes with adjacent marble.

Remarkably light carbon occurs in samples (14, 15) of clastic limestones from the pelite nappe that lies below the Muttletbury nappe but above the nappe containing the Lovelock Formation. The samples are from different beds, each about 1 m thick, of carbonaceous calcarenite containing nearly 50% quartz and feldspar sand. The calcite is recrystallized, and textures provide little information on the nature of earlier rock. The recrystallization is not, however, a product of metamorphism because adjacent pelites are unmetamorphosed and because algal limestones (16, 17) of group 1 occur stratigraphically within 50 m of these calcarenites. The geologic setting makes it unlikely that such beds contained gypsum, now calcitized as in the marble. Studies by Murata et al. (1969) and Hathaway and Degens (1969) lead to the idea that much of the carbonate in these clastic rocks was cement or a diagenetic replacement. In

both cases, biogenic carbon from the calcarenite beds or the adjacent pelite may well have been incorporated in the nondetrital carbonate. Recrystallization, however, makes proof of the hypothesis difficult.

PETROGENESIS OF BRECCIA

The conclusion that marble in the Muttletbury Formation is calcitized gypsum indicates that sulfate rocks were once a major constituent of the formation. The existence of gypsum in both the Muttletbury and Lovelock Formations, however, shows that calcitization did not go to completion. Gypsum was perhaps widespread in the Muttletbury Formation at the time of generation of intraformational breccia: The breccia, thus, could be a solution breccia, and field and petrographic studies support such an origin (Speed 1974a, 1974b). The problematic calcite-quartz sand in the breccia therefore likely represents grains originally disseminated in the gypsum and accumulated on dissolution of the gypsum. The carbon isotopic composition of calcite sand in the breccia (table 1: 8-11) is like that of calcite sand currently disseminated in gypsum (table 1: 13) and limestones of group 1. The sand in the breccia is thus clearly not a product of calcitization of gypsum along with marble.

ACKNOWLEDGMENTS.—We are grateful to John R. MacMillan for assistance in the fieldwork. Work by Speed was supported by NSF Grant GA1574; work by Clayton was supported by NSF Grant GA 22711.

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block are autochthonous; those in the western block are bounded by normal faults; the problem is the initial relationship of the two sequences.

The eastern block contains the terrigenous Grass Valley and Winnemucca Fms. Between them is the Dun Glen Fm., 170 m thick and 85% limestone, largely skeletal micrite with local in place benthonic faunas that indicate a shallow shelf environment. Rocks of the eastern block are gently folded north of Fencemaker Canyon, but they are increasingly deformed to the south over 6 km beyond which they lie below an allochthon of basinal Triassic pelites. Exposed in the western block are an estimated 1500 m of intercalated terrigenous rocks and sparsely fossiliferous micrite that contains age equivalent faunas to the Dun Glen Fm. The limestone totals about 100 m in thickness, and contrasts further with the Dun Glen in having typically thinner individual units, far more intercalated terrigenous rock, and no recognized in place fossils. Beds of the western block are in refolded recumbent isoclines.

Lithic differences between the two successions suggest the western block contains more offshore deposits. Structural contrasts between them imply the western is probably allochthonous, and has undergone shelfward transport relative to the eastern. We believe the western block is probably a remnant of the Fencemaker allochthon, an extensive terrane of basinal Triassic rocks that overrode the platform, probably in Middle Jurassic time.

Stillwater
Soda Lake
Churchville
Nev.

HYDROLOGIC RECONNAISSANCE OF GEOTHERMAL AREAS IN BLACK ROCK DESERT AND CARSON DESERT, NEVADA

Olmsted, F. H., Water Resources Division, U. S. Geological Survey, Menlo Park, California 94025

Two geothermal areas in northern and central Nevada were selected for a preliminary hydrologic reconnaissance: (1) Black Rock Desert, particularly the Gerlach KGRA (Known Geothermal Resources Area), where the discharge area includes hot springs; and (2) Carson Desert, particularly the Stillwater-Soda Lake KGRA, where two known principal hydrothermal systems do not include hot springs but are indicated from wells and other indirect evidence. The hydrothermal areas were outlined by means of test drilling; temperatures at depths of about 50-150 feet (15-46 meters) and geothermal gradients were used to delineate areas of large heat flow associated with rising thermal ground water. At Gerlach, in the Black Rock Desert, the hydrothermal discharge is estimated to be about 1,500-2,000 acre-feet per year (1.8-2.5 cubic hectometers per year). Geochemical data indicate a temperature of about 160° - 180°C for the source of Great Boiling Springs, the principal outlet of the system. Within the Stillwater-Soda Lake KGRA in the Carson Desert, the two hydrothermal systems delineated were: (1) one centered at Stillwater, and (2) one between Soda Lake and Upsal Hogback, about 8 miles (13 kilometers) northwest of Fallon. In both systems the thermal water appears to rise along concealed fault conduits into near-surface aquifers, whence it moves horizontally along the hydraulic gradient. At Stillwater, the maximum temperature of the rising water exceeds 156°C (geochemical data); at Soda-Lake-Upsal Hogback, the maximum temperature is not known but it probably is much more than 100°C.

METAMORPHIC, IGNEOUS

DEATH VALLEY, CALIFORNIA

Otton, James K., D.

University, Univ.

The complex of metamorphic rocks in the Death Valley area can be subdivided into a younger success and a younger schist. The complex is rite previously assigned to a relatively unmetamorphosed bonate-clastic sequence. The metamorphism related to the lithologically similar Pahrump units in the area was later intruded by plutons largely or wholly with the latest plutonic complex they intrude with northwest crustal plutons and intruding the are commonly subhorizontally. Folding of the metamorphic rocks coincide with the well known plutons and probably the normal faults postdate the late Miocene in age. The Black Mts. probably be

A TECTONIC MELANGE ON COLUMBIA, A FRANCISCAN

Page, Richard J., D.

of Washington, Se

TECTONIC BLOCKS (< 4 m) of primary-lithic sandstones, out as coastal exposures on the Gorda Peninsula. Several tectonic blocks include: (1) metabasites, organized and (2) mafic gneiss (serpentine-bearing limestone blocks of late Miocene age), are cut by brittle sheared limestone and sheared matrix of metabasites and boudined radiolarites and sheared metabasalt. A trending thrust fault juxtapose major blocks.

The chaotic structure of rocks of differing ages and compositions, including perthite metagreywackes, are characteristic of the unit in tectonic contact with the Lewis and Clark Inlet fault complex. Within this

February 21, 1979

MEMORANDUM

TO: Dennis Nielson & Howard Ross
FROM: Bruce Sibbett
SUBJECT: Report on Trip to Soda Lake, Nevada

While at the Nevada Bureau of Mines and Geology. I learned that a second open file report on the Carson Sink #1 hole (NURE Project) was available. A copy of this report was obtained and the cuttings on file were compared with the lithologic descriptions. This log was found to be satisfactory. The report includes thin section descriptions also. A slice of the core from the bottom of the hole (8502') was obtained. The lithologic descriptions for the U.S.G.S. O.F. Rep. GJBX - 53(78) were done by Ed Bingler, formerly with the Nevada Bureau of Mines, now with the Montana Bureau of Mines. Ed Bingler has logged the cuttings from many holes in the Carson Desert, including Chevron's. He will be publishing a report on correlation within the sink, but the date has not been set.

I talked with Dick Benoit and Bob Forest of Phillips at their office in Reno. Dick Benoit has been in charge of Phillip's work at Brady (Desert Peak) KGRA. He will publish a case study based on Phillips information next fall as a Nevada Bureau of Mines & Geology publication. Included with this publication will be the mapping which John Heiner did for his thesis. Heiner will have his thesis defence in a couple of months, after which it should be public information. Heiner is working for Phillips and was out of town when I was there. John Heiner will be giving a talk on Desert Peak at the Reno GRC. next September.

I discussed geophysical methods in geothermal exploration with Dick Benoit and his boss (district geologist?) Bob Forest. They have looked at Chevron's data for Soda Lake. They made the comment that seismic work was of "no value" in the Basin and Range Province. Benoit said that Phillip's vibroseis survey at Roosevelt Hot Springs failed to even find the range front fault. They also felt that MT was not useful. Dick Benoit said that he "hadn't seen a MT survey in the Basin and Range that didn't show a magma chamber at depth", implying that the method or its interpretation indicated magmas which did not actually exist. In reference to resistivity they thought that there were too many near surface factors such as salinity variation, surface thermal springs, etc. for the method to indicate the thermal reservoir. Rush (1972, Hydro. recon. Soda Lakes) reported that 900,000 tons of salts have been lost from Soda Lake in the last 70 years. This salt has probably moved into aquifers to the north.

I would strongly recommend that you call Mr. Forest and Benoit (702-386-2273) and discuss their feelings about seismic and other methods in geothermal exploration.

The most interesting thing that I found in the field was several small areas of silica cemented sand north of Soda Lake, near the Soda Lake 1-29 hole. Some opal or Chalcedonic material is present as well as silica replaced wood and grass. These outcrops and the small area of hematite stained alteration around the old steam well were mapped.

Bruce Sibbett

BS/kg