THE ORIGIN OF CRYSTALLINE MAGNESITE DEPOSITS

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CONTENTS

ILLUSTRATIONS

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ABSTRACT

Although the formation of magnesite by a hydrothermal-metasomatic process is possible its operation requires nicely balanced conditions which are difficult to maintain. On examination of the operation of this process in the field it is found to produce, with minor and doubtful exceptions, brucite rather than magnesite.

Crystalline magnesite is most often found in dolomite formations of heterogeneous sedimentary sequences. These are characteristically laid down in the shallow unstable shelf areas of geosynclines. It is suggested that magnesite is also deposited under like conditions in restricted embayments in the shelf in which the magnesium content of the sea water is locally enriched by reworking and dissolution, in the presence of sulfate, of previously or contemporaneously deposited dolomite. Dolomite may be forming diagenetically in the same embayment from aragonitic muds. Diagenesis takes place in the disordered state in the transition from aragonite to calcite. In this interval magnesium is incorporated in the lattice of the calcite. The dolomite formed, being far less soluble than the other carbonates, is concentrated when the excess calcite is dissolved by reworking of the muds.

In such environment, the initial precipitation of magnesia, facilitated by organic activity, is probably in the form of a basic hydrous magnesium carbonate which is converted to the anhydrous form by an increase in sedimentary loading or removal to deeper parts and higher CO_2 concentrations of the embayment by local currents. It may also be converted to magnesite in the carbonate muds at the site of precipitation in the same manner that geodes are formed.

The magnesite at Gabbs occurs in the topmost dolomite member of the Luning formation which is part of a 10,000 foot section of sediments consisting of alternating sequences of conglomerate, argillite, limestone, and dolomite deposited in a local eastward embayment of the Upper Triassic sea. These sediments are folded and faulted, and intruded by a variety of dikes and a stock of granodiorite. The magnesite antedates the intrusive and orogenic history because it is already involved in all of it. The principal effect on the magnesite has been to dolomitize it along stressed zones or to convert it to brucite in pendants of the carbonate engulfed in the granodiorite. The formation of brucite is accompanied by the development of relatively minor amounts of magnesium silicate minerals.

The deposits at Cranbrook, British Columbia; Chewelah, Washington; and Kilmar, Quebec, in spite of their complexity, show striking parallels to those at Gabbs. The extensive development of diopside and serpentine at Kilmar is explained by a process of autosilication.

An examination of magnesite deposits in interior basins supports the evidence from marine environments that they were laid down contemporaneously with the sediments, and that the magnesia was derived from sea water or from the weathering products of magnesia-rich rocks brought into the basins by sedimentary processes and not introduced by hot springs or other hydrothermal sources.

A brief examination of the formation of magnesite in serpentine reveals here also the inadequacy of the hydrothermal hypothesis. Weathering processes with subsequent transformation, purification and concentration of the first formed basic magnesia compounds by several cycles of solution, migration and precipitation offer a more reasonable explanation. The reasoning is extended to the deposits at Currant Creek, Nevada, the "show piece" for hydrothermal replacement. Here the deposits are more adequately accounted for by a two-step process - the processes operating to produce bone magnesite in serpentine working on magnesia-rich, tuffaceous, carbonate sediments.

It is concluded that magnesite deposits of any consequence formed by hydrothermal or replacement processes are extremely rare.

INTRODUCTION

The mineral magnesite, $MgCO_3$, or more correctly, $(Mg, Fe)CO_3$ as iron replaces magnesium to some extent in all natural magnesite, occurs naturally in two principal forms known as "bone magnesite" and "crystalline magnesite". Crystallographically identical, they differ in texture, to some extent in hardness and color, and in mode of occurrence.

The cryptocrystalline "bone" variety is found most often in veins and irregular masses in altered basic igneous rocks such as serpentine ($Mg_3Si_2O_5(OH)_4$) from which it is undoubtedly derived. Under the proper conditions, bone magnesite may also be formed by weathering from other rocks high in magnesia. It is usually a hard, dense, porcelaneous appearing material which breaks with a conchoidal fracture. Generally it is nodular in structure, white or creamy white in color but stained quite often by limonite on surfaces exposed to weathering or subsurface solutions. The principal contaminents in bone magnesite deposits are silica and lime with lesser amounts of iron and alumina. The origin of bone magnesite is a separate although related problem which will be touched upon only briefly. One might, however, mention at this point the host of hydrous basic magnesium carbonate minerals of which hydromagnesite ($3MgCO_3(MgOH)_2 3H_2O$)

is typical, as well as the most important geologically, which are related to both bone magnesite as well as crystalline magnesite as intermediate products in their formation. Hydromagnesite is a soft, white finely crystalline, sometimes clay-like, mineral. It occurs most often as the principal ingredient in a mixture of such carbonates in the soils, or in local deposits in shallow, undrained, swampy basins in areas of serpentine rocks. In addition to the carbonates the mixture may also contain opaline silica, brucite ($Mg(OH)_2$), and a variety of hydrous magnesium silicate minerals, such as deweylite ($3MgSiO_3Mg(OH)_25H_2O$). In a different geologic setting but with similar mineral associations, hydromagnesite is also the principal product resulting from the weathering of brucite at Gabbs, Nevada. It is here an intermediate product between brucite on the one hand and bone magnesite on the other.

The "crystalline" variety of magnesite forms by far the most extensive deposits. The following will be concerned principally with these. They generally, though not always, occur within dolmite horizons of sedimentary sequences. Most have been intruded by a variety of igneous rocks and operated upon by several geologic processes. They may be folded, faulted, and to some extent, metamorphosed and recrystallized so that the magnesite is rarely distinguishable from the associated dolomite. This may also largely obscure, if not obliterate, most of the original structures.

The deposits are generally aligned with some dominant structural trend most often paralleling the trend of the enclosing rocks. Superimposed upon this is the structural complexity resulting from folding, from several periods of faulting and from the emplacement of several generations of intrusives. This picture may be further complicated by extensive thrusts and by the effects of hydrothermal, metamorphic, and metasomatic processes.

Complexity, however, does not end there. The deposits are rarely uniform throughout. Bands of dolomite within the magnesite are common and the change from magnesite to dolomite may be abrupt although it is frequently gradational. Textural and color changes are, for the most part, unrelated to changes in composition but in a few deposits, particularly some in Austria, the color and texture of the magnesite are distinctive enough to be used in sorting the material underground. Textures vary from coarse and granular to aphanitic. The colors range from reddish pink through buff and gray to almost black. Weathering may produce a sandy, pitted surface colored, almost invariably, some shade from buff to chocolate brown.

Mineralogically the deposits are relatively simple consisting overwhelmingly of crystalline magnesite and dolomite. Calcite, though present in some, is rarely abundant. Proliferation of mineralogical detail is found when one examines them more closely. This is the result of the many processes that, however, effect the magnesite only locally. Tremolite, diopside, and garnet, and occasionally periclase, may reflect the effects of contact metamorphism. Serpentine, forsterite, and brucite are the results of metasomatims. Vein dolomite, vein serpentine, talc, rarely calcite, some kaolinite, possible deweylite, and perhaps some bone magnesite, are produced by hydrothermal processes. The above is only a partial list of minerals but it indicates the range of possibilities.

Out of this profusion of detail must be culled the essence of our idea as to how the deposits came into being. Up until a few years ago there was hardly a doubt that they were the result of metasomatic replacement of dolomite by magnesite. The source of the magnesia was at first attributed to magmatic activity which also furnished the driving energy for the process. Since the work of Faust (1949) and others demonstrated the possibility of deriving magnesia-rich solutions by natural processes analogous to the partial calcination of dolomite to periclase and calcite and the dissolution of the periclase by hot carbon dioxide charged magmatic waters, this process has been repeatedly invoked to account for practically all magnesite deposits. Recently Bodenlos (1954) has applied it with indifferent success to account for the crystalline magnesite deposits in the Serra das Eguas, Brunnado Bahia, and elsewhere in Brazil. This point of view has been so often examined, fortified, and refined in the literature that it needs no further documentation. This hypothesis rests largely upon three elements: a source of magnesia such as dolomite, magmatic activity and emanations which presumably operate upon the dolomite and furnish the driving energy for the replacement process, and the evidence for the operation of this process in the deposits themselves.

It will be the burden of this paper to show that whereas magnesite bodies might originate in the manner outlined above the process does not adequately account for the deposits at Gabbs, Nevada, and further, for similar deposits elsewhere. To this end the occurrences at Gabbs, with which the writer is most familiar, will be described in some detail. Brief descriptions will also be given of the deposits at Cranbrook in British Columbia, those at Chewelah, Washington, and at Kilmar, Quebec, all of which the writer has visited. Structurally and petrographically they vary from the simplest to the most complex and together cover a large part of the range of occurrences. In spite of their diversity, however, the writer hopes to show that all owe their origin principally to sedimentary processes and that the metasomatic and other effects often observed have been imposed, for the most part, after the deposits were emplaced.

THE MAGNESITE DEPOSITS AT GABBS, NEVADA

Geographic and Economic Setting

The deposits at Gabbs, Nevada, are located on the west flank of the Paradise Range in the northwestern Nye County. The principal bodies occupy an area roughly one mile long by a half mile wide, and lesser occurrences have been found in a belt more than four miles long and at places almost two miles wide. The reserves of magnesite containing less than 5% lime were estimated by Callaghan and Vitaliano (1948) to be in excess of 25,000,000 tons. Reserves of the better grade material are, of course, considerably less. More recent estimates for the entire area are, not, as yet, available.

Since their discovery in 1927, almost two million tons of magnesite and somewhat less than a million tons of brucite have been mined from these deposits and processed into a variety of granular refractory materials which are used as furnace linings by the steel industry. In addition, close to a million tons of magnesite were used during World War II for the production of magnesium metal. Magnesium oxide from Gabbs is also going into magnesia cements, sound proofing and insulating board, and into the processing of uranium ores.

Previous Geologic Investigations

The systematic exploration of the deposits was begun in

1930, and a preliminary geologic study was made by Eugene Callaghan (1933) of the United States Geological Survey in cooperation with the Nevada State Bureau of Mines. During World War II, Callaghan and Charles Vitaliano, assisted by other members of the Survey, made a more detailed investigation. The result of this work has been published by the United States Geological Survey as Mineral Investigations Field Studies Map MF 35 (1956), a reduced reproduction of which is appended as Plate I for convenience.

Subsequently Vitaliano et al (1957) extended the area previously mapped to about 105 square miles of the northwestern part of the Paradise Peak quadrangle which includes the magnesite deposits. This map MF 52 is reproduced as Plate II. Neither map MF 35 nor MF 52 shows the finer structural details later discussed. These have not, as yet, been completely unravelled.

The stratigraphy of the Mesozoic rocks in the area has been described by Muller and Ferguson (1939) and a petrographic description of the igneous rocks is given by Callaghan (1935). Ferguson and Muller (1949) have also outlined the structural development of the region in some detail. The location of the area and the outline of the Luning Embayment, dotted where it is not definitely known, are shown on Fig. 1.

Geologic Setting of the Magnesite at Gabbs

The rocks on the west flank of the Paradise Range in the vicinity of Gabbs consist of a 10,000 foot sequence of metamorphosed dolomites, limestones, and shales with minor conglomerates of Upper Triassic age which are overlain conformably by Lower Jurassic shaley limestone and dolomite. These were laid down in a shallow, restricted easterly arm of the late Triassic sea which Ferguson and Muller (1949, p. 8) have named the Luning Embayment. Our story begins with the unconformity at the base of the Upper Triassic, which is not exposed in the Gabbs area. Elsewhere in the embayment it is usually marked by a basal conglomerate of chert, hornfels, and metavolcanics derived from the Middle Triassic Excelsior formation upon which the Upper Triassic, the Luning formation (Muller and Ferguson, 1939, p. 1594), rests with some discordance. Upward from the base, this lowermost of the three members of the Luning grades into argillite and thin dark limestone, and in some areas, siliceous dolomite. The middle member is characterized by thin bedded green limestone and black shales alternating with conglomerates. Occasionally it also contains metamorphosed tuff and flows. (Vitaliano, et al 1957), (Ferguson & Muller, 1939, p. 1596). The uppermost member is represented by two facies.

In some areas it consists principally of dark gray dolomite and subordinate limestone. In others, it is made up largely of black, thin-bedded limestone, lesser argillite, and occasional sandstone. The dolomite contains the magnesite deposits. Overlying these are the dark, carbonaceous siltstone and shale of the uppermost Triassic Gabbs formation which grade upward into dark thin-bedded limestone and dolomite. The lowermost Jurassic Sunrise formation, which is similar to the Gabbs in lithology, overlies the Gabbs formation, without a depositional break. This is overlain in turn - unconformably in most places - by the Dunlap formation which consists largely of conglomerates. The appearance of this coarse clastic material heralds the renewal of warping and folding which ultimately culminated in thrusting on a grand scale in early Jurassic time. This was already intimated by the previous instabilities as suggested by the many facies changes, particularly throughout the 8,000 feet of Luning deposition. Muller and Ferguson (1939, p. 1595) comment on this in the following words:

"Scattered occurrences of intercalated coral reefs over the entire area suggest general shallowness of the sea with some irregularities of the strand line.----The area in which the Luning sediments were deposited was dominantly sinking in Luning time, but locally there were marked oscillations evidenced by a frequent recurrence of conglomerates in the near shore belt, by repetition of banks of Alectryonia montis-caprilis, and by a repeated occurrence of thin coral reefs."

Evidence that this instability persisted to the end of the Luning deposition can be found also in the uppermost dolomite that, at Gabbs, lies in the overriding plate of the Paradise thrust. This dolomite is a dark, almost black, very fine-grained to dense, medium- to thick-bedded rock, somewhat metamorphosed as indicated by occasional tremolitic zones and scattered tiny flakes of talc. There are also more intensely silicated horizons of which more will be said later. Locally the dark, dense dolomite may be recrystallized to a dense white marble. There are thin, persistent, layers of black hornfelsic shale, and siliceious beds containing fragmented, lenticular, nodules of chert. Faults of small displacement involving only a few beds are also common. And not too rare are beds of dolomitic or chert conglomerate. Weathering occasionally brings out the ghosts of plastically deformed, blocky, intraformational breccia, and at one place in Tungsten Gulch a thin layer of quartzite is preserved in a small, overturned intraformational fold.

Structural Development

The structural history of the region has been described in some detail by Ferguson and Muller (1949, p. 9) and only the broad outline beginning with the development of the Luning Embayment will be given here. We are most concerned, of course, with the post-Luning events and how these affected the Luning formation and the magnesite deposits which are found within the uppermost dolomite member.

The end of the folding, as indicated by the unconformity at the base of the Luning, which involved the Middle Triassic Excelsior formation and older rocks and marked the beginning of Luning deposition has been mentioned as well as the repeated uplift of the bordering lands and warping within the embayment throughout Luning time.

"The diastrophism of Jurassic age was superposed on rocks which were previously folded, and this Jurassic folding and thrusting appears to have been most intense in the area of the Luning Embayment" (Ferguson and Muller, 1949, p. 7).

The warping, local folding, and thrusting during Dunlap time affected first the Luning sediments in the southern part of the Luning Embayment and spread northward during repeated phases of increasing intensity and complexity, affecting not only the Luning rocks but older sediments as well. This was accompanied, in some areas, by volcanism which probably began shortly after the close of Sunrise deposition (Ferguson and Muller, 1949, p. 10). Much of the detailed structure in the area of the magnesite deposits at Gabbs remains to be unravelled. Only the latest and most obvious faults are shown on Plate I.

The major events during this period of orogeny were the complex folding of the sediments generally along a north-northwesterly trend, and the development of the gently westward-dipping Paradise thrust (Plate II) which traverses the area in a north-south direction. A considerable displacement is involved (6 miles or more, Ferguson and Muller, 1949, p. 40). Along it, dolomite, limestone, and slate of the uppermost Luning formation override the limestone and slate of the Luning as well as the Jurassic rocks. Lesser thrusts associated with it imbricate the dolomite sequence of the upper plate. There is evidence within the latter of at least moderate normal faulting prior to the thrusting. Vitaliano, et al, (1957) also mention pre-thrust normal faulting of considerable magnitude in the Paradise Range farther to the east. Later thrusts, (Plate II), both to the north and south of Gabbs, cut and override the Paradise thrust and bring the Excelsior and lower members of the Luning formation over the upper dolomite member.

Except for the faulting associated with the development of the Basin Range structure, which is still continuing, post-granodiorite movement has been persistent, but generally not very extensive. This is illustrated best in the magnesite deposits and will be treated more fully in connection with them.

In the Gabbs area in general, the regional structural trends - those of the major faults, folds, and intrusives - are to the north-northwest below the Paradise thrust. Above it, in the dolomite of the upper plate, they are more to the northwest. The dip of the bedding, which is generally less than 30°, is dominantly to the southwest.

Igneous Activity

The close of diastrophism saw the intrusion first of several varieties of andesite dikes followed by granophyre dikes, then granodiorite. After the granodiorite and aplite, malchite, dacite, and tertiary rhyolite dikes were intruded. The igneous rocks have been described in detail by Callaghan (1935, p. 302-307) and arranged in their proper time sequence as determined from their cross-cutting relationships by Vitaliano, et al (1957). The only comment that needs to be made here is that elsewhere in the area, particularly eight miles to the north of Gabbs, the plutonic rocks include, besides the granodiorite, diorite and granite. The granodiorite is believed to be the latest phase and the granite the earliest (Vitaliano, et al, 1957). Ferguson and Muller (1949, p. 13) relate the age of the granitic intrusives to those of the Sierra Nevada and to the Nevadan orogeny.

Mentioned above are andesite dikes which antedate the granodiorite. It should be emphasized that some dikes which also traverse the magnesite deposits, are already involved in the earliest Jurassic thrusting. This indicates that they are either older or at least contemporaneous with the thrusts. These may be a phase of Dunlap volcanism mentioned earlier.

DESCRIPTION OF THE MAGNESITE OCCURRENCE AT GABBS

Physical Characteristics and Structure

The magnesite occurs interlayered with recrystallized dolomite, within, and near the top of, a 3,000 foot section of the dark gray to black, dense dolomite of the topmost member of the Luning formation which, at Gabbs, occupies the imbricated upper plate of the gently westward-dipping Paradise thrust. The principal deposits lie in two local synclinal troughs on both sides of a northerly-trending apophysis of the main granodiorite stock. The magnesite zone is about 400 feet thick, but pendants of it extend to 700 feet in the granodiorite. In texture, color and other physical characteristics the magnesite is indistinguishable from the associated crystalline dolomite. Each, however, is readily distinguished from the main mass of the dark dense dolomite with which both, separately or together, interfinger. Except where it is marbelized, the weathered surface of the dense, dark dolomite is invariably gray, whereas that of the magnesite and magnesitic dolomite is usually buff or rusty brown, and commonly it has a sugary texture. On a fresh break the texture may be fine-grained, sandy, or coarse-granular; and the color, white to very dark gray, sometimes mottled. A buff mottling usually means that dolomite is present in the magnesite and this may indicate a recrystallized fault breccia. In the magnesitized dolomite zone, all gradations of the mixture of the two carbonates, from magnesite to dolomite, exist.

Bedding in the dense, dark dolomite is readily apparent. In the magnesite it is more obscure but discernible enough to aid in unravelling the structure, and to leave no doubt as to the bedded nature of the deposit. Rarely, a thin layer of black hornfelsic shale may also be found interbedded with the magnesite.

On the weathered surface, the color changes across bedding contacts, which are usually, although not always, conformable, as well as fault contacts, (of several generations), of the magnesitized zone with the main mass of dense, dark dolomite, are rarely gradational.

The transition from the dark, gray-weathering dolomite to the buff-weathering magnesitized dolomite is accomplished within a few inches. Although the color change across both bedding and fault contacts is abrupt, across the bedding contact the chemical as well as texture transition may be more gradational. Generally there is an increase in the silica content through an interfingering contact zone which may be manifested by the presence of tremolite on both sides of the color boundary. This is particularly true if the change in the upward direction is from dolomite to magnesite. This is illustrated in the series of section of diamond drill holes (Plate III) on which the lime content is represented by bars extending to the right from the diamond drill hole and the silica content, or rather, the acid insoluble content (which is essentially silica), is plotted to the left. The open bars mean that the silica content is in excess of 30% and usually represent dikes. Sections 8 and 10 in particular show an oscillating high silica transition zone of about 50 feet near the bottom of the holes through part of which, however, the two carbonates may interfinger. The top of the dark, dense dolomite is marked by a line. Along the early fault contacts this higher silica zone is generally absent.

Weathering will sometimes bring out the outlines of fragmental material of the dark, dense dolomite in a matrix of buff dolomite sand along the bedding contacts. In some areas, particularly those in which the two carbonates interfinger or where the bedding contact steepens, this sort of transitional conglomerate may range through a score of feet or more. Both bedding and fault contacts are, of course, completely recrystallized so that the features discussed above are apparent only on weathered surfaces. Generally, however, the contact zone is also marked by a much coarser crystallinity in the buff-weathering carbonate so that this often gives a hint of its presence even on a fresh quarry face. It would appear that during recrystallization the greater space available along contacts permitted coarser crystal growth. The interpenetrating veinlets and reciprocal replacement features, tending to eliminate the boundary, which one would expect to find at the contacts of any dissimilar carbonates under like conditions, may also have been developed during this recrystallization.

Several generations of fault contacts have been mentioned. It should be added that the earliest are probably contemporaneous with the bedding contacts. It may be recalled that the crustal instability which was characteristic of Luning time extended through the deposition of the upper dark, dense Luning dolomite. Evidence for such instability is also found in the magnesite deposits. The most striking is exposed in the present quarry. Locally down-warped magnesite beds are overlain unconformably by a two-foot bed of dark, dense dolomite which is unaffected by the warping. Overlying the dolomite conformably is more magnesite. That small scale normal faulting, at places involving only a few beds, has occurred during the deposition of the magnesite is shown most clearly near the eastern boundry of the deposit. Here the contact with the dark, dense dolomite gives the impression of a giant stairway descending toward the west. Steep fault contacts alternate with flat bedding contacts, which on opposite sides of the fault plane may show a slight divergence in dip. The faults may, at times, terminate against later transgressing magnesite beds. In contrast to the bedding contacts, these early faults are rarely siliceous. Within the magnesite they are completely obscured but may be marked by a zone of slightly coarser crystallization. Their general trend is toward the north-northeast but locally they may have any direction. These faults antedate all other structures except the bedding.

Faults of a second, later set, trending northeasterly and also recrystallized, are more readily traceable on the surface within both the dolomite and magnesite by the weathered zone of brownish discoloration and ghosts of the original breccia. On a fresh quarry face they are obscure but may sometimes be distinguished by a zone of mottling and heterogeneous sizes of the crystal . The upper parts of diamond drill holes 602 and 612 in sections 10 and 12 (Plate III) pass through one of them. Again we note the dolomitization along the fault and the absence of silicification. This holds true in general within the magnesite areas that zones of stress are usually dolomitic (Martin, 1956, p. 1774).

Reverse faults of a third, flatter, west-dipping set, of about the same age as the second (both are probably related to the period of early thrusting) follow the northwesterly regional trend of the area and are often the channelways for the granophyre and other intrusives. These are characterized by wider and more obvious breccia zones which, where they have not been altered by the intrusives, are also recrystallized.

Also following the regional northwesterly trend, zones of dolomitization were encountered in the course of mining operations for which at first no adequate explanation could be found. Subsequently more detailed mapping disclosed, however, that these usually coincided with zones of local flexure which would, at places along their trend, become faults.

Three sets of Tertiary and later structures add to the complexity of the deposits. The most prominent of these are faults which strike almost easterly and dip steeply to the south. Where they cut the various dikes, they are characterized by a siliceous gouge, silicified carbonate breccia, and local hydrothermal alteration of the wall rock. Along one of them, the Tungsten fault, scheelite has been deposited. The extent of the movement along them has been minor and mostly horizontal. Associated with the faults are local bedding slips which may die out within a short distance of the main fault.

Doloarenite Dikes

The last is a system of steep open fractures and shear zones parallel to the east-northeasterly trending set of dominant joints. These are often filled with pulverulent secondary carbonates - aragonite for instance - and clay-like mixtures of hydrous magnesium carbonates and silicates. At the intersections of dikes and faults, where the shearing has been accentuated and brecciation has been intensified, subterranian percolation of meteoric waters may develop solution breccias. These sometimes collapse. The increased and swifter flow of water through the collapsed breccia then begins to scour and erode it from the top and to fill the interstitial spaces with carbonate sand, usually dolomite, not only within the breccia itself, but along the underground channel beyond it. If the clastic burden of the underground stream becomes too great, deposition will take place. The lighter stream load permits an increase in velocity and the stream may now begin to scour, not the smooth bed, but the uneven roof and sides, so that as it aggrades the floor, the stream degrades the roof of the channel at the same time. If this continues for any length of time, an underground "dike" of clastic dolomite is built up retaining all the features of its origin, such as cross-bedding, graded bedding - even ripple marks. Further, as headward erosion of the roof continues and the channel fills, the previously deposited sands will become cemented, reducing the area through which percolation can take place, thus increasing the stream flow and, at the same time the load, at the headward end. This accelerates the filling until the channel is closed. Several collapsed solution breccias and clastic dolomite dikes have been exposed in the quarries at Gabbs. The iron-coated solution breccia and interstitial doloarenite is not, as yet, recrystallized. Suppose, however, that it were - not only recrystallized but metamorphosed - so that the sedimentary structures were destroyed. The result would be a dike of coarsely crystalline dolomite showing veinlets of coarse dolomite penetrating the walls of the enclosing magnesite and replacing them. This could very well be interpreted as hydrothermal replacement and the dike as occupying the channelway of the replacing solution.

<u>Alteration</u>

The dolomitization along faults and flexure zones within the magnesite has been briefly mentioned. Perhaps "dolomitization" is not the best word in this instance. The process does not consist of the addition of calcite (or lime in some other form) to magnesite in order to produce dolomite. Rather, the more readily soluble magnesite is abstracted, under some stress, from a mixture of the two carbonates which enriches the residue in dolomite.

At the boundaries of dikes, minor silicification of the magnesite has taken place. The dikes themselves are but little affected except where they have been cut by the later faults and acted upon by hydrothermal solutions. There the principal change in the dike has been kaolinization and assimilation of magnesia to form leuchtenburgite and other complex hydrous magnesium-aluminum-silicates. Kerr and Callaghan (1935) list a dozen or more minerals in connection with the scheelite deposit, along the Tungsten fault.

Grade of the Magnesite

The graphic chemical logs (Plate III) and the chemical contour map (Plate IV) show better than any column of analyses the variation in the grade of the ore. The magnesite is contaminated principally by silica and lime, occasionally alumina, and rarely, phosphorus. Except in the vicinity of dikes, the proportion of combined alumina, titania, and iron oxide in the magnesite rarely exceeds 2%. Generally, it is below 1% and in the dolomite it is even lower. The iron accounts for about 50% of the total R_2O_3 content. Phosphorus rises above tolerable limits only occasionally. Except for the structurally disturbed tremolitic and dike areas, the silica content in the better parts of the deposits is below 2% for the most part. Understandably, the lime content, chiefly in the form of dolomite, varies most widely. Although it may be as low as 0.5%, it is commonly greater than 1.0%, even in the best parts of the deposit.

The Chemical Contour Map

The various structures discussed previously and the variations in the grade of the magnesite are made apparent on the chemical contour map (Plate IV in back pocket). The construction of this map and its use in connection with the mining operations at Gabbs, Nevada, have been discussed in detail elsewhere (Martin and Willard, 1957, p. 426). Briefly, the chemical analyses of the drill cuttings from the quarry blast holes are plotted on a map of the bench being drilled, and chemical contours are drawn separating the orebody into the various grade zones to be mined. The plot is in terms of two variables. The CaO content which is shown in variously colored bands going up the color spectrum with increasing lime is the most important. The red areas indicate the lowest lime content (less than 1.5%) and the purple areas the highest (greater than 20%). Superposed upon this are green contours outlining the various zones lying between certain limits of the silica or acid insoluble (insol) content. The various ore grade boundaries are staked out on the bench after a blast, and the material is mined accordingly.

The map covers a small area in the southeast corner of Plate 1, just west of the two dikes and astride the fault, (the Tungsten fault), shown there. The easterly-trending dark gray band traversing the center of the contour map is the high insol zone representing the Tungsten fault. The northwesterly-trending gray colored zones in the upper right and lower right corners of the colored area are dikes. The purple area in the upper right-hand corner west of the intersection of the dike with the Tungsten fault represents a collapsed solution breccia, and the high lime band extending from it to the west, the filled channel of the underground stream. The latter joins a zone of flexure represented by the area of high lime to the northwest and other areas on the same trend to the southeast. The three high lime belts striking about N, 70°E, represent the recrystallized and dolomitized fault zones discussed previously which cut through the flexure zone mentioned before and are approaching another at the western edge of the map. The high insol area (stippled) below the southernmost of these faults on the map means either that the drills have encountered a tremolite layer or a bedding fault - in this instance, the latter.

THE BRUCITE DEPOSITS

If magnesite deposits are formed by hydrothermal processes, then the extensive transformations that have occurred, wherever masses of magnesite and dolomite have been engulfed in the granodiorite and largely converted to brucite, might shed some light on them. The massive brucite deposits at Gabbs, estimated to contain over 3,000,000 tons of brucitic material (Callaghan, 1933) occur in two pendants in granodiorite (lying on either side of the apophysis mentioned earlier) in which most of the magnesite and some of the dolomite has been converted to brucite by the action of the intrusive. The larger of the two deposits is about 1,500 feet long and 500 feet wide at the surface, and 700 feet deep. Diamond drilling indicates that at no great depth the brucite is completely surrounded by the granodiorite which apparently acted as a huge autoclave and also supplied the heat and the fluids to bring about the transformation. The conversion, however, was far from complete and large blocks of dolomite, as well as magnesite, remain unaltered. The brucite bodies are traversed by the various dikes mentioned previously and also by tongues of granodiorite all of which have been, in part, coverted to serpentine. Besides forsterite (Mg_2SiO_4), serpentine ($Mg_3SiO_2O_5(OH)_4$), and rarely periclase (MgO), there are developed a host of complex magnesium silicate minerals of the sepiolite-deweylite type ($3MgSiO_3 \cdot Mg(OH)_2 \cdot 5H_2O$) as well as veins of cross-cutting, coarse, white dolomite, and bone magnesite. Weathering has converted the brucite to hydromagnesite sometimes to a depth of 50 feet or more. The bone magnesite may also, in large part, be derived from further carbonation of the hydromagnesite and downward migration as $Mg(HCO_3)_2$ and redeposition as $MgCO_3$, under some hydrostatic head, along the many openings, particularly in the zone of brecciation along the contact with the granodiorite.

Formation of the Brucite

The transformation of the carbonate to brucite is usually thought to involve two steps: the calcination of the magnesite to periclase, and the subsequent hydration of the magnesium oxide to brucite. However, it is doubtful that at Gabbs this took place. The mineral assemblage, particularly the presence of serpentine and forsterite with periclase would indicate that the temperature for the large scale decomposition of magnesite into periclase and carbon-dioxide was not attained except perhaps quite locally (Bowen 1940), (Bowen and Tuttle, 1949). Weeks (1956 a, p. 258) has calculated the pressure-temperature equilibrium curves from thermodynamic data for the reactions:

a) $MgCO_3 + SiO_2$ -----> $MgSiO_3 + CO_2$

b) $2\ MgCO_3 + SiO_2$ -----> $Mg_2SiO_4 + 2CO_2$

c) $MgCO_3$ -----> $MgO + CO_2$

He shows that at the same pressure, reactions (a) and (b) will take place at substantially lower temperatures than reaction (c). For instance, at one atmosphere pressure reactions (a) and (b) take place at about 250 and 300° C. respectively. At the same pressure reaction (c) will occur at 410°. The same relationship holds at higher pressures. Between (b) and (c) still lie the equilibria for diopside and wollastonite. However, reaction (c) is not necessary for the formation of brucite. If we initiate the release of a small amount of CO_2 at the contact of the intrusive through reactions (a) or (b), and if the CO_2 is prevented from escaping, then in the presence of water, the following reaction will take place:

d) $MgCO_3 + H_2O + CO_2$ -----→ $Mg(HCO_3)_2$

This puts the magnesia into solution. The release of CO_2 or dilution of the solution with the attendant rise in pH will precipitate $Mg(OH)_2$ rather than $MgCO_3$ according to the reaction:

e) $Mg(HCO_3)_2$ -----→ $Mg(OH)_2 + 2CO_2$

Precipitation of the $Mg(OH)_2$ removes some water, increases the concentration of the CO_2 again, and the process may be repeated. In this manner a large body of carbonate may be changed with a relatively small amount of CO_2 in solution at any one time. Thus, the reaction pressure and therefore the temperature are kept to a minimum. This is probably the process that formed the brucitic marbles at Wakefield, Quebec (Goudge, 1940), in the Organ Mountains of New Mexico (Dunham, 1935), and elsewhere. All occur within pendants of dolomite in granitic rocks. The brucite occurs as small granules and rosettes in a matrix of calcite. The transformation is usually postulated as follows:

f) $CaMg(CO_3)_2$ + Heat -----→ $MgO + CaCO_3 + CO_2$

g) $MgO + H_2O$ -----→ $Mg(OH)_2$

The presence of unreacted periclase granules surrounded by brucite is cited as evidence for these reactions. The calculated dissociation temperature for reaction (f) at one atmosphere CO_2 pressure is 506° C (Weeks, 1956a, p. 249). Actually it takes place between 700° and 800° C depending upon the iron content, and as low as 600° C if the CO_2 pressure is low enough. However, we need not climb this periclase temperature ridge. Consider the following reaction given by Weeks (1956a, p. 259):

h) $CaMg(CO_3)_2 + 2SiO_2 \text{------}\rightarrow CaMg(SiO_3)_2 + 2CO_2$

(A similar equation could be written for the formation of enstatite plus calcite rather than diopside). The calculated temperature at one atmosphere pressure for reaction (h) is about 300° C. Again, the liberated CO_2, in the presence of water, puts the dolomite into solution as the bicarbonates of lime and magnesia from which brucite, as well as calcite, may be precipitated as before.

If the transition through periclase is not necessary for the formation of brucite, how, then, are we to account for the periclase, which has been reported in all of the deposits? Consider the reverse of the reaction (g) above. Again at one atmosphere pressure the calculated temperature (as well as that experimentally determined) at which calcination of brucite to periclase takes place is about 290°. (Weeks, 1956b, p. 470). It can be seen that as the temperature continues to rise the first formed brucite will be overtaken, and some of it will be decomposed without necessarily affecting the previously formed serpentine and forsterite. As the temperature drops again, the periclase will be rehydrated and this, at times, incompletely.

Little has been published on the system $MgO - H_2O - CO_2 - SiO_2$ so the various phase boundaries are not as yet well defined in the range of temperatures being discussed. However, the T - P curve for reaction (g) always lies below those for reactions (a) and (b) even at higher pressures so that the reasoning above would still be valid. Kazakov, A. V., et al (1957) have investigated the system $MgO - H_2O - CO_2$ at the 20°, 60°, and 150° C isotherms, and at CO_2 concentrations from 0 to 19,000 mg./l. At 20° C and concentrations up to 270 mg./l. CO_2 brucite is stable. Intermediate temperatures and higher concentrations of CO_2 up to 6610 mg./l. gave mainly basic hydrous magnesium carbonates. At 150° C and concentrations of CO_2 of 843-2528 mg./l., 85% of the solids formed were magnesite. Therefore, low concentrations of CO_2 are necessary at all temperatures for the formation of brucite. These conditions are met in a partly open system in which the CO_2 may escape but where the water is retained and enters into the composition of the products.

PYROXENITE AND SILICATED DOLOMITE

The discussion of pyroxenite and silicated dolomite would appear to take us somewhat beyond the boundary of our present problem which is, after all, the origin of crystalline magnesite. However, some features of these deposits will, perhaps, be better explained, after some examination of these rocks and the processes through which they originate.

The widespread regional metamorphism of the Lunining sediments in the Gabbs area has been mentioned by others (Vitaliano, Callaghan, et al, 1957). The development of bands of tremolite and widely dispersed talc in the dark, dense dolomite of the Luning formation has also been alluded to earlier in this discussion. There is always the temptation to attribute these effects to the influence of granitic intrusions. That the plutonic rocks did bring about changes in the intruded sediments there can be little doubt. The tactite zones developed along their contacts with carbonate rocks, are of course, familiar to everyone. And to a very limited, almost trivial, extent these effects are also evident at Gabbs. What needs further examination is the regional silication involving dolomite masses comparable in volume to the intrusives themselves. The belt of silication along the western foothills of the Paradise Range is roughly eight miles long, if its interruption by the intrusives and the magnesite deposits is not excluded, and up to two miles wide. The general trend is to the east of north - almost at right angles to the trend of the intrusives. To the south of Gabbs, the belt diverges also from the trend of the Paradise thrust. (Plate II). In this area, large masses of dolomite have been, more or less, converted to a diopside rock apparently by the simple exchange of silica for carbon dioxide. An extreme example of this transformation is to be found in the western part of the magnesite area directly south of the plant site (Plate I). Here, a layer of diopside rock, 250 feet thick, overrides the magnesite in the Margie Pit along a local thrust. Although still retaining all of the outward appearance and structure of the dolomite, the rock's chemical composition is essentially that of diopside. The following is a typical analysis of a portion of diamond drill core which has been recalculated omitting the loss on ignition and R_2O_3 content, and compared with the theoretical composition of diopside ($CaMg(SiO_3)_2$).

	Loss %	R_2O_3 %	SiO_2 %	CaO %	MgO %
D. D. Core	1.39	5.85	49.82	25.19	17.55
Recalculated			53.83	27.21	18.96
Diopside (Theoretical)			55.48	25.90	18.62

The underlying magnesite remains completely unaffected by the drastic changes which took place in the sediments presently lying directly above it. Westward, away from the contact with the magnesite, that is, upward in the section, this rock grades into silicated cherty dolomite, then cherty dolomite and siliceous dolomite in which the silica content is less than 10%.

Another example of a similar occurrence lies somewhat east of the magnesite deposits and beyond the southeast corner of the area shown on Plate I. Here several layers or beds of rock, two to six feet thick, of diopside composition are interbedded with hornfelsic shale and the dark, dense dolomite previously discussed. Ferguson (1924, p. 20) also mentions a layer of diopside at the top of the Cambrian Gold Hill formation in the Manhattan district, Nye County, Nevada.

What is the process which introduces such large volumes of silica into these masses of rock, yet it is so discriminating that a brecciated structural boundary or even a sedimentary break will confine it, apparently absolutely, to the individual sedimentary horizon upon which it operates? Furthermore, the silica is doled out so nicely that there is usually less and rarely ever an excess (to remain behind as quartz veins for instance) over that required to convert the dolomite into a proper silicate. The writer has yet to see a quartz vein of any consequence in the silicated area near Gabbs.

The relationships in the Margie Pit area would indicate (if one is inclined to be a little skeptical of the sensitivity of the process) that either the silication took place prior to the thrusting which brought the silicated rock over the magnesite, or perhaps it was brought about early in the course of the thrusting before it had arrived at its present location. In either event the granodiorite could not have mothered the silication as the thrusting has been shown to be much earlier than the intrusion. If the thrusting was instrumental in bringing about the silication, we still have the problem of the source of the silica. In the second occurrence, mentioned above, however, thrusting is not a proximate complication. Except for the regional folding, this locality is undisturbed. The main Paradise thrust zone lies a half mile or so farther to the west or at least a thousand feet down. Conceivably it could have something to do with the silication. However, the Paradise thrust is not silicated to any greater extent than thrusts in similar sediments elsewhere. The problem of the source of the silica and the method by which it was introduced remain. For a possible clue to the solution of the problem we go back to the Middle Triassic and the Excelsior formation.

It will be recalled that at most places where the base of the Luning formation is exposed it consists mostly of a conglomerate derived from the chert, metavolcanics, and hornfels of the underlying, folded Excelsior formation, which fringes the basin of Luning deposition. Recurring conglomerates of the same material are also found higher in the section, particularly throughout the middle Luning which in the Pilot Mountains consists in large part of conglomerates. (Muller and Ferguson 1939, p. 1596). Ferguson and Muller (1949, p. 9) further suggest that there may have been an island of Excelsior within the embayment between Mina and Luning that was being eroded through at least a part of the Luning time.

It is evident from the foregoing that the restricted Luning Embayment was receiving large volumes of highly siliceous detrital material during a major portion of Luning sedimentation. This silica was initially of clastic volcanic origin which means that a substantial part would be in the form of more readily soluble silica glass (Ramberg, 1952, p. 232). The buildup of the silica content under the conditions outlined above and its periodic incorporation in the sediments in some form would seem to follow as a matter of course. The cherty dolomite in the uppermost Luning indicates that at least locally the sea was saturated with silica at that time.

Supposing, then, we had such sediments consisting of a carbonate-silica mixture, or carbonate gel (Vital, 1951). It has been pointed out that recrystallization of the carbonates has taken place on an extensive scale in the Gabbs area. The processes, largely dynamic, which brought this about also operated, of course, on the high silica horizons tending also to reconstitute the silica and carbonates of the mixture. In general, there is a greater chance of interaction between mixtures of solid phases undergoing morphic transformation than there would be between the same phases under equal chemical and physical stress but not involving a change of form. A common example is the incorporation of manganese by quartz in the transition from the alpha to the beta form and the reverse. The reason for this is that in the transformation the solids go through a disordered state (Elcock, 1956) in which they are, as it were, more susceptible to subversion. Using again, only for the sake of consistency and for comparison, the pressure of one atmosphere, the reaction (h) $CaMg(CO_3)_2 + 2SiO_2 \text{------}\rangle\ CaMg(SiO_3)_2 + 2CO_2$ according to Weeks, will take place at about 300° C. Iron replacing some of the Mg, and the presence of other fluids will lower the temperature somewhat. Again, lacking knowledge on a silicate system containing both high H_2O and CO_2 the extent of this lowering cannot, as yet, be determined.

These fluids now become tremendously important. It can be inferred that, in the presence perhaps of fluids of a particular kind, lowering of the reaction temperature and pressure to atmospheric is likely. It should be born in mind that hydrated silicates are being formed during weathering. Rogers, Quirk, and Norris (1956) describe the occurrence of an aluminum-sepiolite in soils. Vital (1951) mentions recovering a gray, viscous, colloidal-appearing substance being formed in Lake El'ton which consisted partly of brucite and hydromagnesite and also in part of sepiolitic silica gel. The particular fluid referred to above may be a gel.

There is still the question of when the silication of the dolomite to pyroxenite at Gabbs took place. Was the silica originally incorporated in the sediments in the form of a hydrated autosilicated compound and the silication only made evident by subsequent metamorphism, or was it deposited with the dolomite as a silica gel and was the silication of the mixture then actually brought about by the metamorphism? Again, how is the general absence of quartz in the silicated areas, not only at Gabbs but also in the Organ Mountains of New Mexico described by Dunham (1935), related to one or the other of the above alternatives? Whatever the answers to these questions, it is evident that the pyroxenite at Gabbs is as much of sedimentary origin as a slate would be. The autosilication of the dolomite was either brought about or made evident by the stress, probably, of the second phase of thrusting in the Gabbs area which overrode the upper plate of the earlier Paradise thrust.

There are implications of the foregoing worth mentioning here. We have on the one hand extensive deposits of dolomite of all ages, on the other, there are masses of chert and other highly siliceous sediments. Interbedded cherts and dolomites are not rare. Crystalline magnesite deposits, considering the amount of magnesia available in sea water, occur rather infrequently. But as far as the writer is aware, there are no significant deposits of interbedded chert and magnesite except of the most trivial sort to be mentioned later. There are, however, masses of chert "intruded" by equally large masses of serpentine. Could the process of autosilication outlined above to account for the diopside at Gabbs also explain the other anomolies? Reactions (a) and (b) discussed in connection with the brucite deposits would seem to point in that direction. The writer has also encountered an impressive body of rather specific field evidence to support it. One instance might be instructive to recount. In drilling for chromite in the ultra-basic omplex of the Camaguey area in Cuba, a substantial body of partly silicated crystalline magnesite was encountered in the serpentine. It is possible that autosilication of high magnesia sediments might account, in part, for the origin of some ultra-basic rocks.

The simple system, $H_2O-CO_2-SO_2$ in the form of a colloidal silica gel, may yet turn out to be one of the most important in nature. Modified by combination with other ions, it appears strongly to determine some properties of soils. It may also play a significant role in the lithification and metamorphosis of sediments and, one suspects, their transformation into "igneous" rocks. Its ubiquitous influence has remained obscure because the changes which it facilitates in the rocks also destroy its identity. However, reconstituted, it finds expression as quartz.

THE ORIGIN OF THE MAGNESITE DEPOSITS AT GABBS

The Role of Thrusting

It should be clear from the foregoing discussion that a genetic link between the intrusives and the magnesite deposits is untenable. Could the magnesite deposits have been formed during the thrusting?

The remnant of the trust sheet, in the upper plate of the Paradise thrust lying directly south of Downeyville, (Plate II), along which the Excelsior formation overrides the middle and upper Luning, must at one time have extended over a much larger area, possibly over the present site of the magnesite deposits and beyond, conceivably joining the South thrust (Ferguson and Muller 1949, p. 40) cropping out just south of the map area. This thrusting could have been responsible for changing the siliceous dolomite into a silicated rock in the manner previously outlined. The CO_2 released by this process would, in the presence of H_2O, attack the dolomite and, at the temperature and pressures prevailing, would dissolve the magnesia preferentially as suggested by Faust (1949, p. 789-823) and by Faust and Callaghan (1948, p. 59), in their account of the Currant Creek deposit at Currant, Nevada. Leaving behind the calcite, the magnesium-rich solutions would reprecipitate their load as the pressure dropped and CO_2 was lost. If this process actually took place one would expect to find, in addition to the larger masses, fringing networks of dolomite and magnesite veins, serpentinization of previously existing dikes, some control of the magnesite deposition by previously existing structures, and remnants of the calcite that would be left behind or precipitated out to produce the magnesia-rich solution. None of these, of course, are in evidence in the magnesite deposits at Gabbs. Except for the effect of the bedding, the lack of structural control is one of their outstanding characteristics. Even in the brucite pit areas, where one would expect that the conditions outlined above would more likely have been fulfilled, only a trivial amount of bone magnesite, if any, was deposited in this fashion.

The brucite deposits near Oatman, Arizona, (Basic, Inc. private report) occur in veins in tertiary volcanics. Again the conditions would appear to have been ideal for the deposition of magnesite. Yet it is little in evidence. The brucitic dolomite at Rutherglen, Ontario, (Keith, 1946, p. 967-984), Wakefield, Ontario (Doudge, 1940, p. 481-505) or in the Organ Mountains of New Mexico (Dunham, 1935, p. 26-160) rarely contain magnesite. Wherever magnesia was deposited under undoubted hydrothermal conditions, the form appears invariably to have been brucite. The explanation may be that hydrothermal solutions refuse to stand still to be cooled before moving up to horizons of lower pressure so that an excessive loss of CO_2 and rise of pH occurs before the temperature has dropped sufficiently to retain the CO_2 at the lower pressure and for magnesite deposition to take place. Usually, therefore, brucite will be precipitated. How, then, do we account for the origin of the magnesite deposits at Currant Creek, the showpiece for hydrothermal replacement, or for that matter, the origin of magnesite in serpentine? More will be said about the Currant Creek deposit in due course.

The Depositional Environment

It is not impossible, but highly unlikely, that the magnesite at Gabbs was deposited under hydrothermal conditions. However, let us see if a different process will not account more adequately for its origin. It has been brought out earlier in this discussion that the magnesite was involved in most if not all of the events that have affected the Upper Luning sediments as a whole. The earliest of these were local warping and small-scale contemporaneous faulting. That the magnesite emplacement antedates all structures is further emphasized by the lack of evidence for any structural control of the deposition. The persistance of discordant bedding and recrystallized dolomite sands and conglomerates at the contacts of the high magnesia rocks with the underlying dark, dense dolomite all point to a sedimentary environment for the emplacement of the magnesite. If this is so, then how and when did the deposition take place?

Let us recall for the moment the two facies of the Upper Luning. Beneath the thrust it consists dominantly of black, thin-bedded limestone with minor interbedded argillite. The rocks in the upper plate, on the other hand, are made up for the most part of the dark, sense dolomite, in which the magnesite is found. According to Ferguson and Muller (1949, p. 5), the carbonaceous thin-bedded limestone and argillite represent the near-shore facies of which the dolomite is the essentially contemporaneous offshore equivalent brought to its present position by the Paradise thrust. Both were deposited in the same shallow basin receiving, at the same time in their respective areas, only the finer clastic and carbonate sediments. The black color of the thin-bedded limestone and also that of the dolomite is, no doubt, due to occluded carbon which testifies to considerable organic activity during the time these sediments were laid down. The magnesite deposition took place, then, in the same offshore depositional environment as that of the surrounding dolomite. However, dolomite sands and conglomerates often found at the base and along the boundaries of the magnesite would indicate near-shore conditions, and, what is more, near a land mass of considerable relief. On the other hand, the low silica content of the magnesite (except as this was subsequently introduced) and the exceedingly rare and very thin hornfelsic argillite horizons would indicate that the deposition took place far enough from the shore to be essentially beyond the reach of siliceous clastic contamination. Unless we postulate a sudden uplift at this time which is not born out elsewhere in the sediments, how are these facts to be reconciled?

The structural instability of the Luning embayment throughout Luning time has been mentioned before. Eight thousand feet or more of Luning sediments have now accumulated. The sinking has slackened. The basin is filling. Conglomerates, even around the margins, have given way to finer sediments, and the carbonates have spread over a greater area. Life is abundant in the warm shallow sea. Considering the comparatively small size of the embayment, perhaps some adjustment is due. This adjustment has, of course, been going on all along, but it was apparent in the sediments mostly only near the bordering lands. The sea is shallower now, and even a small amount of warping within the basin brings the bottom locally close to the surface perhaps raising an occasional island above it. Isolated domes line up into ridges, and as they rise within the limits of wave action, the tops will be broken and scoured, and the material deposited in the intervening troughs. As warping continues and the sides of the ridges steepen, the still plastic carbonates tend to slide into the troughs folding and breaking up in the process. Now and again a part of a ridge will be faulted forming a graben, or the sediments in the troughs will be faulted due to compaction. A buried fault ridge may remain between the trough and the graben - steep on the side of the graben, sloping more gently on the side towards the trough - over which later beds may be draped apparently assymetrically folded. If the floor of the trough or the graben drops low enough, the sediments it receives may consist for a time of the same kind being deposited in the larger basin so that there would be ridges and tongues of one kind of dolomite interfingering with and interpenetrating a second kind - that which was being reworked. As the whole area rises or the sea falls, the material in the troughs may be reworked a second time. This continual plowing and hoeing, together with the pulsing of the sea and the changing cycles of life, will tend at times to dissolve the dolomite, at others to precipitate magnesite or possibly gypsum* according to whichever process the physical, chemical and ecological conditions may favor.

That such semi-emergent conditions, or possibly even complete local emergence, may have existed in the Luning Embayment at the end of Luning time is suggested by the sedimentary break and radical change in facies between the upper Luning dolomite which was followed, abruptly and without gradation, by the carbonaceous siltstone of the lower Gabbs formation. The carbon in the siltstone suggests organic activity, probably in shallow water. Although no fossils have been found in either the magnesite or dolomite, the

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* There is a small gypsum deposit in upper Luning limestone in the lower plate of the Paradise thrust four miles north of the magnesite deposits at Gabbs.

occluded hydrogen sulfide which is liberated from the magnesite when it is being drilled or crushed is readily detectable, particularly when several drills are going at the same time. A rather rough, but hardly ever-failing, test of the pit foreman for the grade of the ore at Gabbs is, "It must be good, it smells bad." Occasionally the pit samples also contain a per cent or so of free,graphitic carbon.

When the elevations of the boundary between the dark, dense dolomite and the magnesitic dolomite, marked on the diamond drill holes, as shown, for instance, on sections 8 and 10, are plotted and contoured, the result shows up as troughs and ridges whose trend does not necessarily coincide with the trend of the local folding as mapped at the surface. Rather, the troughs are aligned more with the regional trend of the sediments in the lower plate of the Paradise thrust. This, however, may be fortuitous as not enough of the subsurface structure of the magnesite area has been studied in detail.

Dolomite and a variety of magnesium carbonates are being deposited at the present time in several embayed bodies of salt water and saline and fresh water lakes. A most instructive example alluded to before is described by D. A. Vital (1951). He tells of recovering a gray, viscous substance of colloidal appearance floating near the shore of a saline lake (Lake El'ton). One part of this mass was not definitely identified other than that it displayed aggregate polarization with index N of 1.507 to 1.512. This was thought to be a sepiolitic silica gel. The other part of the mass upon analysis consisted of a plastic mixture of brucite, hydromagnesite and other hydrous basic magnesium carbonates, and contained 8.4% carbon.

Kazakov, et al (1957) cite the deposition of magnesite in muds of Kara-Bogaz, an easterly embayment of the Caspian sea, and dolomite in the Kaidak bays of the Caspian Sea, and in the muds of Balkash Lake.

In Australia, Alderman and Skinner (1957, p. 561-567) describe the deposition of dolomitic sediments in saline lakes and in a shallow inlet of the sea 10 miles north of Kingston near the southeastern tip of the province of South Australia. Here, a dunal area, recently vacated by the sea, is covered with Pleistocene and Recent marine and fresh water sediments - mostly carbonates - underlain by partly dolomitic limestones of Tertiary age. Fine-grained plastic sediments consisting of dolomite, calcite with some halite and a little quartz and clay are deposited in ephemeral lakes - mostly during the spring and summer rains - from waters whose salinity may at times be half that of sea water. In some areas gypsum is being deposited, and in others there are carbonate beds 2 to 4 feet thick mostly of fine-grained dolomite. These lie above shelly beds and dark, sulfur-bearing muds. Sedimentation in the lakes is most rapid when plant growth is most vigorous. During these times the pH of the water rises from a normal 8.2 to 9.3. Besides the pH and the concentration of magnesia and carbon dioxide, the presence of other ions, particularly SO_4 and Cl, apparently also affect the precipitation of magnesium carbonate. What subtle control organic activity may exercise in addition to being a pH regulator and SO_4 generator is difficult to assess.

The pH, of course, varies with a wide variety of factors in addition to the concentration of CO_2. The pH of the saturated solution in contact with several solids at standard conditions is tabulated below. From the Handbook of Chemistry and Physics (33rd Ed.) we have

Solid + Solution	pH
MgO	10.5
CaO	12.4
$CaCO_3$	9.4

Faust (1949, p. 812) gives

$Mg(HCO_3)_2$	7.5	
$Ca(HCO_3)_2$	6.35	
CO_2	5.0	
$MgCO_3$	9.5	(determined at Gabbs)

Normally then no magnesium carbonate would be formed at a pH above 10.5. Brucite would be the expected product. Between 10.5 and 9.5 the hydrous basic carbonates would be formed, and below that the hydrous carbonate would precipitate. Magnesite does not form even at the pH of 9.5 except at somewhat higher temperatures or at elevated pressures of CO_2. In general, as the CO_2 concentration decreases, the carbonates formed become more basic. A decrease in the concentration of CO_2 can be brought about, of course, either by a rise in the temperature at the same pressure or a drop in the pressure at the same temperature. Kazakov, et al, (1957, Table 20) produced a mixture of hydromagnesite and magnesite at 60°C from a solution containing 200 mg./l. MgO, and 380 mg./l. CO_2 at pH of 9.43. This would seem to be the lower limit of $MgCO_3$ precipitation at that temperature. Neither dolomite nor magnesite have been precipitated in the laboratory under normal conditions. However, Rogers, et al, (1956, p. 177-184) describe nodules of dolomite in an alunimous-dolomite-sepiolite marl below surface soil of aluminous-sepiolite, montmorillonite, illite, quartz, and dolomite. The writer likewise has found nodules of a bone-like carbonate in small open caverns in the magnesite deposits at Gabbs which by chemical and X-ray analysis proved to be dolomite. Nodules of bone magnesite are common, of course, in the soils of serpentine areas. Nature does, then, produce dolomite and magnesite under surface conditions although man has yet to duplicate the process in the laboratory. How, then, is magnesite or, for that matter, dolomite formed in the kind of shallow basin postulated as the site of origin?

The hydrous carbonates are converted to the anhydrous forms with a moderate increase in pressure (Faust, 1949, p. 810). The decay of organic material would supply the carbon dioxide and a rise in the sea level, or simple sedimentary loading would increase the pressure. Again the colloidal carbonate might also be carried to deeper water and greater CO_2 concentrations by local currents. All of the above processes are probably involved at one stage or another.

Geodes and Magnesite Nodules

Another, somewhat different, process may also play a significant part. Consider a drop of water. It retains its spherical shape because of the force of surface tension. This force is the product of the tension along a unit length times the total length, in this case, the circumference. This is balanced by the force of the liquid within the drop which is a product of the pressure (that is, force per unit of area) times the area, in this case the area of a plane passing through the diameter of the drop. If T is unit surface tension (dynes per centimeter), and P is the pressure, d, the diameter of the drop, and k, a constant indicating proportionality, then from the foregoing

$$kdT = d^2P, \text{ or } P = \frac{kT}{d}$$

As the size of the drop changes, T remains constant. P therefore varies with the diameter of the drop below a certain maximum diameter for a particular temperature. If the drop is made very small, the pressure becomes very great. There is no need to add the actual computations. The point is, the pressure can be made as great as we like by making the drop small enough. What has been said about very small drops applies also to very thin films.

To return now to our problem. We have seen that under "normal conditions" that is, the range of pressures, temperatures, concentrations and ecological environments to be encountered in the seas, brucite and hydromagnesite are precipitated, rather than magnesite. The initial precipitate is probably colloidal which becomes gel-like and clay-like as water is expelled. These plastic masses are held together, in large part, by the tension of inter-particle films. If in this mass is developed a seed of a mineral of which the mass contains the ingredients, or if it contains such seeds of detrital origin, processes would be initiated which would tend to eliminate the local imbalance and to restore the system to homogeneity. Whether the seed was assimilated or the mass crystallized would depend upon the balance between the free energy changes and surface energy changes involved in the transformation. At any rate, in whichever direction the change, it rarely goes to completion. The best that is attained is a stable arrangement of the system with its surroundings and gradients will remain within the system as long as the surroundings are not homogeneous.

The first step is for the mass to enfilm the strange particle. At the pressure prevailing in the fluid within the film, magnesite precipitation is possible. This also liberates water, which diffuses back and has the over-all gross effect of increasing the size of the "drop" thus reducing the pressure. The steep concentration as well as pressure gradient produced across the fluid of the film by the release of water and the precipitation of MgO and CO_2 as magnesite forces a rush of new material into the film, and the process is repeated. If the process goes to completion and the system is closed, nothing should be left at the end except water and magnesite. If it is repeatedly interrupted, small nodules or oolites will result. The formation of geodes or "thundereggs" is brought about by the same process. The gelatinous silica agglomerates by diffusion through the clay soil and grows to a certain size, almost like a drop of water, which depends upon the conditions within itself and its surroundings. It begins to crystallize first around the boundary with its surroundings both inward and outward. However, if the outward conditions change, growth in this direction may stop and solution may even set in to smooth down the more readily soluble sharp edges of the previously formed crystals. However, inward the process goes on in what now amounts to a closed system until nothing is left but water and crystals. The five-pointed star often developed in geodes may result from an excess of water (the drop is too large) in the original gel which favored the precipitation of a hydrous compound (opaline silica for instance) rather than crystals. During the process of such dehydration the amorphous mass may be subjected at some stage to internal forces similar to those tending to produce pentagonal mud cracks. The agglomeration of the silica takes place in much the same fashion that oil is displaced by water from oil sands and probably depends on a property analogous to surface tension. This accounts for the mobility of the silica (Ramberg, 1952, p. 233) and for the chert nodules which are a common feature of some sediments. The process of geode formation accounts also for the growth of quartz crystals in clays and probably for a variety of other phenomena such as the formation of fluid inclusions which we ordinarily think of as occurring only in hydrothermal solutions. It explains also the nodular magnesite occurring in serpentine and the nodules of dolomite mentioned before. The formation of nodular manganese in deep sea muds, as well as the growth of nodules of barite and phosphate, may have a similar explanation. The purity of the nodular magnesite might be explained by the agglomeration of the occluded silica gel as outlined.

In the precipitation of dolomite there is an additional complication because of the formation of the double carbonate. The introduction of another step and process, previously mentioned, might open the door to an explanation of its origin. Assume that the first temporary step is the precipitation of the lime as aragonite which immediately starts its transition to calcite. As it passes through the disordered state (Thompson, 1955, p. 974) introduction of magnesia could take place, particularly if the transition occurs at above normal temperatures in a carbonate mud where the facilitating agency of organic activity and the forces of our powerful drop of water can be brought to bear. This is suggested in part by accounts of the formation of dolomitic limestone from coralline aragonitic lime muds in the Bahamas and the fact that aragonite and dolomite are never found side by side in the bore of the Funafuti atoll. (LeBlanc et al, 1957, p. 147). Little aragonite is found in the consolidated rock, in any event. The formation of dolomite from aragonite in closed lagoons is also suggested by Clark (1924, p. 568). The periodic occurrence of dolomite in the Funafuti bore could also be accounted for by ecological changes accompanying climatic changes and the oscillation of the sea level so that more aragonite was precipitated by organic activity at those times rather than that dolomite was produced because of the changes in pressure. The effect of temperature and therefore of climate upon the precipitation of aragonite in the shells of organisms is well documented by Clark (1924, p. 572). This, of course, could be equally true of dolomite occurrences in general. This explanation would do away with the hypothesis that the carbonate rocks have to be buried, the deeper and longer the better, in order to facilitate the access to them of magnesian waters which would convert them to dolomite.

From all of the foregoing, it would appear that given a source of magnesia in restricted shallow sedimentary basins, the process outlined above can account for the precipitation of magnesite as well as dolomite.

<u>Magnesite Deposited in Interior Basins</u>

Up to now we have assumed that the sea always supplied the magnesia. However, there are sedimentary deposits of magnesite which were laid down in continental basins. Of these, the best known in the United States are those in Clark County, Nevada, south of Overton, (Hewett, et al, 1936, p. 119-141) and near Needles, in San Bernadino County, California (Vitaliano, 1950, p. 357-376). The deposits at Overton consist of thin interbedded white, clay-like siliceous magnesite, dolomite, siltstone and tuffaceous beds, 155 to 325 feet thick, occurring in the Miocene (?) Horse Springs formation. The siliceous material corresponds to parasepiloite in composition ($2MgO.3SiO_3.2H_2O$) (Hewett, et al, 1936, p. 130). The California deposits are similar. In the Needles deposits, the horizon containing the magnesite is 80 feet thick and consists of dense clay-like white magnesite and dolomite with some shale. Hewett and Vitaliano agree on the continental and sedimentary nature of these deposits and both agree also with Longwell (1928, p. 85) that the magnesia was brought into the lake basins by deep-seated hot springs. Hewett, et al (1936, p. 135) speculate that the dolomite from the Paleozic rocks in the nearby Muddy Mountains might be the source of magnesia for the Overton magnesite, but dismiss this idea as unlikely because of the problem of accounting for the calcite that would be left over. Let us see if that can be done.

The problem is two-fold and the reverse of what it has been before: first, to get the dolomite back into solution, and second, to separate the lime from the magnesia, and to do this under what we have been calling "normal conditions" (see above). How Faust and Callaghan (1948, p. 11-74) and later Faust (1949, p. 789-823) derive a magnesia-rich solution has already been mentioned. Their processes require elevated temperatures and pressure. At one atmosphere CO_2 pressure, both calcite and magnesite are more soluble than dolomite, and when it does dissolve, its solution takes place congruently. Their solubilities in millimoles of bicarbonate per 1000 grams of solution under the conditions stated are: calcite, 9 millim., magnesite, 16 millim., dolomite, 3.2 millim. Under normal conditions the solubility of dolomite is negligible. (O. K. Yanat'eva, 1949, p. 479-481). Yanat'eva (1950, p. 252-268) studied the solubility of dolomite at 25° C. and one atmosphere CO_2 pressure in the presence of NaCl up to 2% and in the presence of gypsum ($CaSO_4 \cdot 2H_2O$), magnesium sulfate ($MgSO_4nH_2O$) and calcite ($CaCO_3$) and various combinations of these.

The presence of NaCl increased the solubility of dolomite from 3.2 to about 6.4 millimoles per liter at 2% NaCl, but equilibrium was delayed, with increase of NaCl, to 165 days at the highest concentration used. Gypsum tended to depress the solubility, in other words, stabilize the dolomite. Calcite also reduced the solubility of dolomite. Magnesium sulfate increased the congruent solubility of dolomite. At higher concentrations it gave (omitting H_2O)

i) $CaMg(CO_3)_2 + MgSO_4 \text{-----}\rangle\; CaSO_4 + 2MgCO_3$.

It should be remembered that under normal conditions the hydrous carbonate would be precipitated. Reaction (i) does very well as long as the $MgSO_4$ holds out, as for instance in the Gulf of Kara-Bogaz cited before, for which the reaction, again omitting water, is

j) $Ca(HCO_3)_2 + MgSO_4 \text{-----}\rangle\; MgCO_3 + CaSO_4 + CO_2$.

In the above reaction a source of SO_4 is needed in addition to the magnesia.

Taking the last first, this is not too much of a problem in an environment teeming with life. The processes of organic decomposition and oxidation which ultimately produce SO_4 are well known. In the presence of $Ca(HCO_3)_2$ gypsum is precipitated as above. This solves the part of the problem of getting rid of the lime.

Yanat'eva (1956, p. 1473-1478) restudied the solubility of dolomite in the presence of gypsum at 25° C. and CO_2 pressures of one atmosphere and lower. She found, as before, that at one atmosphere CO_2 pressure and in a saturated solution of $CaSO_4$, the solubility differed little from that at the same pressure in pure water. The higher SO_4 concentrations resulted in longer times for equilibrium to set in. However, at .0012 atmosphere (normal CO_2 pressure = .0032 atmosphere) CO_2 pressure, in the presence of gypsum, the solubility of dolomite increased almost tenfold over that in pure water - almost the same as that in pure water at 1 atmosphere CO_2 pressure. Thus the reaction

k) $CaMg(CO_3)_2 + CaSO_4$ ------) $2\ CaCO_3 + MgSO_4$

moves to the right at low CO_2 pressure. This would lead to the buildup of $MgSO_4$ in the waters of the lake basin which then would be periodically precipitated according to reaction (j) above. As this concentration increases further and the $Ca(HCO_3)_2$ is used up, reaction (i) becomes possible. The further possibility of increasing the solubility of dolomite by the combined effects of low CO_2 concentrations and the addition of NaCl in the presence of gypsum was apparently not investigated. In the absence of further information, and in the face of continually changing CO_2 and SO_4 concentrations, it would be difficult to determine the form in which the magnesia would be precipitated at any one time. Low CO_2 concentrations would tend to bring about the precipitation of the basic hydrates. A lowering of the pH due to an increase in SO_4 concentration would tend to precipitate the hydrous carbonates.

The above, then, might have been the process by which the dolomite in the Muddy Mountains may have contributed to the magnesium content of the lakes in which the Overton magnesite was deposited. It follows from the foregoing that gypsum deposition should preceed the magnesite to allow time for the latter to build up in the waters of the lake, and that the location of deposition of a shallow basin in a structurally unstable area may shift from time to time so that the gypsum and magnesite need not be superposed or interbedded. According to Hewitt, et al (1936, p. 168) "At several localities near the south end of the Muddy Mountains the Horse Springs formation contains thick beds of massive gypsum." Elsewhere in the area some gypsiferous beds also occur within the magnesite.

From the discussion preceeding, there remains little doubt that the magnesite deposits occurring in continental lake basins can be accounted for in sedimentary terms. They do not need hot spring or other hydrothermal solutions as the source for the magnesia.

The Currant Creek Deposits, Nevada

Before leaving the subject of origin, one should mention the bone magnesite deposits at Currant Creek, south of Ely, Nevada, described by Faust and Callaghan (1948, p. 11-74) and repeatedly cited as an illustration of hydrothermal deposition. However, before proceeding with that we will need to digress for a moment and discuss briefly the origin of magnesite deposits in serpentine. These generally occur as veins and nodular masses in shear zones and along faults in the ultra basic rocks. The mineral assemblage is simple, consisting principally of bone magnesite with some opaline or chalcedonic silica, rarely quartz, occasionally vein dolomite and calcite. For reasons previously discussed in connection with brucite deposits, one can, at the outset, seriously doubt their hydrothermal origin. Such an origin, as we have seen, would most likely result in the formation of brucite. The writer is not aware of any brucite ever having been reported from a bone magnesite deposit occurring in serpentine. The most reasonable explanation is that the magnesite was derived from the serpentine through weathering and that through several cycles of solution and precipitation, it migrated downward and filled shear and fracture zones in the underlying rock. Initially it may have been precipitated in the soil zone, where organic activity kept the pH high, as brucite or hydromagnesite. Richmond (1938) describes several low-lying, poorly-drained or undrained areas, sometimes a score or more of acres in extent, overlying serpentine in British Columbia containing, under a thin soil cover, a mixture of soil, opaline silica and white clay-like masses of hydromagnesite and other magnesia minerals. The writer has also seen similar deposits in the chrome-serpentine area near Camaguey, Cuba. This material may be reconstituted into nodular magnesite as outlined previously, or it may migrate as the bicarbonate down fracture zones and, at the proper hydrostatic pressure, be precipitated as vein magnesite.

To get back now to the Currant Creek deposits. They occur in a water-worked, bedded to thin-laminated calcareous rhyolite tuff of Tertiary age, 430 feet thick, which rests on basaltic andesite and is overlain by quartz latite flows. The magnesite occurs at several horizons in the limy tuff associated with bedded dolomite, sometimes quartz, and various hydrous magnesium silicate minerals, and as nodules at the surface and also sparsely in faults and fracture zones. It does not occur in either the overlying or underlying volcanics. There is no brucite. The deposit is interpreted by Faust and Callaghan (1948, p. 1-11) as a replacement of the rhyolite by hydrothermal carbonate solutions, largely because the associated hydrous magnesium silicate minerals and quartz are interpreted as being hydrothermal. It might, however, be better explained as the product of a two-stage process. First the magnesite was emplaced as part of a carbonate-tuff fresh water sedimentary sequence, as outlined previously in discussing the Overton deposit. The second stage consisted of the weathering of the first stage material and its redeposition as nodular magnesite in the manner suggested for the deposition of magnesite in serpentine. The second stage may also have taken place during interruptions of the first. Such secondary deposition of magnesite is not uncommon. Bodenlos (1954, p. 128) mentions cryptocrystalline magnesite deposited as veins in the crystalline magnesite of Serra Das Equas, Brazil. However, he interprets this as part of the fourth stage of a hydrothermal metasomatic paragenesis.

The best deposits at Currant Creek lie in low places, some still covered with soil and detritus. The magnesia was probably derived originally from the dolomitic sequences of the Paleozic rocks underlying the Horse Range in which the magnesite occurs or from the White Pine range somewhat farther to the west.

OTHER CRYSTALLINE MAGNESITE DEPOSITS

There can be little doubt from the evidence presented that the crystalline magnesite deposits at Gabbs are of sedimentary origin and that the metasomatic and hydrothermal effects observed have been superimposed subsequent to their emplacement. However, there are other crystalline magnesite deposits in the world whose metasomatic origin has never been questioned. Among the best known of these are, of course, the Austrian deposits in the Millstatt and Veitsch areas (Friedrich, O. M., 1952, p. 401-402), those at Chewelah, Washington (Campbell and Loofbourow, 1957), Kilmar, Quebec (Bray, 1951, p. 49-59), and at Serra Das Eguas and elsewhere in Brazil (Bodenlos, 1954). Could their origin be attributed to sedimentary processes also? However diverse they may be individually, they all occur within dolomites of variable sedimentary sequences usually consisting of alternating dolomite, shale, quartzite, argillite and limestone. They are all intensely folded and faulted, recrystallized and metamorphosed and characterized by a variety of magnesium silicate minerals. One, not too well known, is the crystalline magnesite deposit located northwest of Cranbrook in Southeastern British Columbia, which is worth more comment. (Cairnes, 1932, p. 101-104).

The Cranbrook Deposits, British Columbia

The Cranbrook deposits differ from the others in that they do not occur in dolomite, although dolomite is present. They lie directly south of Marysville across the St. Mary's River on the eastern flank of the Purcell range, which here consists of 37,000 feet of folded pre-Cambrian (Beltian) dolomites, dolomitic shales and shaley quartzites (Rice, 1937). These are overlain unconformably by 7,000 feet or so of Lower Cambrian quartzites and shales, the lowest member of which, the Cranbrook formation, contains the magnesite deposits. The best of the magnesite is 30 to 50 feet thick, but the entire magnesite horizon is 150 feet or more in thickness. It is 4-1/2 miles long and cut off at both ends by faults. It rests on 200 feet or so of white to purplish-pink, in part cross-bedded, quartzite and is overlain by, and interbedded with, greenish quartzite and gray quartzitic shales. At some localities the upper quartzite grades into blue-gray dolomitic limestone. Some local beds of magnesite are also found in the lower quartzite.

The belt south of Marysville trends northeasterly. The sediments are overturned and dip westerly into the slope of the hill at an angle of 60 degrees. The magnesite is generally light gray to creamy-white in color and almost always coarsely crystalline. Even the best of it appears to be contaminated by quartz and quartzite grains, most of which are readily visible to the unaided eye.

Along Boulder and Wallinger Creeks in the rugged terrain of the Rocky Mountains 16 miles northeast of Cranbrook and 22 miles east of the Marysville deposit, the magnesite horizon crops out again.

Here the quartzite overlying the magnesite grades upward into blue limestone. Farther to the east the magnesite horizon grades into conglomerate and sandy conglomeritic carbonate rocks. Apparently the magnesite deposition took place in a shallow sea transgressing an uneven land surface.

It is difficult to interpret the above in any other way than that this uncomplicated deposit is of sedimentary origin.

Chewelah Deposits, Stevens County, Washington

The Chewelah deposits occur within the Stensgar dolomite which is a part of a 16,000 foot folded, faulted and metamorphosed sequence of sediments of Algonkian age, consisting of shales, slates, quartzites, and dolomite, occupying the west flank of the Huckleberry Mountains in central Stevens County, Washington. (Campbell and Loofbourow, 1957). The northeasterly trending Stensgar dolomite, 300 to 800 feet thick, crops out discontinuously for 25 miles extending from the Colville River in the north to the Spokane River in the south, where the sediments are intruded by the Lune Lake granite of Cretaceous age. The Stensgar dolomite grades downward into the McHale slate (which in turn rests on the Edna dolomite) and is overlain by the slate and phyllite of the Buffalo Hump formation. Most of the magnesite deposits are found in the northern part of the belt, particularly in the area lying 8 miles west of Chewelah. The magnesite, ranging in color from creamy white through gray and pink to almost hematite red, is generally medium to coarsely crystalline, and occurs as steeply-dipping, concordant lenticular masses, often 200 feet thick. The magnesite is contaminated by greenstone dikes, chert, quartz, lime in the form of dolomite, and a variety of magnesium silicates. The chert occurs principally in a persistent horizon, one to two feet thick, near the base of the magnesite, in the dolomitic transition to the McHale slates, a score of feet or so above the contact. The magnesite is also traversed by several dolomitized faults and other dolomitic zones. Some in particular, dike-like and slabby, are known locally as dolomite sidewalks.

There are some parallels here with the deposits at Gabbs - the dolomitized faults, for instance. The other dolomitized zones may correspond to the dolomitic flexture zones and recrystallized early faults. The dolomite sidewalks could be the recrystallized equivalents of the clastic dolomite dikes at Gabbs. A sedimentary origin for these deposits also seems to be the most reasonable explanation.

<u>Magnesite at Kilmar, Quebec, Canada</u>

Kilmar lies 10 miles north of Calumet, which is about halfway between Montreal and Ottawa. The magnesite occurs in a northerly-trending belt of pre-Cambrian (Grenville series) metamorphosed, steeply-dipping sediments composed of sillimanite-garnet gneiss, quartzite, argillite, and in part serpentinized, crystalline, high-magnesia carbonate rocks. The magnesite layer is about 200 feet thick and is confined within a zone of dolomite and dolomitic limestone 1,000 feet or so across. This is bounded on the west by quartzite which, farther to the west, gives way to granite, and on the east by granite and granitic gneisses and schists. Silication of the carbonates has been extensive. The dolomite has been in part converted to diopside, and magnesite to serpentine with the development, in places, of very pure short fibre asbestos. Mobilization of magnesia has also taken place. Some of the quartzite, chert, and quartzitic horizons in the carbonate rocks have been converted to magnesium silicates. These processes operated in several stages. Most of them took place before the intrusion of the granite. Others, no doubt, are attributable to it.

Surprising is the absence of any appreciable quantities of brucite or bone magnesite which, if the granite had much to do with bringing about the silication of the magnesite or magnesitization of the quartzite, should be everywhere in evidence. A possible explanation might be that in the presence of much silica, autosilication takes place, and the silicates are formed rather than the hydroxides. In any event, and in spite of the mineralogical complexity of these deposits, one can still discern an interbedded sequence of dolomite, in part silicated to diopside, and magnesite, partly silicated to serpentine, and quartzite. More chert may have been present, but it has been largely used up in the silication of the carbonate rocks. A sedimentary origin offers the best explanation for these deposits also.

CONCLUSIONS

It is clear from the arguments presented that hydrothermal replacement or depositional processes play but a trivial, secondary role in the formation of magnesite deposits. The crystalline variety originates in marine environments through sedimentary processes. The clay-like beds deposited in interior basins are likewise completely of sedimentary origin. The bone magnesite deposits occurring in serpentine are derived from the serpentine through weathering.

The concept of the sedimentary origin of crystalline magnesite deposits has had a precarious emergence. The prestige of hydrothermal-metasomatic hypothesis, which has been applied with so much success particularly by Emmons and Lindgren, in explaining the origin of metalliferous deposits, made it always a safe haven for the timid souls who might be disconcerted by contradictory evidence in the rocks. It is time for a searchingly critical re-examination of this hypothesis in order that its application may be refined where it offers the most reasonable explanation and discarded where it clearly does not apply so that the door might be opened to new thought.

A sedimentary origin for crystalline magnesite deposits has been suggested before. Niinomy (1925, p. 25) first suggested such an origin for the deposits in Manchuria. Lately this suggestion has been repeated by Nishihara (1956, p. 698-711).

The crystalline magnesite deposits, clearly of sedimentary origin, lying astride the boundary between Spain and France in the western Pyrenees have been described by Joaquin Gomez de Llarena (1952). Ore bodies of thin-bedded to platey, fossiliferous magnesite occur at several horizons in the shale and carbonate parts of Upper Devonian and Lower Carboniferous sediments consisting of conglomerate, sandstone, graywacke, shale, and carbonate rocks. A characteristic feature of the magnesite is the black banding by organic material consisting mostly of marine plants. In a later work (Gomez de Llarena, 1953, p. 55-62) he suggests, after examining the literature, that the

Veitsch deposits in Austria might also be of sedimentary origin. A sedimentary origin for the deposits at Gabbs has also been previously suggested. (Martin and Willard 1957, p. 426).

Even though magnesite occurs most often in dolomite sequences, it is apparent that this is not the only criterion. Extensive areas of dolomite, as for instance those of Silurian age in eastern Ohio, western Indiana and southern Michigan, though they may also contain gypsum and show evidences of abundant organic activity, need not necessarily have been laid down under conditions favorable for the deposition of magnesite. Rather, the dolomite-magnesite deposition under discussion is a local facies occurring in special physical as well as ecological conditions in the unstable, embayed shelf area of an extended geosynclinal trough - an area lying between an emergent and eroding low land mass on the one hand, and the deeper sedimentary basin on the other. In this zone one finds the frequent changes in sedimentary environments, reworking of previously deposited sediments, and alternating deposition and erosion within restricted lagoonal embayments teeming with organic activity. If some of the sediments being reworked are dolomite, it is possible that magnesite may be deposited.

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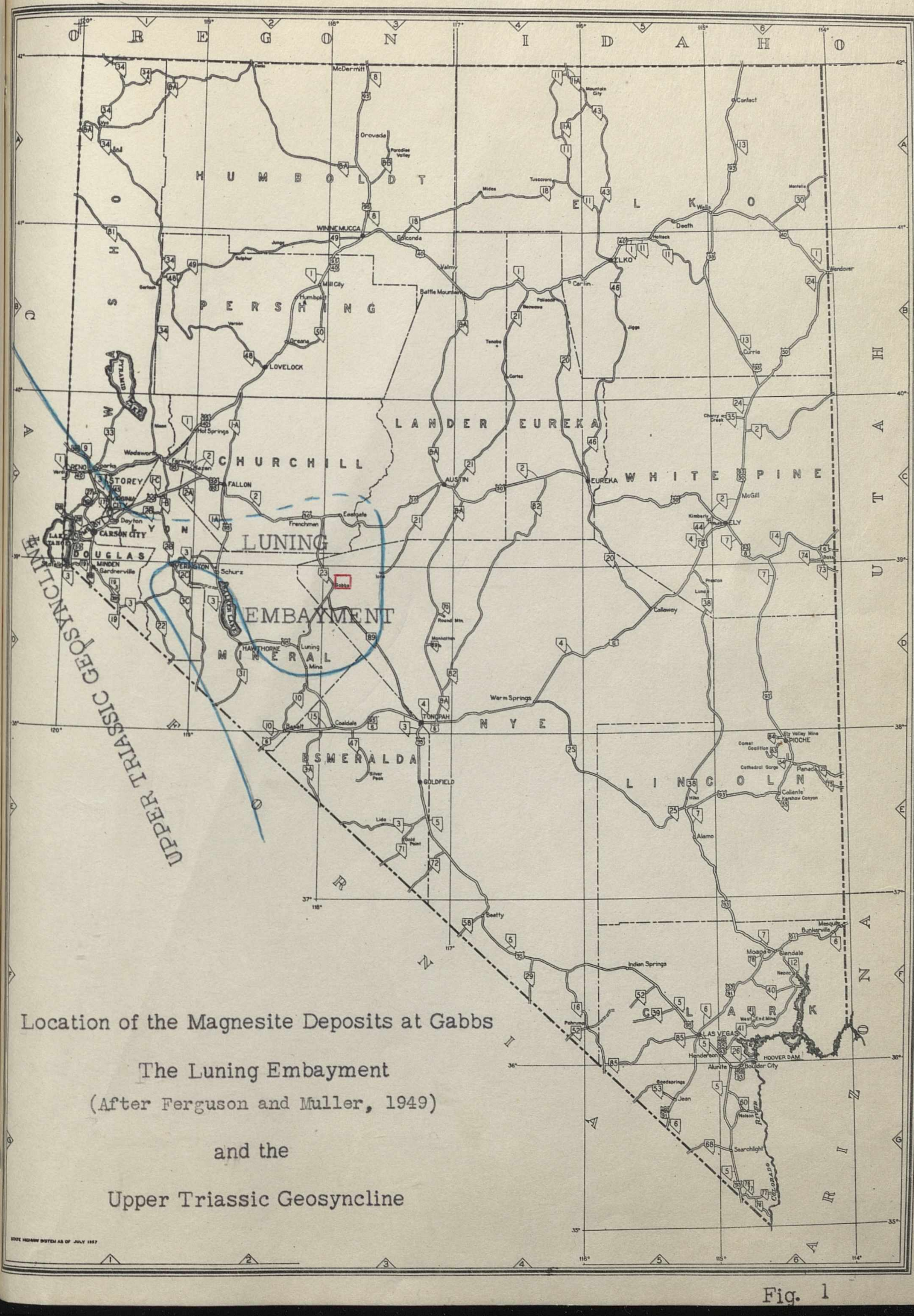

Location of the Magnesite Deposits at Gabbs

The Luning Embayment

(After Ferguson and Muller, 1949)

and the

Upper Triassic Geosyncline

Fig. 1