Geology of the Magnesite Belt of Stevens County Washington

By IAN CAMPBELL *and* JOHN S. LOOFBOUROW, JR.

CONTRIBUTIONS TO ECONOMIC GEOLOGY

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Geologic setting and brief description of the magnesite deposits

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By IAN CAMPBELL and JOHN S. LOOFBOUROW, JR.

ABSTRACT

The magnesite belt of Stevens County, Wash., is a narrow zone of closely folded metasedimentary and intrusive rocks in the northeastward-trending Huckleberry Mountains. Most of the rocks belong to the Deer Trail group, generally believed to be late Precambrian. This group, the oldest rocks exposed within the area and probably of Belt age, comprises a series of metasedimentary rocks composed dominantly of fine-grained clastics, and of lesser amounts of quartzite and dolomite, totaling at least 10,000 feet in thickness. Formations mapped within the Deer Trail group are the Togo formation (oldest), Edna dolomite, McHale slate, Stensgar dolomite, and Buffalo Hump formation. Lying unconformably on these rocks is the Huckleberry formation (conglomerate and greenstone members). Unconformably above the Huckleberry formation lies the Addy quartzite of Early Cambrian age.

These formations, together with greenstone intrusives, have been closely folded. The dominant structure is the steeply dipping locally overturned west limb of a northward-plunging anticline. In many places longitudinal faults have caused parts of the section to be repeated. Drag folds, indicating relative northward movement of westerly blocks, are conspicuous. Emplacement of a Cretaceous granitic batholith has considerably altered parts of the dolomitic and argillitic facies of the metasedimentary rocks, converting dolomite to diopside hornfels and argillite to biotite-cordierite phyllite.

Commercial magnesite deposits in the belt are confined to the Stensgar dolomite of the Deer Trail group. The magnesite probably resulted from replacement of dolomite by hydrothermal solutions related in origin to the granitic batholith. Since their discovery in 1916, the magnesite deposits have constituted the most important mineral resource of the area, producing several million tons of refractory-grade magnesite, and reserves of a few million tons remain.

More than $3 million worth of silver, lead, and zinc was produced in the area before World War I, but subsequent production of metalliferous ore has been sporadic and minor. Other mineral deposits of actual or potential interest include antimony, barite, copper, quartzite, slate, talc, tungsten, and uranium.

INTRODUCTION

In 1920, Weaver (p. 319) recognized that the magnesite deposits of northeastern Washington are limited to a narrow but well-defined zone lying within the northeastward-trending Huckleberry Mountains. Since then, the term "magnesite belt" has come into common use (Bennett, 1941, p. 3) as a convenient means of designating this area. As employed in this report, the term refers not only to the Stensgar dolomite, to which the magnesite deposits are restricted, but also to adjacent formations. Our studies, initiated as a war-time minerals project, were first directed toward providing an outcrop and distribution map of the Stensgar dolomite as a supplement to the work of Eugene Callaghan, who in 1941 had mapped the principal magnesite deposits. As our fieldwork advanced, the original plan was enlarged to include a moderately detailed map not only of the Stensgar dolomite but also of the adjoining formations insofar as this information might add to the accuracy of the areal and structural interpretations. Thus, some of the formations mapped and discussed in this report have had little more than cursory study, whereas others have been studied in considerable detail.

LOCATION AND ACCESSIBILITY

The mapped area, approximately 33 miles long and from 2 to 5 miles wide, covers slightly more than 100 square miles in the south-central part of Stevens County, in northeastern Washington (fig. 1). Chewelah, in the northeastern part of the area, is the principal town and shipping point for the processed magnesite. Chewelah is 55 miles north of Spokane on a branch line of the Great Northern Railway and on U.S. Highway 395, which also passes through Valley and Springdale, smaller communities south of Chewelah and slightly east of the mapped area. From Chewelah a hard-surfaced road leads to the Finch and Allen-Moss magnesite quarries, 4 miles to the southwest, where the mill of the Northwest Magnesite Co. is located. Good secondary roads lead to the Keystone and Red Marble magnesite deposits, which may also be reached by road from Valley. The local Forest Service maintains an extension of the Red Marble road to Stensgar Peak Lookout, thus permitting easy access to one of the highest points (5,808 ft) in the Huckleberry Mountains. In the central part of the area a well-graded road from Springdale to Hunters affords the only good route across the Huckleberry Mountains, though an unimproved road by way of the Cleveland mine also crosses the mountains 2 miles north of the Springdale-Hunters highway. Additional roads, in various states of repair, afford access to the ranches, logging camps, mines, and prospects of the region.

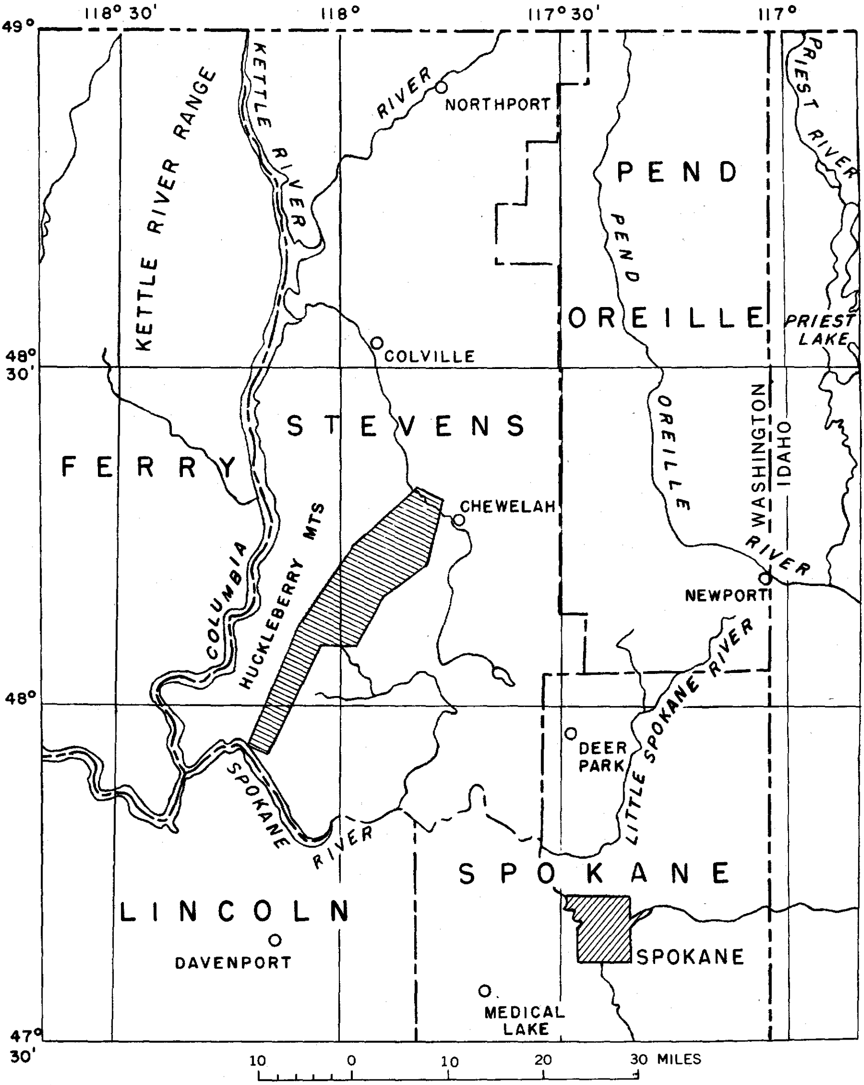

FIGURE 1.—Index map of northeastern Washington showing location of mapped area.

TOPOGRAPHY

The two dominant topographic features of the area are the Huckleberry Mountains and Colville Valley. The greater part of the magnesite belt lies within the Huckleberry Mountains, a north-

northeastward trending range with rugged slopes and a fairly even crest, much of which is more than 5,000 feet in altitude and 3,300 feet above the Colville Valley. The Huckleberry Mountains are one of the minor ranges making up the Selkirk system (Weaver, 1920, p. 30). Colville Valley and its alluviated trench flank the magnesite belt in the northeastern part of the area west of Chewelah.

Many small streams drain the slopes of the Huckleberry Mountains. On the east side, from north to south, Huckleberry, Deer, and Camas Creeks are the principal streams tributary to the north-flowing Colville River. The only major stream on the west side of the mountains is Hunters Creek, which flows into the southward-flowing Columbia River. A few small glacial lakes occur in the area, especially in the northeastern part.

CLIMATE AND VEGETATION

A 13-year record (U.S. Dept. Agriculture, 1941, p. 1173) for Chewelah (alt. 1,700 ft.) shows an average annual precipitation of 18.58 inches. The driest months are July (0.38 in.) and August (0.56 in.); the precipitation is rather evenly distributed over the remaining months, each of which averages from 1 to 3 inches. Much of the precipitation from October to April is in the form of snow. The minimum temperature recorded is —38°F, the maximum is 107° F; average temperature for January (the coldest month) is 23.8°; for July (the hottest month), 65.2°.

Lumbering has been one of the major industries of the region for more than 40 years, and not many of the original stands of western yellow pine now remain. Fires, too, have taken their toll, but nevertheless the greater part of the area is still heavily forested. Much of the cover now consists of second-growth fir, cedar, hemlock, and alder, forming at times an almost impenetrable jungle which stands in unfortunate contrast (from the viewpoint of conservationist and field geologist alike) to the open parklike glades of yellow pine.

FIELDWORK AND ACKNOWLEDGMENTS

As part of the Geological Survey's Strategic Minerals program, we mapped the magnesite belt in 1942 and 1943. Campbell continued work in the area in 1946, assisted by L. S. Gurney, and again in 1949, assisted by T. M. Phetteplace. We were aided in the mapping by others concerned with the Survey's program, particularly by Eugene Callaghan, who very kindly made available a complete set of his field maps of the magnesite quarries and helped us in many other ways, and by C. F. Deiss, who spent several days in the field with us and provided measured sections of the Stensgar dolomite.

To the officials and engineers of the Northwest Magnesite Co.—in particular to Mr. E. A. Garber, president; Mr. C. Arthur Sargent, vice-president; Mr. Howard A. Ziebell, chief engineer; and Mr. Roger L. Fisk, mine superintendent—we are indebted for permission to visit the company properties, for much pertinent information, and for the use of maps and other data. Dr. Harold E. Culver, formerly Supervisor, Division of Geology, State of Washington Department of Conservation and Development, provided helpful information in both the field and the office. We are further indebted to Dr. Culver for opportunity to consult R. H. B. Jones' original maps and manuscript on the Chewelah quadrangle. We have benefited greatly by field conferences with Dr. W. A. G. Bennett, also of the Washington Department of Conservation and Development, and even a cursory comparison of our map with the earlier map published by Bennett (1941, pls. 1, 2) will testify that we owe much to his careful work in this difficult area. We record also the cooperation of personnel of the Washington State Department of Highways, the State Division of Mines, and the U.S. Soil Conservation Service. The California Institute of Technology provided us with office facilities during preparation of the report.

Many base maps were used in the course of the work. The Chewelah quadrangle topographic sheet, at a scale of 1:125,000, was used for some general reconnaissance. Parts of four sections between the Keystone quarry and Huckleberry Creek were mapped by planetable at a scale of 1,000 feet to the inch and with a 20-foot contour interval. The Phoenix, Nogue, Mountain View, and Allen-Moss deposits were mapped by planetable at a scale of 100 feet to the inch and with a 10-foot contour interval. Aerial photographs were used in the mapping, mostly at a scale of 1:36,000 except in the northern part of the belt, where the scale used was 1:20,000, and in the extreme southern part, where it was 1:32,000. Through the kindness of Richard T. Kriebel, of the Polaroid Corp., we were provided with vectographs for some of the more critical parts of the area. We also had the advantage of multiplex topography prepared at a scale of 1,000 feet to the inch and with a 10-foot contour interval. From all these sources our final map, at a scale of 1:36,000 and with a 100-foot contour interval, was compiled by Mr. E. H. Quayle, of the University of California at Los Angeles.

PREVIOUS WORK

The first mineral production in Stevens County was about 1865, but it was not until the late 1880's that mining became active in the county and not until the 1890's that the Deer Trail (Cedar Creek) district—the most important of the metalliferous zones within the magnesite belt—came into production (Bancroft, 1914, p. 2). Although this attracted mining engineers and geologists to the region, information published before 1920 is scant; most of it mentions the geology only briefly, incidental to description of the metal mines and prospects. Among such may be mentioned the papers of Bethune (1891), Burdsal (1896), Hodges (1897), Thyng (*in* Landes and others, 1902), Leith (1906, p. 195), Collier (1907), Stiles (1912), and Hill (1912). Nonmetallic resources of the county did not receive attention until about the turn of the century, when clay, limestone, and marble began to be mentioned in the literature: Landes (1902, 1906, 1911); Shedd (1903, 1910, 1914); Eckel and others (1905); Rathbun (1906). These scattered brief reports and the somewhat more comprehensive geologic observations of Bancroft (1914), and of Pardee (1918) on an adjoining area, constitute the bulk of the literature available until the time when World War I focused attention on magnesite. From the literature of that time, it seems that the rocks of the region had been recognized as consisting principally of metamorphosed sandstones, shales, and limestones, variously assigned to the Precambrian and the Paleozoic, intruded by greenstone dikes and sills and by granitic batholiths, the last probably of Mesozoic age.

Although Bauerman (1885) in 1859–60 made a general geologic reconnaissance of the region, no comprehensive report appeared until Weaver's bulletin (1920) on mineral resources of Stevens County. This report not only described all the mines and prospects of Stevens County, but it provided the first geologic map of the area and established the first stratigraphic section. To Weaver, too, we owe the first relatively detailed account of the magnesite deposits, though earlier and coincidental notices had appeared: Yale and Stone (1921), Jenkins (1918), and Dolman (1920).

With the establishment of a magnesite industry in the area in 1917, attention was drawn to the magnesite belt itself and a number of papers appeared in the next quarter century. Whitwell and Patty (1921) relied on Weaver's work for the general geology, but they added some detail on the magnesite deposits and gave an excellent summary of the state of the industry as it then existed. Many reports dealt with the origin of the magnesite, the extent of reserves, and so forth, such as the papers of Petrascheck (1924), Bain (1924), Siegfus (1927), Landes (1934), Glover (1936), and Culver (1939). Yet, as regards the general geology and structure of the magnesite belt, only a few investigations since Weaver's time deserve mention. In 1920 Bailey Willis with several assistants spent a season mapping the magnesite belt. Willis prepared the first topographic map of the area and drew important conclusions as to stratigraphy and structure and the relation of the magnesite to these features. Unfortunately, Willis' only published work (1927) on the area is largely a discussion of the paper by Siegfus (1927), but Hodge (1938, p. 36) published a map which Bennett (1941, p. 6) believed was a reproduction of Willis' geologic map of the area. On the basis of fieldwork by E. K. Judd, Hodge presented a condensed statement (1938, p. 35–38) on the structure and stratigraphy. In the meantime, R. H. B. Jones had mapped the Chewelah quadrangle in fulfilling the requirements for a master of science degree at Washington State College. Jones' only published paper (1928) is rather brief; in his original manuscript and map[1] he differentiated several additional stratigraphic units, and to him goes the credit for first recognizing a conglomerate (conglomerate member of the Huckleberry formation) at the base of the principal greenstone flows. In 1941 Bennett published a preliminary report on the magnesite deposits, including an excellent map. He considerably revised and improved upon the earlier structural and stratigraphic interpretations of Weaver, Willis, Jones, and Hodge. Our own indebtedness to Bennett is very great, and the differences that appear between our map and structural interpretations and those of Bennett are largely due to the additional detail which our larger scale base and the availability of aerial photographs permitted us to record. A later paper by Bennett (1943) gives detailed information on the Turk magnesite deposit and has proved most useful in extending the stratigraphy of the magnesite belt to the south. Good annotated bibliographies of the literature on the region are available in the papers by Weaver (1920) and Bennett (1941), and little additional information has appeared since 1941.

BEDROCK GEOLOGY

The rocks of the magnesite belt (pl. 1) consist mainly of a thick series of mildly metamorphosed sandstones, shales, and dolomitic limestones of Precambrian age. The thickness of these rocks within the mapped area is more than 10,000 feet, but, as the base is not exposed in the area, the total thickness may be considerably greater. All the known magnesitized zones are confined to a dolomite in the upper part of this series. Greenstone dikes and sills occur throughout the series.

Overlying the metasedimentary rocks in most of the area is a greenstone conglomerate, locally as much as 3,000 feet thick (Bennett, 1941, p. 8). This conglomerate, in turn, is overlain by relatively massive greenstone flows, aggregating nearly 3,000 feet, which probably are the youngest Precambrian rocks in the area. The flows are uncon-

[1] Jones, R. H. B., 1927, Geology and mineral resources of the Chewelah quadrangle: Master of Science thesis, on file in the Division of Geology, Washington State College, Pullman, Wash.

formably overlain by a Lower Cambrian quartzite, the top of which was not mapped but which is known to be at least 3,000 feet thick (Bennett, 1941, p. 9).

Small masses of granite of Cretaceous age occur within the area and are related to larger bodies, such as the Loon Lake granite (Weaver, 1920, p. 87), and possibly the Colville batholith (Pardee, 1918, p. 30–34; Waters and Krauskopf, 1941, p. 1358). Rhyolite intrusives, possibly belonging to the same igneous epoch, are local minor features. In the extreme southern end of the belt, a Tertiary andesite flow overlaps the older formations.

Structure is complex, consisting of broad to tight folds, locally overturned, and strike faults of large displacement and pronounced drag, which, in parts of the area, have repeated some beds as much as three times.

PRECAMBRIAN ROCKS

DEER TRAIL GROUP

Recognizing the impossibility of a satisfactory correlation of rocks in Stevens County with formations established by Daly (1912) along the international boundary about 60 miles north of the magnesite belt, Weaver (1920, p. 50) tentatively proposed the name Stevens series for all the metamorphosed sediments exposed in this county. He proposed (1920, p. 59) the name Deer Trail argillite for an assemblage of argillites, slates, phyllites, quartzites, dolomites, and limestones covering a broad belt in the central part of the county and including much of the magnesite belt. To this same assemblage Bennett (1941, p. 7) gave the name Deer Trail group, and this designation we have retained. Bennett recognized that, although the group consists predominantly of slates, it also includes two conspicuous quartzite and two dolomite beds. These he designated Upper quartzite, Lower quartzite, Stensgar dolomite (in this instance retaining a name first proposed by Weaver), and Lower dolomite.

Throughout most of our first field season we found that the units established by Weaver and Bennett served satisfactorily, but as work progressed it became apparent that additional quartzite beds had stratigraphic value. To avoid the confusion of such terms as "lower quartzite," "lowermost quartzite," and "lower middle slate," we propose new formation names for several of these units. We have divided the Deer Trail group into the following formations, from bottom to top: the Togo formation, Edna dolomite, McHale slate, Stensgar dolomite, and Buffalo Hump formation.

TOGO FORMATION

The oldest rocks in the Deer Trail group we designate the Togo formation because of good exposures in the vicinity of the Togo mine (sec. 6, T. 29 N., R. 38 E.). From Huckleberry Creek south the Togo formation makes up the greater part of the eastern border of the mapped area. Since the base of this formation was not mapped, it is not known how much farther it may extend to the east. Rocks that total more than 4,000 feet in thickness are included in the formation in the southern part of the belt, but the total thickness may be much greater. Much of this formation consists of monotonously dark-gray to almost black rather slaty argillite that erodes easily and provides few outcrops. Towards the top of the formation distinctive thin beds of laminated black and white (or light-gray and dark-green) siliceous slate provide many more outcrops. Thin impure dolomitic and calcareous beds that cannot be traced continuously occur near the top of the formation.

At the top, and serving to some extent as a marker throughout much of the belt, is a quartzitic facies which in most places gradationally succeeds the argillitic facies. This quartzitic facies thickens notably from north to south. In the latitude of the Double Eagle (secs. 19, 29, and 30, T. 31 N., R. 39 E.) it is only a few tens of feet thick, but in the latitude of the Orazada mine (sec. 10, T. 29 N., R. 37 E.) it is nearly 1,000 feet thick. Its quartzitic character, too, becomes more pronounced to the south. In much of the magnesite belt this quartzite unit of the Togo is not a truly massive quartzite (as are many of the quartzite beds higher in the section), for it ranges from quartzitic slate and argillite to slaty quartzite and includes a few thin beds of nearly pure quartzite. Intercalary beds of the laminated siliceous slates mentioned above are common. This unit corresponds, we believe, to the "lower quartzite" of Bennett (1941, p. 8), which is "overlain by a gray dolomite bed [this corresponds to part of our Edna dolomite] about 100 feet thick and underlain by a distinctive slate zone that is dark gray to black with white bands."

The actual contact of the Togo formation with overlying Edna dolomite generally is poorly exposed, but the lithologic change from the distinctive quartzites in the top of the Togo to the dolomite of the Edna is rather abrupt: commonly an outcrop of quartzite will be found, and up the dip, 50 feet or more away, will be a characteristic small outcrop of Edna dolomite. Although this contact must be approximated on the map, it has proved to be one of the most valuable datum planes in a complex and often confusing stratigraphic section.

EDNA DOLOMITE

Bennett (1941, p. 7-8) recognized and mapped a persistent dolomite below the Stensgar and referred to it on his map as "lower dolomite." This unit is here named the Edna dolomite, from characteristic outcrops in the vicinity of the Edna mines (sec. 9, T. 31 N., R. 39 E.).

The Edna can be traced from the northern border of our map, north of Colville Valley, almost to the southern border, where it disappears beneath the Tertiary Gerome andesite. The formation is dominantly an impure dolomite, but it includes facies of quartzite, argillite, calcareous slate and phyllite, and talcose slate. The total thickness of the formation probably is from 1,500 to 2,500 feet in most of the belt, but we could make no accurate measurement because of poor exposures. The Edna commonly underlies low areas, and, in much of the magnesite belt, the formation has been projected largely on the presence of swampy areas, faint suggestions of karst topography, and heavy vegetation.

The dolomitic facies is particularly well exposed in the bluffs north of the county road in secs. 5 and 6, T. 31 N., R. 40 E. The rock is a light to dark gray, locally almost black, fine-grained, siliceous, ferruginous, and calcareous dolomite. Closely spaced fractures, many filled with secondary quartz, are common; relatively deep weathering is characteristic. Bedding in some places is very thin and may show considerable minor contortions; in other places, due to weathering, mineralization, or lack of original bedding, outcrops are devoid of bedding, and attitudes are difficult or impossible to determine.

Chemical analyses (table 1) of five selected specimens from widely separated localities indicate a general uniformity of composition over long distances. The consistently higher content of iron and silica ("insoluble") and lower content of magnesia make an interesting comparison with the Stensgar dolomite (table 2).

The quartzite unit that we have mapped at or near the top of the formation is well exposed along the Red Marble road, in secs. 19 and 20, T. 31 N., R. 39 E., where its presence enabled us to map a small fold (fig. 2) that otherwise would have been almost indistinguishable. The rock is a white or bluish-gray, locally reddish-brown, fine-grained to almost flinty quartzite; bedding planes, in contrast with those in the quartzites of the underlying Togo formation, are rarely determinable. This quartzite unit in the Edna can be traced, with few uncertainties, from the latitude of the Double Eagle quarry to the latitude of the Cleveland mine road. Elsewhere it is somewhat erratic both in position and distribution, and in places more than one quartzite may be present.

The slaty argillite facies shown in the southern half of the map area is well exposed on the Springdale-Hunters road. The rock is dark gray and poorly bedded.

McHALE SLATE

Bennett (1941, p. 8) mentions that below the Stensgar dolomite there occurs a "considerable thickness of Deer Trail slate," but on the map accompanying his report none of the many slate units within the Deer Trail group is specifically differentiated. As a matter of fact, the abundance and the very great similarity of these slaty units constitute one of the chief obstacles in mapping the geology of the magnesite belt. In our work we were rarely able to place any confidence in lithology as a means of distinguishing one body of slaty rocks from another within the Deer Trail group. The only basis for separating and recognizing different units is their stratigraphic position with respect to more readily distinguished markers, such as the Edna dolomite, the Stensgar dolomite, and the quartzite of the Togo formation.

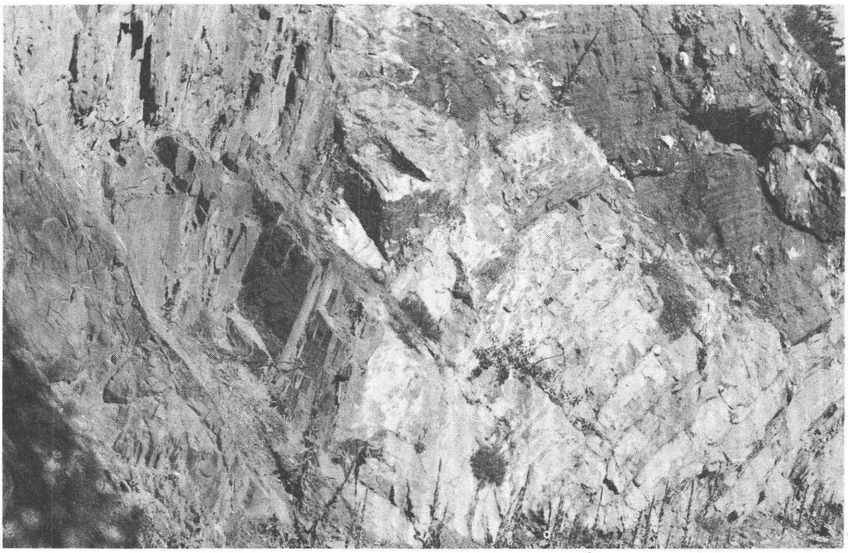

FIGURE 2.—A fold in the Edna dolomite exposed by a cut in the Red Marble road in the S½ sec. 20, T. 31 N., R. 39 E. White areas are thin coatings of milky quartz on joint planes in the dolomite. A strongly jointed greenstone intrusive appears in the left foreground.

Throughout the entire length of the magnesite belt, the Edna dolomite is conformably overlain by a slate which, because of good exposures in and north of McHale Canyon,[2] we have named the McHale slate. It should be recognized that this term includes facies that could be more correctly termed "argillite" and "phyllite." The thickness of the McHale slate is difficult to estimate because the formation has been so greatly folded and squeezed, but throughout most of the belt it seems to be between 1,000 and 1,500 feet. It crops out exten-

[2] McHale Canyon is a local name for a rather striking east-west canyon, northeast of Finch quarry. A secondary road shown on our map in sec. 19, T. 32 N., R. 40 E. indicates the course of this canyon.

sively, and good exposures, in addition to those in McHale Canyon, were found along the Red Marble road and along the Springdale-Hunters highway.

TABLE 1.—*Analyses of Edna dolomite from Stevens County, Wash.*

[Analyses by V. North, U.S. Geological Survey, 1944]

	CM-9	CM-10	CM-11	CM-12	CM-13
Insoluble	28. 14	27. 02	25. 24	30. 95	19. 67
R_2O_3	. 35	. 27	. 23	. 15	. 20
CaO	21. 96	22. 08	22. 58	21. 55	25. 56
MgO (calculated)	14. 16	14. 08	14. 70	13. 04	15. 44
FeO	1. 65	2. 39	1. 98	1. 97	1. 38
CO_2 (calculated)	33. 74	34. 16	34. 97	32. 34	37. 75

CM-9 Red Marble road, 350 ft east of the junction of the Wells-Fargo road; SE¼ sec. 20, T. 31 N., R. 39 E.
CM-10 Red Marble road, 2,400 ft east of the junction of the Double Eagle road; SE¼ sec. 19, T. 31 N., R. 39 E.
CM-11 SE¼ sec. 27, T. 32 N., R. 39 E.
CM-12 U.S. Highway 395, about 3 miles north of Chewelah, in the NW ¼ sec. 4, T. 32 N., R. 40 E.
CM-13 Springdale-Hunters road, 400 ft west of the summit, in the SE¼ sec. 16, T. 30 N., R. 38 E.

Colors of this formation are chiefly the darker shades of gray, green, and brown; but a common and sometimes striking characteristic is a color banding due to alternations of light- and dark-colored layers, locally only fractions of an inch thick. Such bands may represent original bedding. Slaty cleavage is locally well developed, but at other places no cleavage is discernible (argillite facies). However, the greater part of the formation shows a rather notable cleavage or schistosity which may parallel the bedding just mentioned or may cut across it at low to high angles.

An unusual facies, first called to our attention by Dr. Bennett, is an ankeritic slate, best exposed in the southeast corner of sec. 19, T. 32 N., R. 40 E., near an old fork in the McHale Canyon road. Here a very fissile rather light-gray slate contains abundant metacrysts of pyrite and ankerite, both of which have been largely altered to limonite. Similar facies have been found elsewhere in the area, but unfortunately not in such position as to give the facies any value as a marker horizon. It seems likely, however, that the development of ankerite and pyrite depends as much on proximity to an intrusion (greenstone dikes are near all occurrences) as on any original sedimentary feature.

Other attempts to place stratigraphic value on the diverse facies in the formation have also been unsuccessful. The distinctively thin-banded slate, for example, was found in many parts of the magnesite belt, chiefly in the lower parts of the formation; but similar facies were at times also found in middle and upper zones. Toward the top of the formation, there is some indication of a gradational change toward the Stensgar dolomite, the next overlying formation. Here the slate is lighter colored, has a less well developed cleavage, and is interbedded with zones of carbonate.

STENSGAR DOLOMITE

Conformably upon the McHale slate lies the Stensgar dolomite, which is a particularly important formation in this study, not only because it contains all known occurrences of magnesite, but because it is one of the few formations in the area sufficiently distinctive to be recognized with confidence in almost all outcrops. It has been used, therefore, as a major marker and has furnished a key to much of the structural pattern.

This formation was first named by Weaver, who included it with his Deer Trail argillite (Weaver, 1920, p. 58–59) but described it separately (1920, p. 57) and differentiated it on his map. Bennett (1941, p. 7–8) retained the name and recognized the Stensgar dolomite as a distinct formation within the Deer Trail group.

The Stensgar dolomite can be traced from north of Colville Valley to a point about half a mile south of the northern boundary of the Spokane Indian Reservation. In parts of the belt, it has been completely cut out by strike faulting; elsewhere, as in the Red Marble section, thrust faults have repeated the formation as many as three times.

The Stensgar dolomite not uncommonly appears in bold outcrops, but nothing approaching a continuous section is available except at the magnesite quarries and along U.S. Highway 395 north of Chewelah, all of which present atypical sections. On the whole, the typical Stensgar is a fine-grained dense light-bluish or pinkish-gray dolomite which weathers medium gray or light buff. Except where extensively recrystallized, the dolomite commonly is somewhat thin bedded and bedding-plane attitudes can be measured at most outcrops. This is fortunate, for such data can seldom be obtained from the other formations of the area.

At the northern end of the Stensgar belt, an abandoned quarry in the NE¼NE¼ sec. 32, T. 33 N., R. 40 E., affords excellent exposures. At this site C. F. Deiss measured a section of Stensgar dolomite from the base upward along the crest of a nearly isolated hill on the east side of Colville Valley, approximately 4.7 miles northwest of Chewelah. Most of the formation is exposed on the crest of the hill, and the transition from the underlying McHale slate to the Stensgar dolomite is completely exposed on the old quarry road at the southeast end of the hill. Black-gray well-banded slaty argillite grades up into dull-green and maroon interbedded banded argillite which becomes dolomitic up section, becoming interbedded with dolomite, and finally grades into continuous beds of buff-weathering dolomite. The base of the Stensgar was arbitrarily taken at the base of the first thick bed of dolomite above which dolomite exceeds argillite in the section.

The Stensgar at this locality is overlain by a thin lens of the conglomerate member of the Huckleberry formation, but at the northwest end of the hill the conglomerate member is cut out by an erosional unconformity and the Stensgar is overlain directly by the greenstone member of the Huckleberry. The actual contact is covered with soil and glacial drift.

Section of Stensgar dolomite of Precambrian age, 4.7 miles northwest of Chewelah, Stevens County, Wash.

	Feet
Top of Stensgar covered.	
17. Dolomite, as in unit 16	16
16. Dolomite, as in unit 15 but contains even more white quartz in veins 1/16 to 2 in. thick. All beds much jointed and some brecciated. Beds in upper part of unit stained dull red	61
15. Dolomite, gray and tan mottled; some beds mottled red; medium to coarsely crystalline, recrystallized, thick-bedded; contains large dark gray-blue dolomite crystals, much limonitic clay, and veins and lenses of white quartz as much as 1 in. thick. Rock weathers drab tan brown. Some beds partly brecciated	78
14. Dolomite and argillite: dull-gray finely crystalline platy-bedded dolomite and red and gray dolomitic argillite in parting beds. Near middle of unit several beds of blue and tan coarsely crystalline dolomite contain veins of white quartz as much as 4 in. thick	17
13. Dolomite, pale-blue and gray mottled, coarsely crystalline, thick-bedded, hard; contains flakes and nodules of limonitic clay and many 1- to 2-mm veins of quartz. Some veins ½ to 1 in. thick. Strike N. 24° E., dip 36° NW	21
12. Unit four-fifths covered. Dolomite, in lower part, cream-gray, platy-bedded, slightly argillaceous. Drift and soil cover upper part of unit. Well exposed in quarry	50
11. Unit covered by glacial drift. Probably dolomitic argillite and pure argillite	58
10. Dolomite, finely crystalline, thick- and thin-bedded; more distinctly bedded in lower part. Contains much limonite and quartz in veins and irregular lenses. Top bed of unit mottled red coarsely crystalline dolomite which forms large dip slope at crest of ridge	39
9. Dolomite, buff-gray, cryptocrystalline, thick-bedded to platy; contains much silica; weathers drab tan and forms partly covered slope on crest of ridge	30
8. Dolomite, pale-gray, dense, finely crystalline; much thicker bedded than in unit 7; contains many stringers and veinlets of white chalcedony and much limonite. Field estimate that some beds are 10 to 12 percent silica. Dolomite weathers buff brown and forms broken ledges along top of ridge. Some irregular masses or lenses half an inch or more thick of pure white quartz. Many small areas of pink-red coarsely crystalline dolomite or magnesite; irregularly distributed. Upper part of unit thin-bedded and tan on fresh fracture; contains veins of dolomite and dendrites of hematite	59

Section of Stensgar dolomite of Precambrian age, 4.7 miles northwest of Chewelah, Stevens County, Wash.—Continued

	Feet
7. Dolomite and argillite: gray and buff-tan finely to medium crystalline dolomite interbedded with tan and greenish-gray thin- and wavy-bedded argillite. Many veins and stringers, 1 to 3 mm thick, of chalcedony in most beds, and much limonite and some bands of hematitic clay. Upper part of unit platy bedded	65
6. Dolomite and argillite: pale-gray, finely crystalline, dense, hard, thin-bedded dolomite; stained buff and separated by parting beds of maroon and faint green-gray argillite. Upper half of unit contains less argillite and the dolomite is thicker bedded. Parting beds ⅛ to ½ in. thick between ½- to 4-in. beds of dolomite in lower half of unit gives wavy banded appearance to cliff faces. Unit weathers drab tan	90
5. Unit covered. Beds equivalent to this unit on the hill N. 40° E. are red, green, and buff argillite and interbedded buff-weathering dolomite	68
4. Dolomite, buff, stained pink in lower part, buff in upper part, coarsely crystalline, thick-bedded; contains veinlets and irregular lenses of cherty dolomite. Much limonite stain on some beds	17
3. Dolomite and argillite: dull maroon finely crystalline hard, dense dolomite in 4- to 24-in. beds in lower part of unit; grades up into pink medium crystalline hard thick-bedded dolomite which weathers drab buff and is wavy banded in upper 3 ft. Some irregular parting beds of purple-maroon and green argillite. Many irregularly distributed 1- to 2-mm veins of white quartz on joint surfaces. Dolomite slightly recrystallized. One 6- to 8-in. breccia zone along a joint face contains much limonite	18
2. Dolomite and argillite: darker purple-maroon massive dense finely crystalline dolomite and some parting beds of argillite. Unit banded red, buff, and faint green. Upper bed weathers drab tan	9
1. Dolomite and argillite: maroon dolomitic argillite interbedded irregularly with thin beds of dark- and light-green sericitic argillite and red and gray mottled crystalline siliceous dolomite. Dolomite increases in amount and thickness of beds up section, and some beds weather buff. Cherty veinlets and irregular veinlets common in upper beds. Unit forms small quarry face and weathers red, banded irregularly green and buff	34
Total thickness of Stensgar dolomite	730

It should be noted that this section is not typical of the Stensgar dolomite. The formation here is considerably thicker than is typical, and the zone of reddish dolomitic slate at its lower contact, with the McHale slate, is not found elsewhere at this horizon. In sec. 18, T. 32 N., R. 40 E., there is, within and near the top of the Stensgar, a local intercalation of red slaty dolomite, 400 feet thick at its greatest development, and rather similar lithologically to the basal phase to the north. Farther to the south, a red facies of the dolomite is well exposed in the steep walls of McHale Canyon. This is a relatively pure dolomite and lacks any slaty features. It probably is not more than 150 feet thick and is of limited lateral extent.

From McHale Canyon south as far as the Springdale-Hunters road, the Stensgar maintains (except in the parts that have been magnesitized) a rather uniform lithology: a fine-grained to almost aphanitic light-colored gray-weathering thin-bedded dolomite. Nodules of chert and secondary quartz veinlets are not uncommon. Exposures of rather typical Stensgar in this part of the belt are provided by cuts made for the road to Stensgar Peak lookout. A section here, in the NW¼ sec. 25, T. 31 N., R. 38 E., at an altitude of about 4,200 feet, was measured by C. F. Deiss. Here, as elsewhere, much of the section is covered; in fact, less than 5 percent of the Stensgar dolomite in the entire mapped area is actually exposed.

Section of Stensgar dolomite of Precambrian age, 14.5 miles west of Valley, Stevens County, Wash.

	Feet
Top of Stensgar covered.	
Dolomite, pale- and dark-gray, finely to microcrystalline, dense, hard, thick-bedded; weathers light gray and contains arenaceous lenses, 1/16- to ½-in. thick, which weather in relief	16
Unit largely covered. Dolomite in upper 15 ft similar to that in overlying unit	88
Dolomite, blue-gray in upper part, pale gray in lower part; cryptocrystalline, medium- to thick-bedded, finely and wavy-banded	15
Unit covered	125
Unit covered. Soil or float contains tan, red, and buff dolomitic argillite	258
Dolomite, pale-gray, finely crystalline. Contains 1- to 2-mm veinlets of white dolomite. Upper part of unit finely banded with salmon-colored clay laminae	8
Total thickness of Stensgar dolomite	510
Contact of Stensgar dolomite with underlying slates and argillites covered.	

From the Springdale-Hunters road southward, the Stensgar shows more extensive recrystallization, and much, though not all of it, might best be called a dolomitic marble. This facies is medium to coarse-grained, massive, white, and weathers white to gray.

From Huckleberry Creek northward to the southeast corner of sec. 7, T. 32 N., R. 40 E., the Stensgar is overlain, with possibly a small angular discordance, by a quartzite of the Buffalo Hump formation. From this point northward, it is overlain unconformably by members of the Huckleberry formation. From Huckleberry Creek south to the Spokane Indian Reservation, the Stensgar is overlain, with apparent conformity, by a slate of the Buffalo Hump formation. South of the Spokane Indian Reservation boundary, the Stensgar either pinches out due to nondeposition or is cut out by a small erosional unconformity, so that the slate of the Buffalo Hump formation immediately overlies the McHale slate. The Stensgar is thickest, 730 feet, at the northern end of the area where the top of the formation is concealed; at the southern end of the area, it thins and finally disappears. For much of the belt, the average thickness of the Stensgar is from 300 to 400 feet.

Locally the Stensgar dolomite changes—in some places gradually, in others rather abruptly—into magnesite, which appears to be a volume-for-volume replacement of dolomite. The magnesite is found at scattered localities along much of the belt, from the Finch quarry in the north to the Turk deposit in the south—a distance of almost 20 miles—and in masses ranging from a few pounds to millions of tons. Magnesite is harder and less soluble than dolomite and most of the more extensively magnesitized parts of the Stensgar dolomite therefore crop out as topographic highs (Finch, Allen-Moss, Keystone, Red Marble deposits). This feature was an obvious aid to prospecting, and it provides a basis for assuming that few, if any important exposed deposits within the area, are still unknown.

Many analyses of the Stensgar dolomite, especially from the magnesitized areas, are available (Bennett, 1941, p. 24–25). Those that are given in table 2 were selected to show the composition of widely scattered outcrops of the typical dolomite as well as of some representative magnesitized parts.

BUFFALO HUMP FORMATION

The rocks of the Deer Trail group that lie above the Stensgar dolomite are chiefly slates, quartzites, and closely related types. This part of the group was recognized by Weaver and mapped and described in more detail by Bennett (1941, p. 7), who differentiated an Upper quartzite overlain and underlain by "mainly undifferentiated slates." Because it proved impossible to map this Upper quartzite as a consistent unit, because other quartzites from which it is indistinguishable are present in this part of the section, and because all these quartzites seem to thicken or thin within short distances laterally and are thus unsuited for marker beds, it seemed best to lump together into a single formation all the rocks that overlie the Stensgar dolomite and underlie the Huckleberry formation. The characteristic assemblage of intermingled slates, argillites, and quartzites is well exposed on a prominent knob—locally called the Buffalo Hump—that rises steeply to the south above Huckleberry Creek in the SW¼ sec. 3, T. 31 N., R. 39 E., and over which passes the aerial tram from the Northwest Magnesite Co.'s mine at the Finch quarry to the Keystone and Red Marble quarries; we have therefore here defined the Buffalo Hump formation as the Upper quartzite of Bennett (1941, p. 7) as well as those "undifferentiated slates" he mapped above the Stensgar dolomite.

TABLE 2.—*Analyses of Stensgar dolomite from Stevens County, Wash.*

[Samples 1-4 collected by C.F. Deiss, analyzed by U.S. Bureau of Mines; sample 5 reported by Bennett (1941, p. 25, No. 64), analyzed by U.S. Bureau of Mines; samples 6-11 collected by Eugene Callaghan, analyzed by U.S. Geological Survey]

Sample	Ignition loss	Insoluble	R_2O_3	CaO	MgO
1	41. 3	10. 3	1. 8	26. 8	18. 7
2	43. 2	6. 1	2. 8	27. 9	19. 1
3	41. 7	9. 4	1. 9	27. 0	19. 3
4	42. 6	8. 1	2. 1	28. 3	19. 5
5	46. 0	2. 0	1. 6	27. 8	22. 0
6	49. 1	3. 3	. 8	. 6	44. 8
7	48. 6	2. 9	. 7	1. 6	45. 0
8	49. 9	1. 5	. 5	. 6	46. 2
9	45. 2	3. 4	. 9	. 7	46. 9
10	47. 5	3. 5	2. 1	2. 1	43. 3
11	49. 8	3. 5	7	. 2	45. 1

1. Dolomite from 156 ft above the base of the 730-ft section exposed in sec. 32, T. 33 N., R. 40 E., 4.7 miles northwest of Chewelah.
2. Dolomite from 163 ft below the top of the 730-ft section in sec. 32 T. 33 N., R. 40 E.
3. Dolomite from the Keystone quarry of the Northwest Magnesite Co., sec. 9, T. 31 N., R. 39 E.
4. Dolomite from the westernmost belt of Stensgar, in sec. 24, T. 31 N., R. 38 E.
5. Dolomite from the U.S. Magnesite deposit in sec. 10, T. 30 N., R. 38 E.
6. Magnesite from the Keystone quarry of the Northwest Magnesite Co., sec. 9, T. 31 N., R. 39 E.
7. Magnesite from the Double Eagle deposit, secs. 17 and 18, T. 31 N., R. 39 E.
8. Magnesite from the Midnight deposit, sec. 7, T. 31 N., R. 39 E.
9. Magnesite from the Turk deposit, sec. 1, T. 29 N., R. 37 E.
10. Fine-grained red magnesite from the Crosby deposit, sec. 18, T. 31 N., R. 39 E.
11. Magnesite from the Red Marble quarry of the Northwest Magnesite Co., sec. 25, T. 31 N., R. 38 E.

The Buffalo Hump formation has been traced from the bluffs south of Colville Valley in sec. 7, T. 32 N., R. 40 E., almost to the southern limits of the area. In the central part of the area, for example in the Red Marble section, the lower 300 to 400 feet of this formation is slate lying conformably on the Stensgar dolomite; except for its stratigraphic position, this slate is in no wise distinguishable from other slates of the Deer Trail group. It ranges in color from shades of light and dark gray to brown and green. It may be very thinly laminated but, on the whole, this feature is less common than in the McHale slate. At the "Slate quarry" on the road just below the Red Marble quarry, cleavage is exceptionally well developed; for the most part, however, the rock is more a slaty argillite than a true slate. Where both cleavage and bedding can be distinguished, their angular relationships differ as do the attitudes noted in the McHale slate. North of the Red Marble area this lower slaty facies of the Buffalo Hump formation thins rapidly, and north of the Mountain View it has not been found. To the south it thins slightly as far as the Spokane Indian Reservation boundary; but from this point south it thins more rapidly, being displaced by a thickening overlying quartzite.

Much of the Buffalo Hump formation is quartzite and quartzitic slate. To the north, the quartzite facies first appears south of Colville Valley, where it rests directly, and with slight unconformity, on the Stensgar dolomite. From a thickness of 500 feet in the McHale Canyon section, it thickens rapidly to almost 1,000 feet in the vicinity of the Allen-Moss quarry. From this point southward it thins (the slate described above appears between the quartzite and the Stensgar dolomite) until, in the Red Marble area, it is less than 200 feet thick. From here south, the quartzite thickens notably. In the central part of the belt, in the vicinity of the Springdale-Hunters road, this quartzite includes a lens of dark-gray slaty argillite which is 700 feet thick but which thins rapidly both to the north and south (pl. 2). About 1 mile south of the Springdale-Hunters road, the upper part of the quartzite is cut out by an intrusion of Loon Lake granite, which continues in approximately this horizon southward for at least 4½ miles. Despite this removal of its upper part, the quartzite continues to thicken southward; in the vicinity of the Spokane Indian Reservation boundary it is nearly 2,000 feet thick, and the top is concealed or removed. It should be noted here that on the basis of lithology and thickness the quartzite that is so well exposed, for example, in secs. 28 and 33, T. 29 N., R. 37 E., southwest of the Orazada mine, might be correlated with the younger Addy quartzite of Cambrian age, which unconformably overlies Precambrian rocks to the north. On the basis of stratigraphic and structural relations, however, we have correlated this southern, very thick unit with the upper quartzite of the Buffalo Hump formation, as Bennett did (1941, pl. 2).

In the central and north-central parts of the belt, the quartzite is typically gray white to yellow and brown—or in some places bluish—fine to coarse grained, and thoroughly indurated. In places it is pebbly and includes small lenses and stringers of quartz conglomerate. Thin slaty beds and lentils, too small to map, occur at many horizons. On the whole, the constituents of the quartzite are poorly sorted, and bedding and other depositional features are commonly so completely lacking that attitudes cannot be determined. Jointing and fracturing are abundant. To the south, as the thickness of the quartzite increases, the sorting is better and bedding planes are more conspicuous. The quartzite becomes medium to fine grained, pebble beds are scant or absent, and the color is white to light gray, and in places purplish.

From the Midnight area in the north to the Turk area in the south, where it is cut out by the Loon Lake granite, the uppermost part of the Buffalo Hump formation is a slaty argillite in which cleavage is developed to various degrees—for the most part only slightly, though some true slates were found. Fresh specimens are relatively dark gray or brown and commonly weather to lighter shades of tan or green. Banding (bedding?) is rare, but it may be represented by very thin faint lines of lighter color. The slaty argillite grades into the underlying quartzitic rocks; its top is cut by an erosional unconformity. Because this unit is overlain by the ledge-forming Huckleberry formation and underlain by rather resistant quartzite, outcrops are poor in quality and scant in number and no reliable estimate of its thickness can be given. South of the Turk deposit it is too thin to show on our map, but to the north it thickens rapidly to about 1,500 feet. This thickness is maintained almost as far north as sec. 22, T. 32 N., R. 39 E., where the slaty argillite is concealed by glacial drift. The slaty argillite is not found in the eastern fault block where the Finch and Allen-Moss quarries are located.

CONDITIONS OF SEDIMENTATION

The general conformity of the formations and the recurrence and similarity of lithologic types make it desirable to consider the Deer Trail rocks as a single group. As a whole, the group is characterized by the absence of coarse clastics. Fine-grained sand, silt, and dolomitic lime were the prevailing types of sediments, and silt was the predominant type. This assemblage indicates deposition in relatively quiet waters from an erosional area in which the stripping process was not very rapid. That the basin of deposition may have been relatively shallow is suggested by several ripple-marked quartzites, sun-cracked argillites, and thin red beds within the section. In all these features as well as in the considerable thickness of the section, the resemblance to the Belt series is striking; and of the various facies of the Belt, the resemblance is strongest, as might be expected, to the Purcell and Coeur d'Alene facies (Fenton and Fenton, 1937, p. 1877).

The question of marine versus nonmarine origin of the Belt has been discussed in many papers and is ably reviewed by the Fentons, who conclude that "with minor reservations, the series appears to be marine" (1937, p. 1938). On their map of the Belt sea (1937, fig. 13) the transition from marine to continental facies is shown as coming just within the boundaries of northeastern Washington. We would be inclined to correlate most of the Deer Trail group with the marine facies of the Belt and thus move the shore of the Belt sea a little farther west; in a sea as shallow as the Belt sea is supposed to have been, it is obvious that the line between marine and continental sediments cannot be closely drawn.

Within the magnesite belt both the proportion of clastic sediments and the size of particles increase toward the south. Thus, the two dolomite formations (Edna and Stensgar) both thicken northward, whereas quartzitic facies of the Deer Trail group thicken notably to the south. This circumstance may perhaps correlate with the observation of the Fentons (1937, p. 1940) that "the lands that furnished Belt sediments seem to have been higher in southwestern Idaho than in Washington, for they provided much greater quantities of both dissolved and suspended sediment during several epochs."

We have found no evidence of the rhythmic alternation of limestone and dolomite (Fenton and Fenton, 1937, p. 1931) which characterizes some of the calcareous facies of the Belt; but, if this feature was originally present, subsequent metamorphism might have obliterated it.

HUCKLEBERRY FORMATION

The Huckleberry formation contains two members, the lower one designated the conglomerate member, the upper one the greenstone member. These extend from the northern end of the magnesite belt southward to the vicinity of the Deer Trail Monitor mine. South of this point (in sec. 24, T. 30 N., R. 37 E.) a thin sliver of the conglomerate member was traced nearly to the Turk magnesite deposit. South of the Turk deposit, neither member was found.

CONGLOMERATE MEMBER

This conglomerate, first recognized by Jones (1928) and further studied and named by Bennett (1941, p. 8), is the most distinctive unit in the entire magnesite belt. A good section is exposed west of the topographic divide in sec. 28, T. 32 N., R. 38 E., and a much less complete, but more easily accessible section, may be observed along the road to Stensgar Peak. Typically, the rock is a schistose conglomerate composed of angular to subrounded fragments of slate, phyllite, argillite, limestone, dolomite, and quartzite. Fragments range from a fraction of an inch to several inches in diameter; but most are under a quarter of an inch. The angularity of the fragments is perhaps due more to the abundance of platy rock types, such as slate and phyllite, than to lack of wear. Quartzite and dolomite fragments are commonly well rounded, but they are scarce in most facies of the conglomerate. Sorting is poor, but alinement of fragments is striking, whether due to primary deposition or to subsequent stresses. The schistose appearance of the member is perhaps due more to alinement of fragments which themselves possess moderately good cleavage than to later metamorphism. The matrix of the conglomerate has a somewhat tuffaceous appearance, although alteration has made it impossible to determine what proportion of the matrix is of pyroclastic origin and what proportion is merely the finer grained part of the original sedimentary rocks. The greater part of the conglomerate is green or pale gray green. Differences in appearance are due principally to differences in the proportion, size, and shape of the larger fragments. The finer grained facies are difficult to distinguish from the underlying slate of the Buffalo Hump formation; hence, in many places the exact location of the contact is somewhat problematical. No consistent size gradation in the conglomerate was noted. Although coarse facies were found at or near the bottom, they seem to be equally common up to and above the middle of the section. There is some gradation toward fine sizes in the upper part, and the proportion of possible pyroclastic material appears to increase as the upper contact with the greenstone member is approached.

For about 16 miles in the central part of the belt, the conglomerate member of the Huckleberry maintains a fairly uniform thickness of about 1,500 feet, and there is little or no evidence of any angular unconformity with the underlying Deer Trail group. However, the conglomerate seems to be absent in the block in which the Finch and Allen-Moss quarries are located, and the upper or greenstone member of the Huckleberry rests directly on the Buffalo Hump formation. In the small area mapped north of Colville Valley, only a few feet of conglomerate was found; therefore an erosional unconformity probably occurs between the Deer Trail group and the Huckleberry formation.

The formations above the Deer Trail group were not studied in detail; hence, a discussion of the genesis of the conglomerate member of the Huckleberry is beyond the scope of this paper. The larger fragments in the conglomerate cannot be specifically identified as to source, but it seems very likely that they were derived by erosion of the underlying formations in the Deer Trail group. A puzzling feature is the predominance of slate, phyllite, and argillite over quartzite and dolomite. To be sure, this reflects the proportions of these types in the Deer Trail group, but one might expect erosion to increase the proportion of quartzite. The presence of dolomite pebbles suggests that somewhere during deposition of the conglomerate the Stensgar was being eroded, but if so, large masses of quartzite of the Buffalo Hump formation, which overlies the Stensgar, should also have been contributing detritus to the conglomerate, and quartzite pebbles are in many places conspicuous by their absence. It is possible, of course, that the larger fragments in the conglomerate came from formations still higher in the Deer Trail group, which may have included small beds of quartzite and dolomite now completely eroded.

GREENSTONE MEMBER

The greenstone member of the Huckleberry formation crops out in a continuous broad belt from the latitude of the Deer Trail Monitor mine northward to sec. 16, T. 32 N., R. 39 E., where it is cut out by granite. The greenstone, instead of following the crest of the Huckleberry Mountains, crosses from the west slopes at the southern end of the belt to the east side in the northern part of the belt. Another broad belt of greenstone lies west of the Finch quarry, and what is doubtless a continuation of this belt appears again north of Colville Valley. The greenstone is a resistant unit and outcrops are numerous. Good sections are exposed north of Colville Valley, in secs. 20 and 21, T. 32 N., R. 39 E.; east of Kline Farm in sec. 1, T. 31 N., R. 39 E.; and along the road to Stensgar Peak. Owing to lack of or uncertainty of attitudes in the greenstone member, no accurate figures for thickness can be given, but Bennett (1941, p. 8) estimates a maximum of 3,000 feet on Stensgar Mountain, which our mapping corroborates. The greenstone thins gradually to the north and to the south.

Although many lithologic types have been recognized, the greater part of the member apparently is composed of extremely fine grained and massive metabasalt. Differences in texture result in local varieties so dense that they may have originally been glassy; again, other varieties give evidence of having been originally medium-grained diabase. Amygdaloidal varieties are known but are not at all common. A relatively abundant variety characterized by numerous millimeter-sized ovoid concentrations of chlorite may have been originally finely vesicular. Other varieties with calcite-filled vesicles and sparsely disseminated chalcopyrite were noted. Toward the base of the greenstone member, where there is a gradational transition from the conglomerate member, pyroclastic varieties occur. These were also found, very scantily distributed, higher in the member.

None of these lithologic variants can be traced laterally, and flow units have not been recognized. Although some intrusive facies may be included, the great thickness of the greenstone, the fact that it succeeds a distinctly clastic unit, the presence of amygdules, the indications of vesicularity, and the very fine grain of most of the rock, strongly suggest that the greenstone member of the Huckleberry represents an accumulation of basaltic flows. Ellipsoidal structures have not been recognized. The greenstone is locally schistose, and these facies seemingly are more common toward the bottom and toward the top of the member. This may have resulted from greater shearing effects developed at the top and bottom during folding.

In the northern part of the area, the greenstone successively overlaps the upper units of the Deer Trail group (Stensgar dolomite, slate and quartzite of the Buffalo Hump formation) and the conglomerate member of the Huckleberry, thereby giving evidence of the unconformity that separates the Deer Trail group from the Huckleberry formation.

PALEOZOIC ROCKS

ADDY QUARTZITE

Resting with marked unconformity on the Huckleberry formation is a distinctive quartzite, the Addy quartzite of Early Cambrian age, first named by Weaver (1920, p. 61–63), who pointed out that it was one of the most extensively developed formations in Stevens County, underlying much of the crest of the Huckleberry Mountains as well as many other topographic highs. The name was retained by Bennett (1941, p. 9), who pointed out that "as originally described by Weaver the Addy quartzite included both the underlying greenstone and conglomerate." It is true that on Weaver's map the Huckleberry formation was apparently included with the Addy quartzite, but in his definition of the Addy, Weaver makes no mention of igneous or pyroclastic facies. Clearly, Weaver intended the name for a sedimentary unit.

Nowhere have we attempted to map a complete section of the Addy quartzite, but Bennett (1941, p. 9) states that its greatest thickness within the magnesite belt is between 3,000 and 5,000 feet. Good exposures, particularly of the lower part of the formation, occur on Stensgar Peak. Where exposed in the magnesite belt, the Addy is a thin-bedded, or, less commonly, a massive ledge-forming quartzite, bluish-gray to white, and thoroughly indurated. A distinctive phase is one in which bedding planes are emphasized by thin purple bands. Interbedded argillaceous layers and pebble lenses are scarce. No basal conglomerate is known within the area mapped, and fragments of underlying formations are notably absent. Where stratigraphic evidence is lacking, the Addy has been distinguished from the quartzites of the Deer Trail group on the basis of its greater uniformity, its lesser degree of iron staining, its more distinctive bedding, and especially by its much greater thickness; exceptions have been indicated in the discussion of the correlation of the thick quartzite in the southern end of the belt (p. F-19).

A quartzite which has been correlated with the Addy on the basis of these criteria forms the western slopes and the crest of the prominent mountain east of Finch quarry, in secs. 20 and 21, T. 32 N., R. 40 E. The same quartzite appears north of Colville Valley, where it has been quarried as a source of silica for the manufacture of ferrosilicon. Because this quartzite is correlated with the Addy, a thrust fault has been mapped along its western border.

Although the upper part of the Addy was not mapped in our work, it is of interest to note that Bennett (1941, p. 9) reports it to be overlain conformably by extensions of the Old Dominion limestone of Weaver (1920, p. 66–68), a formation in which the Lower Cambrian brachiopod *Kutoringa cingulata* (Billings) has been found. It is for this reason that the Addy was first considered to be Cambrian. Subsequently, Okulitch (1951), at the type locality of the Addy quartzite, obtained fossil evidence which definitely places this formation in a horizon "very close to the base of the Lower Cambrian."

TERTIARY SYSTEM

GEROME ANDESITE

In the extreme southwestern part of the area, the rocks of the Deer Trail group disappear beneath flows of andesite. Weaver, who named this overlying formation the Gerome andesite (1920, p. 98–99), stated that it "consists of lava flows, tuffs, breccias, and intercalated beds of shale and sandstone containing impure carbonaceous seams of varying thickness. The total thickness of these igneous and sedimentary members is at least 700 feet." Bennett shows the formation on his map (1941, pl. 2), but does not mention it in his text. We have not studied the formation in any detail because the only facies present in the area mapped is a distinctive microporphyritic hornblende andesite. This rock is composed of a fine-grained gray matrix studded with tiny, very fresh, and often markedly euhedral, black hornblende phenocrysts. It commonly shows a flow texture sufficiently well developed that attitudes can be mapped. The Gerome andesite of Oligocene age lies with marked angular unconformity upon the formations of the Deer Trail group.

INTRUSIVE ROCKS

GREENSTONE DIKES AND SILLS

Throughout much of the area covered by rocks of the Deer Trail group, dikes and sills of greenstone occur. Only a few of the larger ones can be mapped to scale. Similarity in original composition and in the type of alteration (dominant chloritization, with minor epidotization, uralitization, and serpentinization) suggests that these intrusives may have been feeders for the flows of greenstone of the Huckleberry formation. This is further suggested by the relative scarcity of the dikes in the southern part of the area, where the greenstone member of the Huckleberry is absent. Finally, these greenstone instrusives are not known to penetrate any rocks above the Huckleberry; in fact, any occurrences above the Deer Trail group are very dubious. Within the rocks of the Deer Trail group, the greenstone intrusives seem to be most abundant at about the horizon of the Edna dolomite; they are less common in the quartzites, but occurrences are known in all formations of the Deer Trail group. Various exomorphic effects of these intrusives are discussed under "Metamorphism."

The smaller bodies especially and the walls of larger bodies of greenstone exhibit considerable shearing and are locally schistose, but this textural feature differs widely in both degree and distribution. Where structural or stratigraphic relationships are not apparent, it is not always clear whether isolated outcrops of greenstone should be assigned to the intrusive or to the Huckleberry formation. In the NE¼ sec. 1, T. 29 N., R. 37 E., we noted a small unweathered mafic dike with chilled selvages comparable to fresh basalt.

LOON LAKE GRANITE

Weaver (1920, p. 87–89) proposed the name Loon Lake granite for a granitic batholith that occurs in the eastern part of Stevens County, Wash. Near the Spokane Indian Reservation boundary (pl. 1) granitic rock of batholithic dimensions transects the metasedimentary rocks of the magnesite belt; nearby isolated masses of granitic rock are assumed to be apophyses of the same body. Although a very large part of this rock is quartz monzonite or granodiorite rather than granite, we have used Weaver's terminology in mapping the area. Within the area studied, the rock is commonly light gray to light pink with a sprinkling of lustrous black biotite flakes; less commonly it contains small amounts of hornblende. Generally it is medium to coarse grained and even grained, though in some outcrops it is strikingly porphyritic, containing phenocrysts of orthoclase up to 2 inches in length.

Within the area studied, the Loon Lake granite is less resistant than most of the formations with which it comes in contact, so that exposures are poor; but from the areal pattern it is apparent that the intrusion has been both discordant and concordant.

INTRUSIVE RHYOLITE

Numerous small bodies of rhyolite and rhyolite-porphyry, all but one of which are too small to represent on the map, were first noted in the vicinity of the Wells-Fargo mine and were traced southward from there almost to the southern end of the belt. The largest body underlies a small hill half a mile south of the Orazada mine. The distribution of most of the rhyolite suggests that it was intruded in sill-like bodies, nearly all of which are confined to the lower part of the Deer Trail group (Edna dolomite and Togo formation). Except for one doubtful outcrop along the Stensgar Peak road, rhyolite has not been found north of the Wells-Fargo mine. The rock is weathered to various degrees and is commonly pink to light brown. Sparse euhedral crystals, 1-mm long, of quartz and, in the fresher specimens, of feldspar are set in an aphanitic groundmass that in places is faintly flow banded. There is no question, however, concerning the intrusive nature of the rhyolite. Like the Loon Lake granite, this rock shows no evidence of the metamorphism which affected the rocks of the Deer Trail group. It seems reasonable to correlate the rhyolite with the igneous activity associated with the emplacement of the Loon Lake batholith.

AGE AND CORRELATION OF FORMATIONS

Although it is impossible, due to the absence of fossils in any of the formations studied, to make specific age determinations for the formations of the magnesite belt, nevertheless there are some obvious analogies in lithology and stratigraphy that deserve to be mentioned.

Most workers in the area agree that the granitic intrusions are of Mesozoic age, and lead-alpha age determinations on zircon obtained from granitic rocks in the Turtle Lake quadrangle south of the mapped area indicate that they are Cretaceous in age (George Becraft, written communication, 1958). There has been uncertainty and disagreement in the dating of the older formations. Weaver (1920, p. 50), Jones (1928), and others, some with doubts and reservations, placed the rocks of the Deer Trail group in the Paleozoic; Bennett (1941, p. 8) and others have placed the group in the Precambrian. More recent work in the Hunters and Turtle Lake quadrangles confirms the Precambrian age (George Becraft, written communication, 1958).

In the Metaline quadrangle, which lies just northeast of the Chewelah quadrangle, the oldest formation recognized by Park and Cannon (1943, p. 6) is the Priest River group, described as a "complex sequence of metamorphic rocks that includes phyllites and schists, limestones, dolomites, quartzites and volcanics." The rocks are predominantly fine-grained phyllites, and Park and Cannon say "it is somewhat surprising to find metamorphism of so low a grade in rocks that had so complex a geological history as these." They assign the Priest River rocks to the Precambrian. Unconformably overlying the Priest River group in the Metaline quadrangle is a distinctive conglomerate, which Park and Cannon (1943, p. 7) name the Shedroof conglomerate. This conglomerate is overlain by the Leola volcanics (op. cit., p. 9), consisting mainly of homogeneous greenstone. The Shedroof conglomerate and the Leola volcanics are assigned by Park and Cannon to the top of their Precambrian section (op. cit., p. 11). A doubtful unconformity separates these rocks from the overlying Monk formation of Cambrian (?) age, a varied group of phyllites, quartzites, grits, and limestone, which grades upward into the Gypsy quartzite, known to be Cambrian in age (op. cit., p. 13).

In position, lithology, and degree of metamorphism the rocks of the Deer Trail group correspond closely to the Priest River group; in the magnesite belt, the Deer Trail group is overlain by the conglomerate member of the Huckleberry formation (analogous to the Shedroof conglomerate) and the greenstone member of the Huckleberry (analogous to the Leola volcanics). No formation corresponding to the Monk formation is known in the magnesite belt, but the Gypsy quartzite seems to be analogous to the Addy quartzite. On this basis, it seems reasonable to suggest that both the Deer Trail group and the Huckleberry formation are Precambrian; that the break at the top of the Huckleberry represents the break between Belt time and Paleozoic; and that the Addy quartzite represents the basal Cambrian in this area, as confirmed by Lower Cambrian fossils found in the Addy nearby (Okulitch, 1951). In western Montana, where the relations between Belt and Cambrian have received much more detailed study (Deiss, 1935), it is significant that, as in the Metaline and Chewelah quadrangles, the unconformity between Belt and Cambrian is nowhere striking and may even pass unnoticed, and that everywhere the basal Cambrian is represented by a siliceous sandstone or quartzite.

METAMORPHISM

At least three types of metamorphism, exclusive of minor hydrothermal changes, can be recognized in the rocks of the magnesite belt. These are (a) metamorphism associated with intrusions of greenstone dikes and sills, (b) metamorphism associated with folding, and (c) metamorphism associated with granitic intrusions.

Metamorphism associated with the greenstone dikes and sills is distinctly local and generally minor. Where the dikes have intruded quartzite, some knife-edge contacts are exposed and the quartzite commonly is unchanged. The slates, argillites, and phyllites are also affected only slightly (coarser recrystallization) or not at all. In the Stensgar dolomite, however, a zone of dense flinty rock with a greenish hue, ranging from a few inches to several feet, is developed adjacent to the intrusions. In these zones, minerals of the serpentine group are prevalent. Good exposures of such a zone occur in the gulch south of the Keystone quarry.

The Edna dolomite, whether due to its greater content of impurities or to the presence of more abundant and larger greenstone intrusions, is the formation in which contact effects are most pronounced, though they may be wholly lacking in places. The commonest result is recrystallization to a coarse marble that locally contains a small amount of tremolite or talc. More striking evidence of the susceptibility of the Edna dolomite to intrusive effects lies in the copper mineralization, which consists chiefly of pyrite, chalcopyrite, bornite, and tetrahedrite, in a gangue of quartz and coarsely crystalline carbonate. Copper prospects dot almost the entire length of the Edna dolomite, and, significantly, almost every prospect is near a greenstone intrusive.

All the rocks of the magnesite belt older than the Loon Lake granite have been slightly metamorphosed during folding, so that cleavage or schistosity has developed and sericite and lesser amounts of talc, chlorite, and actinolite have been formed. The effect of shearing stress on different beds differs greatly. The fine-grained incompetent argillaceous rocks are the most intensely sheared, and locally some near-commercial quality slate has formed. More commonly, cleavage has not developed beyond that of a slaty argillite. In some localities phyllitic rocks have formed. The quartzites show little or no effect of the shearing and for the most part seem to have reacted as massive competent units. Interbedded argillaceous layers within the quartzites, on the other hand, commonly show very considerable shearing. Where relatively pure, the dolomites behave much as the quartzites, that is, as massive units scarcely affected by shearing. Where somewhat argillaceous, however, they may be converted to talcose schists or calc-phyllites.

Shearing and orientation within the conglomerate member of the Huckelberry formation has already been mentioned (p. F–21). Owing to the original platy habit of many of the fragments, no reliable measure of elongation could be made. However, dolomite pebbles in which the longest axis is four to five times the length of the least axis, have been found. These examples are exceptions, however, and we believe that most components of this formation have undergone no significant elongation.

Shearing near the top and bottom of the greenstone member of the Huckleberry has been mentioned. In general, the larger masses of greenstone (whether intrusive or extrusive) rarely show any effects of shear, but smaller bodies, and in places the contact surfaces of the larger bodies, may be extensively sheared and locally may resemble chlorite-actinolite schists.

Despite the evidence of dynamic metamorphism, the general impression to be gained from all these rocks, especially those of the Deer Trail group, is that they are only moderately metamorphosed, considering their Precambrian age and the folding and thrusting they have undergone. On the whole, all these rocks fall clearly within the greenschist facies as defined by Turner and Verhoogen (1951, p. 465–466).

As might be expected, the highest grades of metamorphism are in the immediate vicinities of granitic intrusions. The zones of these "contact"-metamorphic effects are narrow despite the size of some granitic bodies. In the slaty and argillaceous rocks, metamorphism associated with a granite intrusion commonly produced a fine-grained sericite-biotite-cordierite hornfels. At only one place (sec. 22, T. 39 N., R. 32 E.) was sillimanite found. In some localities bedding and cleavage are entirely obliterated. On the quartzites, granitic intrusions have had relatively slight effect. In some of the less pure quartzites, metacrysts of feldspar formed close to contacts, and in exceptional cases a "granitized" facies has developed.

The conglomerate member of the Huckleberry has been changed strikingly in the vicinity of granite contacts. Near the southern end of the conglomerate it is converted into a rock with the aspect of a medium-grained diorite, consisting of small amphibole metacrysts studding a greenish-gray matrix of recrystallized quartz, sericite, and diopside. In other places, the conglomerate has been converted into a fine-grained dark-green amphibole hornfels from which all trace of schistosity and bedding has been obliterated.

The greenstone member of the Huckleberry is only slightly affected by granitic intrusions. In the southern part of the unit the greenstone is recrystallized to a medium-grained rock of gabbroic appearance, but the minerals are virtually the same as those in the parent rock.

In few places have contact effects of the granite on the Stensgar dolomite resulted in more than a thorough recrystallization. At the Turk deposit, for example, the generally coarser grain, marmorized appearance, and possibly the bleaching of the Stensgar, may be contact effects of a nearby but hidden granite mass. Silicate minerals of metamorphic origin were found in the Stensgar, but they are not particularly common. In thinner bedded and possibly more argillaceous facies, talc and minerals of the serpentine group are not uncommon, but these may have resulted more from stress than from contact metamorphism. An example of contact effects occurs in the dolomite adjoining the small intrusion of granite near the Allen-Moss quarry where, in a zone from a few inches to a few feet thick, the

Stensgar has been converted to a medium-grained mass of forsterite, calcite, and serpentine. The Queen and Seal mine, the Cleveland mine, and other lesser deposits and prospects are located in the southern half of the Stensgar dolomite belt, and it seems not unlikely that these epithermal veins and replacement bodies containing silver, lead, and zinc minerals are related to intrusions of Loon Lake granite. Only in the Cleveland mine has any detailed study of the mineralogy been reported (Jenkins, 1924, p. 16–17), and therefore it is not known whether any significant differences exist between the metalliferous deposits formed in the Stensgar dolomite and those, such as at the Turk and Deer Trail mines, formed in the Edna dolomite. Certainly there appears to be some contrast between the lead-silver deposits that seem to be genetically related to the younger granite intrusives and the copper deposits that seem to be genetically related to the older greenstone intrusives.

The most conspicuous metamorphic effects of granite intrusions occur in the Edna dolomite, which is converted to a fine-grained dark-green diopside hornfels. Commonly this rock is massive, but not uncommonly it may show dark layers of brownish-green diopside, alternating with light layers of greenish-gray diopside and dolomite. Whether these bands represent original compositional differences or whether they are due to metamorphism is not known, for outcrops of these types are very limited in extent and structural relations are not evident. A less common type of metamorphosed Edna dolomite is found near the border of the Spokane Indian Reservation (NE¼ sec. 22, T. 29 N., R. 37 E.). This consists of green actinolite crystals up to half an inch long, arranged in decussate pattern in a fine-grained matrix of quartz-dolomite hornfels.

Metamorphic changes obviously add another obstacle to attempts to work out the details of stratigraphy. This is particularly well illustrated by the Edna dolomite, in which the more considerable metamorphic changes have produced a rock that, unless its origin could be traced, would never be correlated with the Edna. This explains the principal difference between our map and that of Bennett (1941). Bennett does not show the Edna dolomite south of the Springdale-Hunters highway, whereas we have projected it almost as far as the Spokane River. It is in this region that the largest invasion of granite has occurred, and it is here that the Edna shows the greatest changes, some of it being entirely a diopside hornfels, with little or no trace of carbonate left. Mapping the Edna as far south as we did enabled us to reinterpret the carbonate beds lying higher in the section in the southern part of the belt. Thus, in our interpretation the dolomite that extends almost to the southern boundary of the map is the Edna and not the Stensgar, as on Bennett's map, and the Stensgar ends at about the latitude of the Orazada mine.

An estimate of the stage of metamorphic intensity reached in contact aureoles has been a matter of interest, particularly since the systematization of Harker (1932). In the aureoles surrounding the granite intrusions in the magnesite belt, the metamorphism of the argillaceous beds of course provides the most sensitive measure. With only rare exceptions, we found no moderate- to high-intensity minerals such as garnet, staurolite, and sillimanite. The principal neocrystallization minerals are biotite and cordierite, indicating that a rather low-intensity metamorphism prevailed. This is further borne out by the uniformily fine grain of most of the hornfelses. In a few places an approach to knotenschiefer texture was observed, but more commonly all constituents are of even and very fine grain. At least within the magnesite belt, the contact effects of the Loon Lake granite, though striking, represent no more than a low-intensity stage of thermal metamorphism.

STRUCTURAL GEOLOGY

The dominant structural element of the magnesite belt is the west limb of a rather closely folded and locally overturned anticline, whose axial trend is north-northeast. This was first recognized in broad outline by Weaver (1920, p. 108) and was confirmed by Bennett (1941, p. 10). Our work has served chiefly to bring out in greater detail the many complexities, offsets, faults, smaller folds, and overturns that exist in this broad structure.

FOLDS

At the southern end of the magnesite belt the formations are nearly vertical or are slightly overturned to the west. To the north, dips are commonly westward at angles ranging from shallow to steep but averaging about 50° Subsidiary folds increase in number to the north and appear to be drag folds, some of which—as the one in the vicinity of the Double Eagle and Keystone deposits—are thousands of feet across, whereas others, chiefly in the slates, duplicate the large drag folds on smaller scales, some being less than an inch wide. In the northern part of the area—in secs. 7, 18, and 19, T. 32 N., R. 40 E., and in secs. 13 and 24, T. 32 N., R. 39 E.—a series of northward-trending tightly compressed mostly isoclinal folds are overturned to the east and involve all the formations of the Deer Trail group. Dips there, both normal and overturned, range from 35° to 65° W. The plunge is northerly and probably gentle, as indicated by the long outcrop areas of a quartzite of the Buffalo Hump formation and a slate of the Stensgar dolomite, along the axial parts of the two most westerly folds. Farther to the north, all the folds merge into one anticline, which plunges northward.

FAULTS

The major faults shown on the geologic map are for the most part parallel to the regional strike of the beds. Traces of the faults (as nearly as these can be approximated) cut across the topography in most instances in nearly straight lines, suggesting vertical or nearly vertical attitudes. Nowhere is the plane of any of the larger faults actually exposed, and every fault is therefore inferred from stratigraphic relations, which, in turn, are often far from clear.

The westernmost large longitudinal fault or fault zone is best shown in the Red Marble area, where there is a repetition of a belt of Stensgar dolomite that includes the Red Marble, Crosby, and other smaller magnesite deposits. In the northwest corner of sec. 36, T. 31 N., R. 38 E., near the Wells-Fargo mine, this fault zone is displaced by a tear fault having a horizontal displacement of at least 800 feet. From there the zone continues southwestward as far as the Turk deposit, where it joins, or is joined by, a smaller diagonal fault coming in from the north; it continues southward beyond the Queen and Seal mine, ending against granite. The total length of this fault zone is more than 15 miles. In the vicinity of the Crosby and Red Marble deposits, where the existence of the zone is best demonstrated, the stratigraphic displacement is more than 1,100 feet.

A second longitudinal fault or fault zone which also repeats the Stensgar dolomite and associated Deer Trail rocks lies east of the one just described. This second zone has a stratigraphic displacement, in the vicinity of the Crosby and Double Eagle deposits, of at least 1,500 feet. South of the tear fault near the Wells-Fargo mine, it either splits into two longitudinal faults or may be represented by either one of the mapped faults which continue southward, with various complications, beyond the Springdale-Hunters road. North of the Double Eagle deposit this fault appears to be folded in approximate conformity with the bedding. It passes northwest of the Keystone quarry where it may be related to other faults which cause the Stensgar dolomite to be repeated four times. This same fault zone and the one first described may continue north of Huckleberry Creek, where repetition of parts of the section indicates additional faulting; but the considerable blanket of glacial deposits and Recent alluvium in the valley of Huckleberry Creek effectively obscure bedrock relations. Inasmuch as neither stratigraphy nor structure can be closely matched across Huckleberry Creek, it is probable that there has been some transverse faulting in this area. East of Browns Lake and passing north across Colville Valley is another large longitudinal fault zone which has brought Addy quartzite adjacent to Edna dolomite, indicating a stratigraphic displacement of 10,000 to 15,000 feet.

Evidence of normal faulting is found chiefly within the quarries and mines. The displacements are so small that the faults cannot be traced for any distance or represented on the areal map. It is likely that most of the normal faults resulted from the small adjustments made necessary by the major thrusting, which, in turn, may have been related to the folding and overturning to which the Deer Trail rocks have been subjected.

The time of the folding and faulting cannot be determined except that both must have occurred after the formation of the Addy quartzite and probably largely before the granitic intrusion, for some faults seem to stop at granite contacts and the granite shows no evidence of shearing. Presumably the folds and faults in the magnesite belt originated in the period just preceding the batholithic intrusions. The relation between faults and folds is not so clear as might be wished. The transverse faults, as, for example, the tear south of the Wells-Fargo mine, are clearly later than the folding and the longitudinal faults. The longitudinal faults might be expected to be slightly later than, or even partly coincidental with, the folding; yet nowhere is there a clear-cut example of a longitudinal fault cutting or even passing into a fold, and there is good reason to believe that one longitudinal fault has been folded (the one inferred just north of the Double Eagle magnesite deposit).

SURFACE FEATURES

QUATERNARY DEPOSITS

Problems of glaciation involving the general region are discussed in papers by Flint (1936; 1937) and Hobbs (1943). Within the magnesite belt, Pleistocene glaciation was responsible for notable drainage modifications and has left a mantle of glacial debris on the lower slopes of the Huckleberry Mountains. The higher parts of the Huckleberry Mountains appear to be unglaciated, and the ground moraine is believed to terminate to the south at about the latitude of the Keystone quarry. South of this point, therefore, mantle rock may be accepted as a reasonable guide to the character of the underlying bedrock; to the north, due allowance must be made for the vagaries of glacial distribution. In places quartzites and, to an even greater extent, greenstones of the bedrock formations show glacial pavements and glacial striae; good examples are on the quartzite mountain east of Finch quarry and on the greenstone exposures north of Fourmile Lake and in the N½ sec. 33, T. 32 N., R. 39 E. Striae commonly indicate that the ice moved in a direction around S. 20° W.

The Keystone quarry is located on the steep slope of an eastward-facing ridge, which, at the quarry site and also to the north, shows recent landslide scars. The hummocky surface of the landslide blocks cannot readily be differentiated from the hummocky surface produced by the glacial ground moraine, and the outline of the landslide area is therefore somewhat conjectural. Possibly between 50 and 100 acres may be involved, lying just northeast of Keystone quarry. The over-steepening which caused the landslides probably resulted from glacial action, as there is no indication of structural control.

Glaciofluvial deposits and Recent alluvium fill the valleys of the larger streams, such as Colville River and Huckleberry, Woodbury, Deer, and Hunters Creeks. Except that they conceal the character and distribution of bedrock, neither these nor the glacial deposits have significance with respect to the magnesite occurrences; accordingly, all are represented by a single pattern on the areal map.

DEVELOPMENT OF TOPOGRAPHY

There is much evidence of glaciation in the drainage of the area. The most obvious features are the many lakes and the flat, in places U-shaped, trench of the Colville River. A probable case of piracy is in the southwest fork of Huckleberry Creek and the north fork of Deer Creek. The major courses of several streams—Stranger Creek, the two forks of Chewelah Creek, Woodbury Creek, Huckleberry Creek, Deer Creek, and others—trend southeast or southwest, which indicates that the master stream—the Colville River—may at one time have flowed southward rather than northward, as at present. Indeed, C. D. Campbell (1953, p. 138) has suggested that for a time during the glacial period the Columbia River took a southward course in what is now Colville Valley.

The dominant topographic feature of the area is the Huckleberry Mountains, much of which stood higher than the limits of glaciation (Flint, 1937, pl. 5). Glaciation, however, by shaping the valleys, also determined in part the broad outlines of the ridges, and differential erosion played only a minor part in shaping the landforms. The major ridges are not strike ridges, but diverge slightly from the strike of the beds. The formations of the magnesite belt, for example, commonly strike N. 15° to 20° E., whereas the ridges carved from them trend only a few degrees east of north. The best illustration of this divergence is the Huckleberry Mountains. From a glance, one might infer that this is a strike ridge of the Addy quartzite. The highest part of the Huckleberry Mountains does indeed coincide with an outcrop of Addy quartzite; but, if one follows the crestline southward from Stensgar Peak, he soon passes from the Addy quartzite to the greenstone member of the Huckleberry formation. On continuing southward, one finds successively older formations: the conglomerate member of the Huckleberry, the Buffalo Hump formation, the Stensgar dolomite (which crosses the crest near the Cleveland mine), and so on, until in the southern part of the belt he reaches the quartzites of the Togo formation, which make a conspicuous crest south of the Orazada mine. This subparallelism rather than true parallelism between topography and stratigraphy occurs even in rather small and local structures and must constantly be considered when attempts are made to correlate geology and topography.

On the whole, the landforms of the region are in a mature stage of erosion except in the larger alluviated valleys and glaciated parts, where recent downcutting has produced youthful topography locally.

MAGNESITE

The outstanding mineral resource of the region has been, of course, the magnesite. Although this report is concerned with the geology of the belt as a whole rather than with the details of the magnesite deposits, it is desirable to point out here some relationships between the stratigraphy and the occurrence of the magnesite and to include such information about the deposits as we have been able to gather.

OCCURRENCE AND ORIGIN

Search that began about 1900 has failed to reveal a single occurrence of magnesite outside the Stensgar dolomite; but within the Stensgar, distribution of the magnesite is variable, in both place and amount. Zones containing perhaps not more than a few hundred pounds of magnesite (as in the E½ sec. 17, T. 31 N., R. 39 E) contrast with deposits such as the Red Marble, where several million tons occur. Although the major deposits lie in the north-central part of the belt, two deposits (U.S. Magnesite and Turk magnesite) are known in the southern half of the belt. Whether the magnesite occurs at a consistent horizon or several horizons within the Stensgar cannot be determined. For example, although the magnesite appears to occur in the upper parts of the Stensgar dolomite at the Finch and Allen-Moss deposits (northern end of the belt), is seems to be in a middle part at the Red Marble deposit, some 7 miles farther south. Yet, because the Stensgar thins to the south, the magnesite may lie at virtually the same time horizon at both locations.[3]

[3] We are indebted to Mr. Conrad Martin, geologist for the Basic Refractories Co., for first suggesting this possibility to us.

An obvious puzzle is why there has been no magnesitization of the Edna dolomite, particularly if—as will be emphasized below—we accept a hypogene source for the solutions that effected the replacement. But the chemistry and the mechanics of replacement are still much too imperfectly known to explain this seeming anomaly. Schroeder (1948), in his study of the Turk deposit, pointed out that the less pure parts of the Stensgar were in general not magnesitized, whereas those parts (chiefly the middle part of the section) which were rather pure dolomite were most extensively replaced. If degree of purity of dolomite be a factor in determining susceptibility to magnesitization, then the Edna dolomite would be in general a far less favorable host rock than the Stensgar. Detailed drill-hole data and chemical analyses, comparable to those published for the Turk deposit, are not available for other sections of the Stensgar or for the Edna. The few chemical analyses we have of the Edna (table 1) show an average insoluble of 26.2 percent, whereas the analyses of the Stensgar (table 2) show an average insoluble of only 4.9 percent. These figures illustrate the very impure character of the Edna, and thus may serve to substantiate the suggestion that relative purity of the carbonate formation has a part in determining susceptibility to magnesitization.

The problems that bear on the origin of the magnesite have been ably reviewed by Culver (1939, p. 19–23) and further summarized by Bennett (1941, p. 13). The confinement of the magnesite to a single formation might seem to suggest a primary depositional origin—an hypothesis which was supported by Whitwell and Patty (1921) and Siegfus (1927). However, the erratic distribution of magnesite in the Stensgar, the large metacrysts of magnesite in the dolomite, and many other equally clear-cut replacement features lead us to agree with those who have advocated a replacement origin for the magnesite: Weaver (1920, p. 322), Jenkins (1918), Bain (1924, p. 422), Willis (1927).

A fact bearing on the time of magnesitization is that the magnesite appears not to have been affected by the regional deformation (folding and major faulting), though it is affected by the presumably later small normal faults. This gives strong support to the concept that the replacement took place about the time of the granitic intrusions and may therefore have been effected by hydrothermal solutions connected with the granite magma, instead of being related to the much earlier greenstone intrusions, as some have postulated (Jenkins, 1918; Yale and Stone, 1921; and Bain, 1924). Bennett (1941, p. 14; 1943, p. 8) has suggested a connection between the batholiths and the magnesite, and it is of interest to note that in the only other large deposit of "crystalline" magnesite in the United States (the deposit at

Gabbs, Nev.), Vitaliano and Callaghan (1956) have concluded that the replacement of dolomite by magnesite was accomplished by solutions moving upward through the dolomite and genetically related to an earlier granitic intrusive.

Assuming that magnesite is a hydrothermal replacement of dolomite, there remains the question of its localization. Willis (1927) stressed the importance of overthrust beds of quartzite, which in his opinion acted as barriers to hypogene solutions in the Stensgar. Although quartzite is conspicuous on the hanging wall of the Stensgar at the Finch and Allen-Moss deposits, it is not present at all magnesite occurrences, and in any event, we doubt the overthrust relationship.

The fact is that we have not found any consistent feature in either structure or lithology that would explain the localization of the magnesite. Presumably this must in part have been controlled by throughgoing channelways (joints or faults) which reached to the source of the magnesium-rich solutions, but we have no knowledge of the local factors controlling the process. On the nature and source of the magnesitizing solutions, our study throws little light. Bennett (1943) called attention to the somewhat unusual association of forsterite with magnesite at the Turk deposit, and Schroeder (1948), pointing to the presence, also at the Turk deposit, of magnetite and pyrrhotite, concluded that temperatures of the magnesitizing fluids may have been sufficiently high to have come within a hypothermal range. Bodenlos (1954) has recently reviewed the information bearing upon formation of "crystalline magnesite" in general, and concludes that these deposits result from hypogene solutions to which, in the case with which he was concerned, he attributed mesothermal characteristics. He, like Faust (1949) and Faust and Callaghan (1948), accepts dolomite as a logical source (by dedolomitization or other mechanism) for the magnesium in the magnesium-rich solutions, in those places where dolomite can be shown to be present in the area. These generalizations are of specific application to the Stevens County deposits. Although we have found no evidence of dedolomitization within the area studied, the granitic intrusions might well have dedolomitized either the Edna dolomite or the Stensgar dolomite, or both, at greater depths where higher temperatures must have prevailed. Given this source of magnesium-rich solutions, the precipitation of the magnesite where now found was doubtless determined by one or more of those subtle factors which still elude the geologist seeking to explain the localization of replacement deposits.

HISTORY AND PRODUCTION

The early history of the magnesite industry in Stevens County is given by Weaver (1920) and Whitwell and Patty (1921). Despite the demand for domestic sources of refractory-grade magnesite precipitated by World War I, which led to the discovery and initial development of the Stevens County deposits, more companies failed than succeeded in the half decade following 1916, when the first shipments were made. In that year the Washington Magnesite Co. produced 715 tons of crude magnesite. In the following year, 1917, this company was taken over by the Northwest Magnesite Co., which since then has remained in nearly continuous production and has been the only producer of refractory magnesite in the district. In 1923 the Northwest Magnesite Co. bought the Red Marble, Allen-Moss, and Woodbury deposits from the American Mineral Production Co., which had ceased production in 1920. Today the Northwest Magnesite Co. owns all the deposits of any significant tonnage in the magnesite belt, except the Turk deposit, which is on State land.

Early operations involved selective quarrying and hand sorting of the ore to insure a sufficienctly low content of silica and dolomite, the two chief gangue minerals. Until 1943 about 50 percent of all of the quarried rock was thus discarded as waste. Although in some quarries the magnesite has a color (red or black) that permits ready discrimination from dolomite (generally light gray), in the Finch and Allen-Moss deposits the color of the magnesite and of the associated dolomite may be so similar as to make distinction difficult. Instead of color, quarrymen made use of the difference in luster of the two minerals, a feature best seen in sunlight, and largely dependent on the difference in the index of refraction (in magnesite, $n_0=1.700$ to 1.726; in dolomite, $n_0=1.680$ to 1.716). This is the only instance known to us where difference in the index of refraction has been of commercial application in separating two minerals otherwise so similar in appearance.

In 1941 the Northwest Magnesite Co. introduced a two-stage flotation process for removal of silica and dolomite to upgrade their product still further. In 1943 the company established a sink-float process which took advantage of the relatively coarse grain size of the quarried material. Feed to the sink-float plant requires crushing and sizing to only 1/8 of an inch to 1 1/2 inches. The density of the heavy medium (a ferrosilicon suspension in water) is set at 2.87. The company deserves much credit for having pioneered, in wartime, an application of sink-float separation to two nonmetallic minerals, dolomite and magnesite, of such small difference in specific gravity (dolomite, 2.85; magnesite, 3.00). The success of this operation (Wicken, 1945) made it possible to abandon selective mining and the relatively inefficient hand sorting of the quarried rock, and indeed, has made it possible, under favorable circumstances, to exploit some of the old "waste" dumps.

The great bulk of the magnesite produced in Stevens County has been dead-burned for use as a refractory, of which the steel industry is by far the largest consumer. Production has accordingly fluctuated more or less in parallel with the fluctuations in steel production. Production figures from Stevens County have not been published for many years, but it is perhaps reasonable to estimate that since their discovery these deposits have yielded more than 5 million tons of crude magnesite, practically all of which has been for domestic consumption, except for a brief period during World War II when the industry had also to supply some of the needs of our allies.

DESCRIPTIONS OF DEPOSITS

KEYSTONE DEPOSIT

The Keystone quarry (sec. 9, T. 31 N., R. 39 E.) was first opened in the 1890's as a marble quarry and until about 1905 was operated by the U.S. Marble Co. as a source of ornamental stone. As late as 1945, a closely spaced row 10 feet long, of drill holes in the brucite body, which was a unique feature of the Keystone deposit, could still be seen as evidence of the size of blocks that were quarried and of the care taken in breaking the material free. Published analyses (Shedd, 1903) showing the high magnesia content of the "marble" led to its "rediscovery" as a magnesite deposit in 1916, and it was from the Keystone that the first magnesite shipments from Stevens County were made. Discovery and development of the Finch deposit, which was larger and more favorably located and was owned by the same company (Northwest Magnesite), put the Keystone in a standby position for many years. Production was resumed in 1946 after completion of an aerial tramline to the deposit and has continued on a limited basis. Total production is estimated to have been around 500,000 tons of crude magnesite. The ore was rather high grade, averaging only 3 percent silica and 3 percent lime.

The magnesite at the Keystone was for the most part dark gray to almost black. The deposit was roughly triangular and had a maximum width of 250 feet and a length of about 500 feet in a northeasterly direction. The exposed surface was about 63,000 square feet, excluding a wedge of dolomite in the northeastern part. The floor of the deposit lay at about 3,600 feet, or roughly about 250 feet below the highest outcrop. Owing to the very steep slope, the deposit was wedge shaped and had a blunt edge toward the southeast.

Toward the northern end of the deposit there occurred a lens of rather massive almost translucent gray-green brucite, prized by the early "marble" workers because of its attractive color and the comparative ease with which it could be worked (hardness of brucite=2½; of calcite marble, 3). Unfortunately, objets d'art and ornamental stone carved from the brucite showed in the course of time an unpleasant tendency to develop a white powdery coating (hydromagnesite) not characteristic of the better quality of marbles. This was indeed one of the features that led to the curtailment and eventual closing of the early quarrying operations at the Keystone. The brucite, probably, was formed as a hydrothermal alteration, controlled by a fault channelway, of earlier magnesite. In turn, the brucite was locally altered to hydromagnesite. The brucite body was about 20 feet wide at its maximum, and extended some 100 to 150 feet in length. Chemical analysis and microscopic study of the material by Clarke (1903) led him to conclude that, although brucite was the dominant mineral, it was variably intermixed with serpentine, chlorite, hydromagnesite, and dolomite.

FINCH DEPOSIT

Between 1917 and its virtual exhaustion in 1954, the Finch deposit (sec. 30, T. 32 N., R. 40 E) was the principal producer—and for many years the sole producer—of magnesite in the State of Washington. Through two World Wars it supplied the bulk of our domestic needs for refractory magnesia. Total production is estimated to have been well over 3 million tons. The first production from the Finch deposit was developed by the Washington Magnesite Co. in 1916. In 1917 the deposit and the company were taken over by the Northwest Magnesite Co., and for many years the Finch deposit was that company's principal source of production.

The overall length of the main deposit was 1,600 feet, and the width ranged from 300 to 650 feet. This lenticular body had a maximum thickness, normal to the dip, of about 240 feet. Old photographs show a bold outcrop ridge with ravines on either side where the deep lower quarry was developed. This lower quarry, from an altitude of 2,145 to 2,300 feet, was mined from a series of benches, and the magnesite was dropped to an underground haulage level at 2,105 feet. An upper quarry (sometimes referred to as the North Finch) extended from 2,420 to 2,560 feet. It contained some very high-grade material, but the proportion of rejects increased with depth. The magnesite at the Finch was in general finer grained and much lighter colored than that in most of the deposits. Indeed, as indicated above, color alone could not be successfully employed at the Finch to discriminate ore from gangue. Greenstone dikes and sills, minor faulting, and local areas of residual dolomite contributed to the problems of selective mining at the Finch, though probably less than at most of the deposits.

ALLEN-MOSS DEPOSIT

The Allen-Moss deposit (secs. 25 and 36, T. 32 N., R. 39 E., and secs. 30 and 31, T. 32 N., R. 40 E.) was discovered in 1916 and put into production by the American Mineral Production Co. in 1917. This company constructed a standard-gage railroad line from Valley to the deposit and produced actively during World War I. For a few years following the close of the war, a small amount of magnesite was sold to the Northwest Magnesite Co., because of the convenient location of that company's mill at their Finch quarry and their calcining plant at Chewelah. In 1920, the American Mineral Co. ceased operations, and in 1923 the Northwest Magnesite Co. purchased from them their four deposits: the Allen, the Moss, the Woodbury, and the Red Marble. The Allen-Moss remained in standby position until, with the onset of World War II, it was placed in production in 1941. The deposit was exhausted in 1948, after a total production estimated at about 700,000 tons.

The deposit was, in its broad aspects, generally similar to the Finch, except for a notable "island" of unreplaced dolomite which occupied a somewhat central position in the deposit, and for the relatively flat-lying attitude of the strata which, coupled with faulting and with minor "rolls," added to the mining difficulties.

WOODBURY DEPOSIT

The Woodbury deposit (sec. 1, T. 31 N., R. 39 E.) was located in 1916 and in 1917 provided the initial production (amounting to a few hundreds of tons) of the American Mineral Production Co. With the more favorable conditions at the Allen-Moss deposit, operations at the Woodbury were closed down, and the property remained idle until in 1949 the Northwest Magnesite Co. mined out the small reserves (an estimated 50,000 tons) remaining.

MOUNTAIN VIEW, NOGUE, AND PHOENIX DEPOSITS

These three small deposits, all within the drainage area of Huckleberry Creek in secs. 3 and 4, T. 31 N., R. 39 E., contain only insignificant tonages of magnesite—probably at best no more than a few thousand tons at each deposit. The Phoenix deposit, which is a few hundred feet east of the access road for the tramline to the Keystone deposit, and was relatively inaccessible until the construction of the tramline, illustrates very well on a small scale the incipient stages of the replacement process whereby dolomite was converted to magnesite. The magnesite, nearly black, contrasts well with the much lighter colored dolomite, and the control by bedding planes and by vertical joints of the course of magnesitization can be clearly seen.

MIDNIGHT DEPOSIT

The Midnight deposit (sec. 7, T. 31 N., R. 39 E.) is represented by some 5 outcrops of coarse-grained dark magnesitized dolomite, scattered along the strike of the Stensgar dolomite for a distance of nearly 2,000 feet. A small quarry was opened in one of the outcrops, probably during World War I, although no record now remains. Surface indications suggest that no more than a few thousand tons are present in all; but if the magnesite is continuous in the area between the outcrops—which is highly unlikely in our opinion—there could be several hundred thousand tons in this deposit.

DOUBLE EAGLE DEPOSIT

The Double Eagle deposit (sec. 18, T. 31 N., R. 39 E.) was developed by the Western Materials Co. during World War I. Production was somewhat sporadic until 1921, when the company ceased operations. Much of the small production was chemical-grade magnesite for use in plastic magnesia rather than refractory magnesite. The deposit was extensively sampled by representatives of the Westvaco Chemical Co. early in World War II, but no further development has been undertaken. Several hundred thousand tons of magnesitic material may be present, but most of it is thoroughly intermixed with dolomite and is rather low grade. As at the Keystone, the magnesite is rather coarse grained and dark gray to almost black, making a striking contrast with the lighter gray dolomite (fig. 3).

CROSBY DEPOSIT

The Crosby deposit (sec. 18, T. 31 N., R. 39 E.) is credited with a very small production during World War I, but details are unavailable. Exposures in the area are poor, and tonnage estimates are very uncertain. It seems unlikely that the deposit contains more than a few thousand tons of magnesite. The "ore" from the Crosby is distinctive because, except for the Red Marble, it is the only deposit in which magnesite is red. Some of the Crosby ore is indeed almost a cinnabar red, and it is said that the deposit was first located as a quicksilver claim. The magnesite is also somewhat finer grained than in most other deposits.

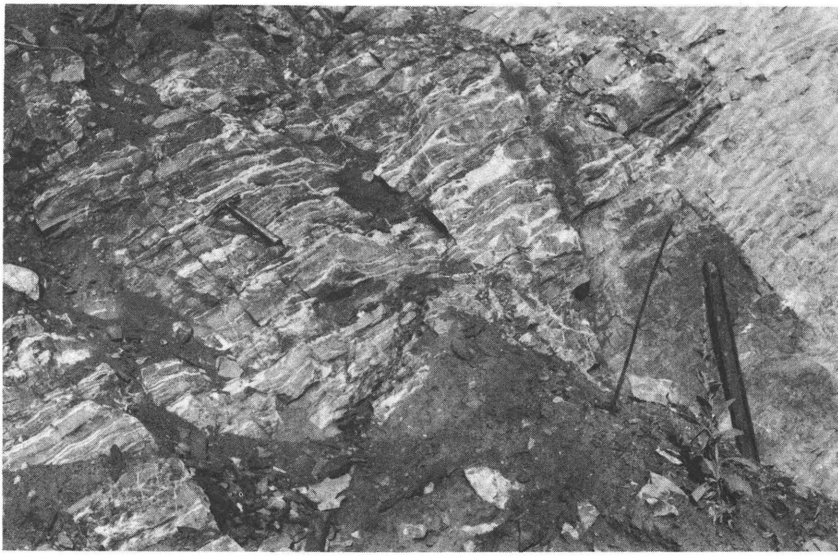

FIGURE 3.—Magnesite deposit in the Double Eagle quarry. Dark thin-bedded magnesite is interlayered with white recrystallized dolomite; grades (to right in photograph) into gray nonmagnesitized dolomite of the Stensgar.

DAVIS DEPOSIT

The Davis deposit (sec. 18, T. 31 N., R. 39 E.) is some 2,000 feet north of the Crosby deposit and about 1 mile south-southeast of the Midnight deposit. It is in the same section of the Stensgar dolomite as the Midnight. The deposit is poorly exposed and seems to be of insignificant tonnage.

RED MARBLE DEPOSIT

The Red Marble deposit (secs. 24 and 25, T. 31 N., R. 38 E.), discovered early in World War I, has long been recognized as one of the major concentrations of refractory-grade magnesite in the United States. It was owned originally by the American Mineral Production Co. That company had built a railroad from Valley to its Woodbury and Allen-Moss deposits, and late in World War I started a branch line to the Red Marble. Grading of the road had been completed and ties had been piled ready for setting out when the end of the war curtailed operations. The railway was never completed, and the grade was later taken over by the county and the U.S. Forest Service for a roadway. Because of its relative inaccessibility, the Red Marble deposit lay idle until late in World War II when the Northwest Magnesite Co., which had acquired the property in 1923, started to extend its aerial tramway from the Keystone to the Red Marble, and initiated also an extensive drilling campaign in order better to evaluate and to develop the deposit. The Red Marble was put into production in 1949. Details of production and of reserves have not been released, but it may be reasonable to estimate that the Red Marble deposit contained originally more than 5 million tons of magnesite. Much, but by no means all, of the magnesite is distinctively red or pink, possibly due to a somewhat higher than average content of ferric iron which confers premium characteristics on the ore, for less iron must be added at the time of calcining.

The deposit lies in a bluff facing north on a deep ravine draining into the north fork of Deer Creek. Magnesitic material occurs through a vertical range of some 400 feet and was exposed at intervals along the strike for more than 3,000 feet. The Stensgar dolomite in which the Red Marble deposit lies is repeated to the west by faulting, but in this section the more westerly block shows little magnesitization. However, a claim known as the Black Bear (sec. 24, T. 31 N., R. 38 E.) a quarter of a mile west-northwest of the Red Marble deposit, in the western block, locally shows some magnesitization.

U.S. MAGNESITE DEPOSIT

The U.S. Magnesite deposit, in sec. 10, T. 30 N., R. 38 E., was developed by the U.S. Magnesite Co. during World War I. Several carloads were shipped in 1917, but operations ceased when the shipment proved unusable because of excessive lime and silica. Development consisted of a short adit and a small quarry, now largely obscured by slumping. It is estimated that not more than a few thousand tons of low-grade magnesitic material is present.

TURK DEPOSIT

The Turk deposit (sec. 1, T. 29 N., R. 37 E., and sec. 36, T. 30 N., R. 37 E.) is of interest not only because it is the most southerly part of the Stensgar dolomite where extensive magnesitization has been noted but because it is one of the larger concentrations of magnesite in Stevens County. Largely on State land, it was tested during World War II by some 5,000 feet of drilling, under the supervision of the Washington State Department of Conservation and Development. Bennett (1943) sets forth the essential geological and chemical data and concludes that slightly more than 2 million tons of magnesite are present. The deposit has not as yet been developed commercially, but it constitutes an important reserve.

Much of the magnesite at the Turk deposit is nearly white, contrasting with the darker shades of magnesite elsewhere in the district.

The suggestion has been made that this feature may be the result of bleaching, because of proximity of granitic intrusives, which elsewhere in the district are generally more distant from the magnesite concentrations.

OTHER USEFUL MINERALS

Although magnesite has been for many years the principal mineral resource of the mapped area, and although no attempt was made to investigate deposits of other minerals, it may be desirable to conclude this report with brief mention of deposits that have been or might become of commercial interest.

In general, it may be said that most of the metallic deposits were discovered in the late 1880's and early 1890's, that most of the ore was produced before World War I, and that subsequent new developments and efforts to revive old mines have been for the most part disappointing. Indeed, at the time of our studies, most of the metal mines were relatively inaccessible. The description given in the reports by Weaver (1920) and Jenkins (1924) are consequently based on more extensive observations than in many instances would be possible today, and the reader is referred to these bulletins for geologic and mineralogic detail. Additional reference material may be found in Huntting (1956).

Most of the metal mines have been developed in complex silver-lead-zinc-copper ores, and the principal products have been silver and lead. One district, the Deer Trail, is reported to have yielded more than $3 million, chiefly in silver and lead. Production from other metal mines has been relatively small.

Antimony.—There are two noteworthy occurrences of this metal in the area. At the Cleveland mine (sec. 9, T. 30 N., R. 38 E.) the lead-antimony sulfide mineral, boulangerite, at one time constituted an important part of the ore, and a few shipments were made, principally of boulangerite, in order to separate the antimony. The Wells-Fargo mine (sec. 36, T. 31 N., R. 38 E.) has undergone sporadic development since it was first located in 1890; but only one small shipment is known to have been made (Purdy, 1951). The ore mineral is stibnite, the gangue quartz. The deposit is in the McHale slate.

Barite.—Many veins of barite occur in the district, most of them within that part of the Edna dolomite lying south of the road from Valley to the Red Marble. Shipments have been made from deposits in sec. 11, T. 30 N., R. 38 E. (Valentine, 1949).

Dolomite.—The Stensgar dolomite, although in places a good grade of dolomite, cannot yet be considered competitive with the more extensive and readily accessible bodies of dolomite in the Old Dominion limestone of Weaver (1920) near the area (Bennett, 1944). The thicker parts of the Stensgar, however, are a potential resource.

Copper.—A little copper has been recovered as a byproduct of mining for lead and silver, and deposits where copper is the dominant mineral have been prospected, principally in the Edna dolomite near greenstone dikes and sills. The dominant minerals in these copper deposits are pyrite, chalcopyrite, bornite, and tetrahedrite in a gangue of quartz and coarsely crystalline carbonate. Secondary minerals include malachite, azurite, and chalcocite. The only recorded shipments of copper ore appear to have been from the Edna mine in sec. 9, T. 31 N., R. 39 E. (Weaver, 1920, p. 181–182). No production is known subsequent to 1904.

Quartzite.—Quartzite is extensively exposed throughout much of the area. The Precambrian quartzites are in general relatively thin and impure, and, except very locally, offer little commercial possibility. The Addy quartzite, in contrast, is a relatively thick, uniform, clean quartzite. During World War II, considerable tonnages were shipped for the manufacture of ferrosilicon. A deposit a few miles north of Chewelah (sec. 4, T. 32 N., R. 40 E.) was being worked as late as 1947 by the Ohio Ferro-Alloys Co. of Tacoma (Valentine, 1949).

Silver and lead.—Silver and lead have constituted the principal metallic resource of the area, though very few mines have been active in the last 30 years. Argentiferous galena is the chief ore mineral, though most of the deposits involve a rather complex mineralogy. Most of the deposits are in the southern part of the mapped area and are largely found in, though by no means wholly confined to, the Stensgar dolomite. Mineralization probably was related to the intrusion of the Loon Lake granite. Details concerning individual deposits may be found in Huntting (1956), Jenkins (1924), and Weaver (1920).

Slate.—Slate was produced at one time from a quarry in a slaty member of the Buffalo Hump formation near the Red Marble magnesite deposit, in sec. 19, T. 31 N., R. 39 E. The quarry has not been active for many years, though a considerable tonnage of fairly good quality slate remains.

Talc.—Limited workings have exposed for some 50 feet a deposit of talc known as the Firminhac, in sec. 15, T. 30 N., R. 38 E. The talc is associated with basic dikes in the Edna dolomite. No shipments are known to have been made.

Tungsten.—The Germania mine in sec. 13, T. 29 N., R. 37 E., was for many years an active producer of wolframite-scheelite ore (Kerr, 1946, p. 129–130). The deposit lies in the upper part of the Loon Lake granitic batholith.

Uranium.—About 2 miles east of the southeast corner of our area, a large uranium deposit, the Midnite mine, was discovered in 1955 near the contact of the Loon Lake granite and metasediments of the Togo formation. Subsequently, other smaller deposits have been discovered in the area. At the Midnite mine the ore near the surface consists principally of secondary uranium minerals—meta-autunite, phosphuranylite, and uranophane. Uraninite and coffinite have been identified from drill cores. The ore is almost entirely within the Togo formation, but a small amount of ore has been produced from secondary minerals in the granite at the contact (G. E. Becraft, written communication, 1957).

Zinc.—Zinc is present, in various amounts, in most of the deposits that have been worked for silver and lead. In a few, such as the Cleveland mine (mentioned under Antimony), it has been an important coproduct; in most it has been a minor byproduct. Sphalerite is the principal ore mineral.

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