PROGRESSIVE KINETIC METAMORPHISM IN THE MISSI SERIES NEAR FLINFLON, MANITOBA.[1]

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ABSTRACT.

The Missi Series near Flinflon, Manitoba, comprises a group of early Precambrian conglomerates and greywackes that lie in structural conformity upon Wekuskoan volcanics. Both the Wekuskoan and the Missi are intruded by quartz diorite and granodiorite and by presumably related dikes and sills of albite and quartz-albite porphyry, ultrabasic dikes and aplite.

All the rocks have been kinetically metamorphosed to some extent, and three zones of metamorphic intensity are recognized in the Missi Series. Greywacke in the chlorite zone is a quartz-albite-epidote-sericite-chlorite rock; through a series of gradational changes it becomes a garnetiferous quartz-oligoclase-mica schist in the garnet zone. Quartz-diorite, albite and quartz-albite porphyry in the chlorite zone are quartz-albite-epidote-chlorite rocks; in the garnet zone they are quartz-oligoclase-biotite-amphibole gneisses. Extrusive rocks of the Wekuskoan Series show similar changes, and the pebbles of the conglomerates, regardless of their original character, attained a remarkable degree of adjustment to their environment during metamorphism.

The field relations are such that neither proximity to an igneous mass nor depth of burial can be held to account for the advance in metamorphic rank of the various rocks. Proof is offered to show that the schistosity developed by internal shearing on closely spaced planes, and as the amount of movement in the higher rank zone is demonstrably greater than in the low rank zones, the hypothesis is proposed that the heat necessary for metamorphism was developed mechanically by and during the intimate internal shearing.

INTRODUCTION.

The Missi series is a group of early Precambrian sedimentary rocks extensively exposed along the southern border of the Canadian shield in northern Manitoba and Saskatchewan. This paper refers only to that portion of the series which outcrops a short distance east and north of the town of Flinflon, located on the Manitoba-Saskatchewan boundary, approximately 400 miles northwest of Winnipeg.

All of the rocks of the district, including volcanics, sedimentary rocks, and intrusives, have been somewhat metamorphosed, and a definite series of progressive morphological and mineralogical changes can be traced in each of the several varieties of rocks, from the low rank chlorite zone through

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to the garnet zone, with a variety of products dependant on the original compositions. (Fig. 1.) The literature on progressively metamorphosed rocks of this character is growing

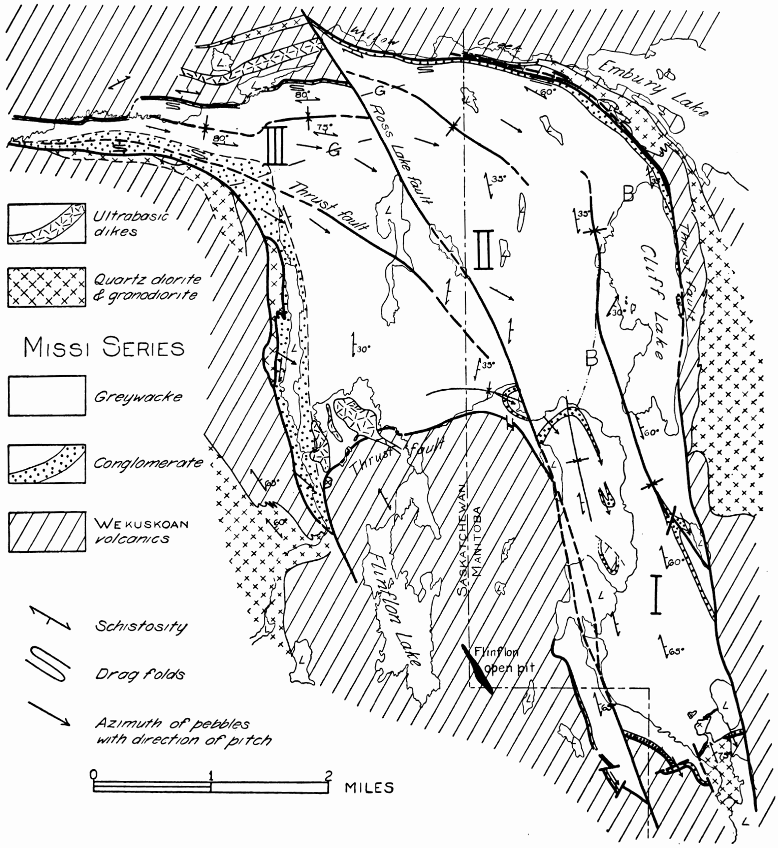

Fig. 1. Sketch map of geology in the vicinity of Flinflon, Manitoba. I, chlorite zone; II, biotite zone; III, garnet zone; B, biotite isograd; G, garnet isograd. Fault lines are labelled. Other heavy lines, normal or intrusive contacts.

rapidly, but there seems to be no previous description of such a series from North America. It may be of value, therefore, to record their presence here, and to draw attention to some of the problems involved in their origin.

GENERAL GEOLOGY.

The Missi series near Flinflon ranges in thickness from 5,000 to 9,000 feet, with the top unknown. It lies in structural conformity upon the Wekuskoan group of volcanics, the oldest rocks in the region. Both the Missi and the Wekuskoan are cut by a suite of igneous rocks that include, from oldest to youngest, major bodies of quartz diorite and granodiorite, dikes of quartz-feldspar porphyry, basic to ultrabasic dikes, possibly lamprophyres, and aplites. Finally, all the rocks were injected by numerous quartz veins.

The sedimentary rocks are folded into an asymmetric, north-trending syncline, with a steeply inclined to overturned eastern and a gently inclined western limb. The fold is cut by east-dipping thrust faults that strike a few degrees west of north from the southern end of the area to about the latitude of the north end of Cliff lake. Thereafter the fold axis, the faults with one exception, the schistosity, and contacts curve to the west in a wide arc that turns the grain of the country through an angle of 90°, from north-south to east-west. Ross lake fault, the exception to the rule, is the youngest fault in the area. It offsets the fold axis, the metamorphic belts, and the older thrust faults through a horizontal distance of 3,300 feet.

The early movements (pre-Ross lake fault), which curved the structural lines so drastically, formed beautiful drag folds everywhere in the west-striking beds. (Figs. 1 and 5, B.) The shape of the folds indicates that the rocks on the north moved west relative to those on the south. The continuity of the strike for over 20 miles to the west shows that the structure is not merely of local development; every one of the older structural features of the district is bent into conformity with it. Furthermore, within it, in marked contrast to the beds farther south, the rocks are highly schistose and intensely deformed. Evidently they have been affected by profound east-west movements that, though on a regional scale, were localized to a considerable extent along this belt. These movements are believed to be of fundamental importtance in the interpretation of the metamorphism of the area, as noted in a later section of this paper.

THE MISSI SERIES.

The Missi series is particularly well suited for study and comparison of progressive changes in intensity of metamorphism. The greywackes are astonishingly uniform in grain size and composition from top to bottom. Aside from thin lines of magnetite grains or local lenticular pebble beds, the rocks are massive and without bedding; hand specimens from one part of the section cannot be distinguished from those from any other part of the section in areas of equal metamorphism. The similarity and persistence of pebble beds from the lowest into the highest rank zone, the uniformity of the metamorphic product in any one zone, and the small areal extent of the series, all offer firm support for the assumption that the composition of the greywackes was originally essentially uniform over this whole area. Thus it is possible to use the series as a key on which to delimit the zones, confident in the belief that the mineralogical changes observed from one to the other are due, not to original differences in composition, but to environmental changes during metamorphism. The zones so defined can also be used as control to enable comparison of the metamorphic products of the interbedded conglomerates, the underlying volcanics, and the intruded igneous rocks, either within the same zone or from one zone to another.

DEVELOPMENT OF THE ZONAL CONCEPT.

The first zonal mapping in rocks of this type was done by Barrow in 1893 (pp. 330-358). He studied an area of metamorphosed sedimentary rocks in the southwestern Highlands, and showed that zones of different metamorphic intensity could be distinguished by mapping the distribution of certain critical minerals. He used staurolite, kyanite, and sillimanite as critical minerals. and named the zones after them. Other zones were added by further work in 1912.

Tilley extended Barrow's work in 1925 (pp. 100-112) and since then, he and his students have vigorously pursued the study of similar groups of rocks. Tilley added precision to the term "zone" defining it as an area bounded by lines or "isograds" drawn between points of first entry of the distinctive new minerals into the assemblage. Such a line, joining for example, the points at which biotite appears in the suite, must have been a line of equal pressures and temperatures during metamorphism. According to Tilley (1924, p. 169), an " 'isograd' . . . . . may be defined as a line joining points of similar P.T. values under which the rock as now constituted originated. In reality, an isograd is the intersection of an inclined isogradic surface with the earth's surface. So we may speak of a biotite isograd, garnet isograd, etc. . . . ." Theoretically, a separate isograd might be drawn for each newly developed mineral in a progressively metamorphosed series. As a matter of common practice, however, only five zones are recognized. They, with their bounding isograds, are as follows:

Chlorite zone
  Biotite isograd
Biotite zone
  Garnet isograd
Garnet zone
  Kyanite isograd
Kyanite zone
  Sillimanite isograd
Sillimanite zone

Of these, only the first three are developed in the Missi series.

DELIMITATION OF ZONES IN THE MISSI SERIES.

The approximate locations of the biotite and garnet isograds in the metamorphosed greywackes are shown on the accompanying map. The garnet isograd, marked by the appearance of garnet in the suite, and by a pronounced and abrupt increase in the degree of development of schistosity, is recognized much more readily in the field than the biotite isograd and the accuracy with which it is located is correspondingly greater. In order to locate the points of first entry of biotite, many specimens from critical places would have to be examined microscopically. Although two complete suites of specimens, one on either side of the area, were collected across the zones, in addition to many within the zones themselves, the available data are insufficient to allow more than a first approximation to the position of the biotite isograd throughout most of its length. Fortunately, the exact location of the line is not vital to the argument; an ample number of specimens were collected to show that the southern portion of the area is in the chlorite zone, the middle and largest part of the area is in the biotite zone, and the remaining north and northwest part is in the garnet zone.

Character in the chlorite zone.

The greywacke in the chlorite zone is a slightly schistose rock, dark on fresh surface, weathering light bluish or rusty grey. The quartz grains range in size from 1 cm. pebbles to grains 0.3 mm. in diameter; generally they are less than 1 mm. Many of them have a slight bluish or violet tinge, especially apparent on fresh fractures. Fracture takes place through rather than around the grains.

In thin section the rock is seen to be composed mainly of quartz, chlorite, white mica, albite, and epidote. Quartz predominates, but is somewhat less than 50 per cent of the whole, and white mica, chlorite, and albite are about equal. Apatite and magnetite occur in minor amounts and grains of blue-black tourmaline, zircon, and titanite are not uncommon. Microcline was identified in only one section and orthoclase is absent.

The larger, subangular to well rounded quartz grains are nearly all unbroken, though most of them show strong undulatory extinction. Commonly they are surrounded by flakes of chlorite and white mica which are arranged in radiating flame-like growths perpendicular to the circumferences of the quartz grains. Albite, averaging An_4[2] in composition, is the only feldspar in this zone. In the rocks of lowest grade within the zone the albite crystals are clouded by flakes of white mica, but from intermediate to highest grade within the zone the grains are perfectly clear, practically devoid of cleavage, and rarely twinned. More than half of the twinning present is according to the pericline rather than the albite law.

White mica, in most cases somewhat greater in amount than chlorite, is abundant in all the rocks. Specimens from the extreme southern end of the area (lowest grade) contain felted aggregates of this mica and chlorite flakes with outlines suggestive of original feldspar crystals, but no feldspar other than albite occurs. In rocks of slightly higher grade there are no such aggregates, and the white mica is evenly distributed through the section or grown in flame-like borders around quartz, magnetite, or less commonly, around albite or epidote grains. The chlorite is optically positive, with $\beta = 1.618 \pm 0.002$, properties that serve to classify it as the high aluminum variety, prochlorite. The tiny plates lying parallel to one another in the plane of schistosity, are evenly distributed through the sections; in addition, large, undeformed, more strongly pleochroic flakes occupy the corners of "eyes" of

[2] The subscript indicates the percentage of the anorthite molecule in the plagioclase as determined by optical methods. The same convention is used throughout this paper.

schistosity around magnetite grains. The cleavages of these larger flakes are not oriented with the schistosity and are not uncommonly perpendicular to it.

Epidote occurs in scattered granules, in aggregates of granules, and in single large prismatic crystals elongate parallel to b′. Many of the larger crystals are zoned, with rims that show slightly stronger pleochroism and notably stronger birefringence than the cores. It is invariably optically negative, and must therefore contain more than ten per cent of the iron-bearing molecule.

Apatite is present in small stumpy prisms that show no evidence of deformation. Magnetite (and ilmenite) is commonly in undeformed euhedral crystals with the lines of schistosity sweeping around them. Tourmaline in stumpy prisms or, very rarely, in sheaf-like aggregates of slender prisms, is not uncommon.

All the minerals, with the possible exception of quartz, crystallized either wholly or in part after shearing movements ceased. The white mica and chlorite flakes are undeformed, and the peripheral growths of these minerals around quartz, etc., would surely have been destroyed if the rock had been affected by any internal movements after they were formed. The clear, uncleaved, untwinned character of the albite, as well as its sutured borders against other minerals, indicate it to be a product of neocrystallization. Similarily, the undeformed euhedral apatite and tourmaline prisms suggest that these minerals were formed late.

Character in the biotite zone.

The rock in which biotite makes its first appearance, though with better developed cleavage and with a silky luster on fresh surfaces, is quite similar to greywackes in the chlorite zone. It is made up, as before, of quartz, chlorite, white mica, albite, epidote, minor apatite, magnetite, tourmaline, zircon, and titanite, but it has the significant addition of green biotite. This mineral ranges in amount from negligible at the biotite isograd to 15 per cent of the rock at the garnet isograd. It is moderately pleochroic, with X = pale yellow, and Z and Y = clear brownish green. The flakes are small, undeformed, and intimately intergrown with white mica and chlorite.

Concomitant with the increase in the amount of biotite, white mica and chlorite decrease in quantity. In the highest grade within the biotite zone there is practically no chlorite, and the remaining white mica flakes have increased in size until they may more properly be called muscovite. The albite is slightly more calcic (An_{6-8}), and a few relatively large clastic grains of quartz remain, though most of them have been broken down into lenticular or spindle-form aggregates of smaller grains. At the upper limits of the biotite zone, practically all resemblance to the greywacke of the chlorite zone has been obliterated; the rock is an albite-epidote-mica-quartz schist.

Character in the garnet zone.

The typical rock of this zone is a coarsely crystalline, crenulated garnetiferous quartz-oligoclase-mica schist. The crenulations are small, somewhat irregular, and discontinuous along their length; in places they are well enough developed to give the rock a pronounced linear structure that pitches steeply east.

The garnets, which serve to give the zone its name, are mostly less than 1 mm. in diameter, and, adjacent to the biotite zone, very few in number. They are anhedral, elongate parallel to the schistosity. The schistose planes sweep around the grains, but no definite evidence that the garnets had been rolled was observed. Index determinations suggest that the garnet is high in the almandine molecule.

The quartz is completely recrystallized, and occurs in narrow bands elongate parallel to the schistosity, or in smaller grains intergrown with sutured borders. The only feldspar present is oligoclase-andesine (An_{28-32}). It occurs in clear small anhedrons somewhat elongate parallel to the schistosity. Pericline twins are uncommon and albite twins exceedingly scarce; cleavage is fairly well developed parallel to (001), but is rare parallel to (010).

Both biotite and muscovite are abundant. The biotite is strongly pleochroic with $X =$ pale yellowish brown and Y and $Z =$ very dark brownish green or almost black. It is in small flakes that are oriented, for the most part, parallel to the crenulations of the planes of the schistosity; however, many flakes lie across the schistose planes, and none of them show any evidence of post-crystallization deformation. Muscovite, in contrast to the biotite, forms large flakes, many of which are bent and cross-fractured. The cleavage is so well developed in some of these larger flakes that, in thin section, they

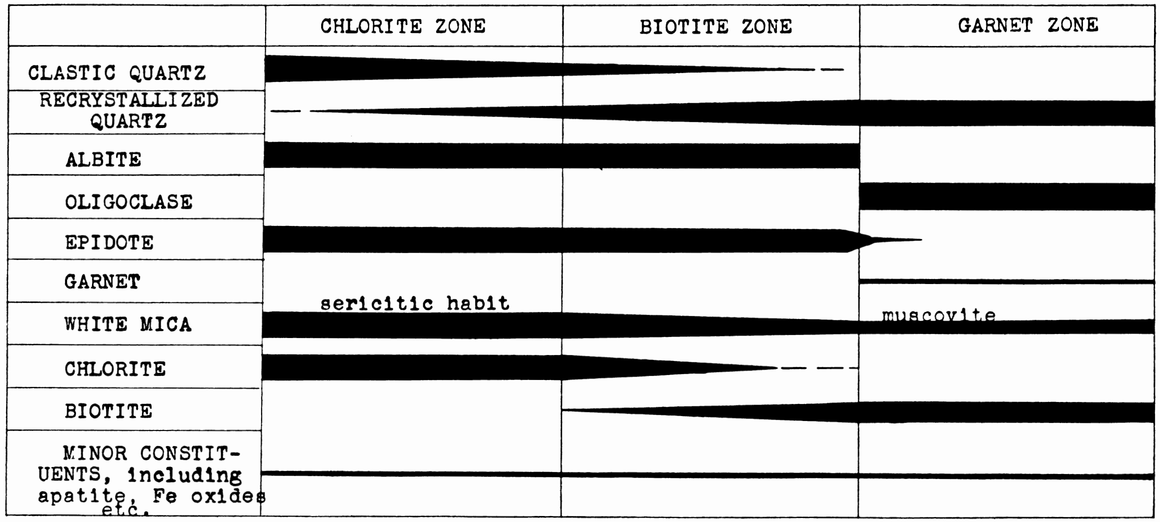

Fig. 2. Diagrammatic representation of mineral paragenesis in progressively metamorphosed greywackes of the Missi Series, Flinflon, Manitoba.

seem to be made up of bundles of fibers; obviously the mineral has been deformed since crystallization.

Tourmaline, zircon, and magnetite occur in small amounts. Apatite occurs in unusually large, somewhat broken crystals, wrapped around by the planes of schistosity. Epidote and chlorite are absent.

MINERAL PARAGENESIS IN THE GREYWACKES.

A diagrammatic representation of the mineral paragenesis observed in the greywackes is given above. Unfortunately, chemical analyses of metamorphic minerals of this type are notably conspicuous by their absence in the literature. Without them, and without bulk analyses of the various rock types, anything approaching a thoroughgoing discussion of the mineral paragenesis is impossible. In the following paragraphs, the mineralogical changes are summarized, and an attempt, in part well supported by the observed facts and by the work of others with similar rocks, and in part purely speculative, is made to interpret them. It is perhaps worthy of restatement here that the metamorphism is on a regional scale, that the greywackes are regarded as uniform and the system isochemical throughout, and that therefore new minerals are formed only at the expense of those already present, without either introduction or loss of material.

Quartz, in detrital grains to 1 mm. in diameter in the chlorite zone, is mechanically crushed and drawn into lenses in the biotite zone; at the garnet isograd no recognizable detrital grains are left. In the crenulated schists of the garnet zone, quartz occurs either in elongated spindle-form aggregates or in narrow ribbon-like crystals that lie with their length parallel to the schistosity. Recrystallization of quartz in the two lower-rank zones is restricted to the formation of sutured borders around the grains. In the garnet zone, the growth of ribbons is proof that recrystallization here becomes a much more active process. In the highly crenulated schists, those described below in which some metamorphic retrogression has occurred, many of the ribbons are broken down again into aggregates of grains with sutured borders. Quartz seems to play little active part in the chemical reactions during metamorphism. Harker considers that, by diluting the more active constituents and separating them from one another, it acts as a retarder; that is, it slows up, but does not otherwise change the reactions involved.

Chlorite remains unchanged up to the biotite isograd. At and beyond that line it combines with white mica to produce biotite. In what may be the first step of the transformation, the white mica takes on a very faint greenish tinge that, as the reaction proceeds, gradually deepens in color until the pleochroism from colorless to green is quite apparent. All, or nearly all the chlorite is used up before the garnet isograd is reached; white mica remains in considerable excess and in the garnet zone forms plates 4 or 5 mm. in diameter.

The only observable change in biotite in the higher zones is a gradual deepening in color. No "normal" red-brown biotite occurs in the series, but very dark-green varieties are associated with garnet. Phillips has analyzed two specimens of biotite, one light-green and the other dark (1930, p. 255), and these analyses suggest that the deepening in color may be due to an increase in the content of ferrous iron.

The plagioclase undergoes a series of reactions that have become recognized as characteristic of this kind of metamorphism. The original feldspars in the greywacke were derived from a field of volcanic rocks of intermediate composition on the average. It might be supposed, therefore, that andesine or other calcic varieties of plagioclase might predominate, but the striking fact is that in the chlorite and biotite zones, the only feldspar present is albite, averaging An_4. In view of the abundant epidote, which occurs both poikiloblastically in albite and scattered through the section, it is probable that the original average plagioclase was more calcic than albite, and that both albite and epidote were formed by its breakdown during the first stages of metamorphism according to the familiar equation:

plagioclase + water + iron = albite + epidote + alumina + silica

The amount of iron taken up by the epidote determines, above a certain critical minimum which must be released, the amount of excess alumina. The albite, in the lowest grade in the chlorite zone, is clouded by flakes of white mica, possibly paragonite produced by reaction of the excess alumina with newly formed albite and water. As the biotite isograd is approached, the mica is either resorbed or eliminated, for the albite there is water clear. If the mica is resorbed and alumina expelled, it may enter into the chlorite, which is here the high aluminum variety, phochlorite.

Water-clear albite persists with little change through the biotite zone. However, with the appearance of garnet, the anorthite content of the plagioclase rises abruptly from An_{6-8} to An_{28-32}. A similar rapid conversion of albite to calcic oligoclase or sodic andesine has been noted in rocks of this type by several workers. (Wiseman, 1934, p. 385; Phillips, 1930, p. 250.) Turner (1933, p. 540), summarizing the occurrence of oligoclase in certain schists states, "Oligoclase normally appears as a product of dynamothermal metamorphism at relatively high grades such as prevail in the zones of almandine and perhaps kyanite. It is accompanied either by deeply colored hornblende, hornblende and biotite, or biotite and muscovite, according to the chemical composition of the rocks in which it occurs." Evidently it is produced by a reaction that can only occur when certain relatively high values of temperature and pressure have been passed, and once initiated, it proceeds rapidly to completion.

Epidote is greatly diminished in amount above the garnet isograd, that is, concomitant with the appearance of oligoclase. It is generally assumed that the necessary anorthite is formed by the reverse of the reaction by which the original plagioclase was broken down. However, by this reaction, ferric iron is released, and if all the calcium of epidote is converted to anorthite, both silica and alumina must be supplied. Quartz, biotite, and white mica are the only other minerals that are present; if the system is isochemical, they must absorb the excesses and supply the deficiencies of the reaction. Biotite darkens in color; presumably it absorbs the excess iron. Quartz can, of course, supply the necessary silica, but the only source for alumina is the micas. If they are pure potash micas, they cannot supply alumina without at the same time forming some potash feldspar, and there is no potash feldspar in the suite. However, if either mica contains the sodium-bearing molecule, that molecule may enter into the reaction, make up the deficiency of alumina, and form a little albite which could, of course, immediately react to produce more oligoclase. The complete reaction could then be written,

$$\text{albite} + \text{epidote} + \left\{ \begin{array}{c} \text{soda mica} \\ \text{potash mica} \end{array} \right\} + \text{quartz} = \begin{array}{c} \text{plagioclase} + \text{potash} \\ \text{mica} + \text{water} \end{array}$$

One other possibility must be recognized. The dark biotite analyzed by Phillips contains over five per cent more calcium than the light-green biotite. It may be, therefore, that the epidote ⟶ anorthite reaction instead of taking up alumina, releases calcium which enters into the biotite.

The problem of the origin of garnet, though there is actually only a very minute amount of that mineral, is one of the most difficult in the mineral paragenesis, certainly impossible of final solution without numerous analyses. It must be formed from some of the minerals in the suite at the garnet isograd, and of these the albite alone can be ruled out. Tilley (1926, pp. 43-44), followed by Phillips (1930, p. 252), consider that garnet forms from chlorite that remains over through the biotite zone, and Harker (1932, p. 192) gives convincing illustrations of garnet pseudomorphic after chlorite flakes. However, in the Missi series, white mica is abundant at the garnet isograd, evidently greatly in excess of the amount necessary to produce biotite from all the chlorite; furthermore, chlorite is absent in all of the sections collected in the biotite zone along the garnet isograd. Again, in other rocks to be discussed later, chlorite, though it remains in excess far into the biotite zone, reacts with epidote to form amphibole and no garnets appear. There is a distinct possibility that epidote, biotite, and magnetite play a part in the formation of garnet, but until more information is obtained, the precise mode of origin of this mineral must remain unexplained.

Retrogressive metamorphism in the garnet zone.

The minerals in the chlorite and biotite zones are all clear and undeformed; evidently their crystallization took place after movements in those zones had ceased. Early minerals of the garnet zone have been subjected to post-crystallization movements; in the extreme northern and western parts of the sedimentary area the last movements were severe enough to produce some retrogression of high-rank minerals to those of lower rank. Staurolite, in addition to the minerals listed above in the garnet zone, is of sporadic occurrence in this part of the area. It appears in anhedral crystals and elongate fragments strewn out parallel to the schistosity. Groups of the fragments in optical continuity with one another are surrounded and separated from each other by a mat of pale-green chlorite; evidently the groups of fragments were once parts of single large crystals of staurolite, a relatively high-rank mineral that has now been changed in part to chlorite. A few of the garnets have thin chlorite rims, but on the whole this mineral proved more resistant to the tendency to retrogress than did staurolite. Other indications of late movements include cracking and granulation of quartz ribbons, development of cleavage and cloudiness in oligoclase, and bending and breaking of biotite as well as muscovite plates. It is of interest to note that although both staurolite and garnet, and possibly some of the biotite, reverted to chlorite, the plagioclase did not alter, but remained as elsewhere in this zone, about An_{30} in composition.

METAMORPHISM OF QUARTZ DIORITE.

Coarse-grained plutonic rocks of medium composition intrude both the Missi series and the underlying volcanics at several points in the area. Although they all are metamorphosed in a similar manner, the tabular body of quartz diorite intrusive into greenstones east of Cliff lake, with its dike continuations along Willow creek west of Embury lake is best suited for careful examination. Exposures are excellent, and it is possible to walk on continuous outcrops and trace every gradation of change throughout a length of over four miles. The largest part of the intrusive lies in the chlorite zone as defined from the metamorphic condition of greywacke; the dikes along Willow creek extend far into the biotite zone; other discontinuous dikes of somewhat similar composition outcrop north of Willow creek, well within the garnet zone.

Character in the chlorite zone.

The main body of quartz diorite exposed east of Cliff lake is in the chlorite zone. Here it is a coarse grained granitic type of rock that shows little evidence of shearing. Estimated from the hand specimen, it is made up of about 25 per cent quartz, 40 to 50 per cent feldspar, and the remainder of chloritized mafic minerals.

In thin section the rock is seen to be made up of quartz, albite, (An_4), abundant chlorite and epidote, some calcite, scattered crystals of apatite and pyrite, and a very little microcline. The albite crystals are hypidiomorphic, with clear rims and centers altered to aggregates of granular epidote, white mica, and albite. Probably the original feldspar was zoned, with relatively pure albite rims that have remained unaltered, and more calcic centers that broke down into the aggregate described. Quartz is interstitial to albite. It has strong undulose extinction, but although some of the albite crystals are bent and fractured, the rock is in no sense intimately sheared.

Microcline is present in small, clear, unaltered grains in the groundmass and in thin, more or less continuous rims around albite crystals. Tongues of microcline project between and through crystals of albite and seem to replace them to some extent. The manner of occurrence, and the presence of calcite and pyrite suggest that the rock may have been somewhat altered by late magmatic solutions. However, the total amount of these minerals is very small, too small to have any effect on the subsequent course of metamorphism; they can be neglected in the discussion.

The original character of the rock cannot be stated exactly. Judged from the large amount of epidote and the zoned appearance of feldspar, the original plagioclase was probably much more calcic than albite. The extremely small content of microcline, even if it be admitted as primary, and the absence of white mica except as saussuritic intergrowth with albite, indicates an original low potassium content. In short, the rock was probably originally composed of quartz, moderately calcic plagioclase, and mafic minerals, and was therefore a quartz diorite. As now constituted, it contains the mineral assemblage appropriate to the chlorite zone, that is, epidote, chlorite, albite, and quartz.

Character in the biotite zone.

Traced northward, the quartz diorite remains quite massive as far as the portage between Cliff and Embury lakes. There it passes into the biotite zone of the greywackes; from there on the foliation is well developed, and the quartz diorite grades into a quartz diorite gneiss. The quartz is drawn out into thin ribbons less than 1 mm. thick and 2 cm. or more long; on weathered surfaces the ribbons stand up in little blades parallel to the foliation.

Under the microscope other notable differences between these rocks and those farther south are apparent. The minerals are all clear and the dust-like particles and swarms of mica flakes that clouded the albite crystals in the chlorite zone have completely disappeared. The original igneous texture is obliterated and the rock is made up of a granoblastic aggregate of approximately equidimensional grains of quartz and albite (An_{4-5}), needles of greenish-blue amphibole, a little brown biotite, epidote, some calcite, and accessory apatite, tourmaline, and titanite. The albite tends to be idioblastic and sporadic crystals are of porphyroblastic dimensions. The amphibole needles and larger prisms are mostly oriented parallel to one another in the plane of schistosity, but many of them lie across or at random in the section, proving that they crystallized after movements ceased. Biotite, in contrast to greywackes in the same zone, is very minor in amount.

Character in the garnet zone.

A large dike of quartz diorite gneiss intrusive into greenstones north of Willow creek is well within the garnet zone of the greywackes. In the hand specimen it looks like typical quartz diorite gneiss, but microscopic examination shows the rock to be considerably more basic than the quartz diorite farther east. It is composed of oligoclase (An_{20}), quartz, greenish-blue amphibole, epidote, chlorite, rare grains of zircon, titanite, and iron ores. Evidence of post-crystallization movement is clear; the chlorite and amphibole crystals are bent into sharp crenulations, and the feldspar, instead of being clear and uncleaved as formerly, is cloudy and badly fractured. The movements may have been sufficient to cause retrogression of amphibole to chlorite and epidote, both of which are more abundant than is expectable in this zone. However, the plagioclase is somewhat less calcic than that in the adjacent greywackes, suggesting that this gneiss either was not so sensitive to metamorphism, or that the dike was not intruded until after the metamorphic processes had passed their climax. The later explanation is borne out to some extent by the tectonic and intrusive history of the region, but a full consideration of this point is outside the purpose of this paper.

METAMORPHISM OF FELDSPAR PORPHYRY.

The Missi series is cut by over 40 dikes of feldspar porphyry. The dikes are of all thickness up to 50 feet, with an average of perhaps 20 feet; several were traced along their strike for over 1,000 feet. They were intruded parallel to the schistosity for the most part, but in some places they cut across the structure with a zig-zag course that suggests partial control of intrusion by joints.

The porphyry, except where greatly metamorphosed, is one of the most easily recognized rocks in the district. Euhedral phenocrysts of pinkish to white albite that range from 2 mm. to 1 cm. in length, make up 20 to 25 per cent of the rock. On fresh surfaces they show up in striking contrast to the dark grey to black, fine-grained or aphanitic groundmass. The total plagioclase content, including that of the groundmass, averages over 50 per cent of the whole. Quartz, though it forms 20 to 25 per cent of the rock, occurs in phenocrysts in only two dikes; in the others it remained in the groundmass. Potassium feldspar is absent.

Character in the chlorite zone.

The euhedral crystals of albite (An_{3-4}) are well twinned according to the albite law. They are clouded with tiny flakes of white mica, granules of epidote, and dust-like particles. The groundmass is composed of quartz, albite, abundant epidote and chlorite, some calcite, and rare grains of apatite, titanite, and pyrite. The rock is considerably sheared, and contrary to the general rule in this zone, the chlorite flakes are much deformed. Late hydrothermal action is suggested by the fact that calcite cuts across and seems to replace some of the albite phenocrysts, and by the occurrence of cubes of pyrite, not only in the porphyry, but also disseminated for several feet through the adjacent greywackes.

Character in the biotite zone.

The phenocrysts of the porphyry, though locally sheared out almost beyond recognition, remain intact on the whole through the biotite zone. The groundmass, however, is much darker on fresh surface, due to the development of megascopically recognizable needles of amphibole and flakes of biotite.

In thin section, the phenocrysts are seen to be albite (An_{4-5}), well twinned according to the albite law. The groundmass contains quartz, acicular greenish-blue amphibole, abundant epidote, chlorite, some white mica, and accessory apatite, zircon, titanite, rutile, and rare flakes of allanite.

Knots and clusters of dark-green biotite flakes with decussate arrangement occur sporadically in some of the dikes. The clusters, generally accompanied by calcite and exceptionally abundant apatite, range up to 2 mm. in diameter. They embay or occupy whole centers of feldspar crystals, or fill large areas

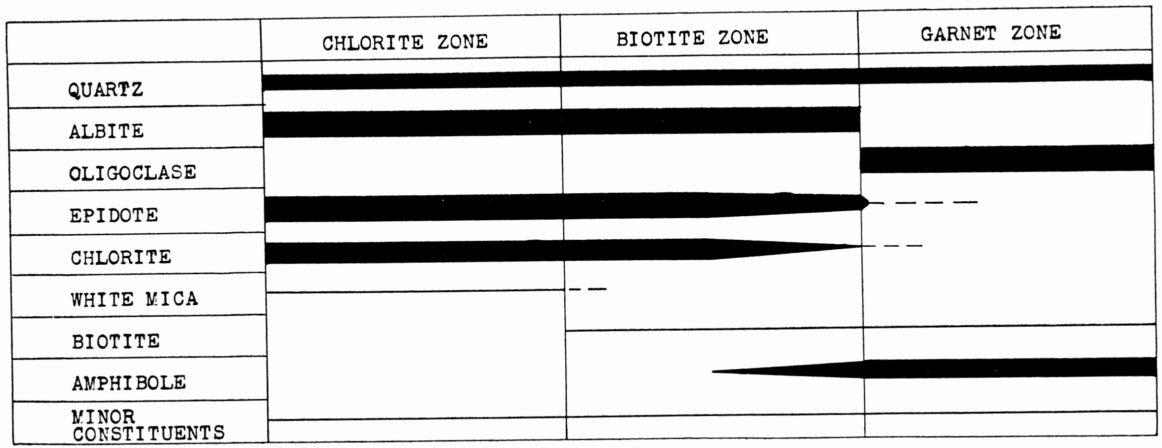

Fig. 3. Diagrammatic representation of mineral paragenesis in progressively metamorphosed quartz diorite and feldspar porphyry near Flinflon, Manitoba.

in the groundmass to the partial or complete exclusion of other constituents. Although nearly all the feldspar phenocrysts are somewhat cracked and broken, the biotite flakes are undeformed. Replacement relationships and the nested arrangement of biotite are phenomena foreign to kinetically metamorphosed rocks. Apparently biotite, calcite, pyrite, and perhaps apatite were introduced by post-shearing magmatic gases or solutions, and are not to be considered as normal products in the progressive kinetic metamorphism of these porphyries.

Character in the garnet zone.

Gradations in petrographic character from the biotite zone to the garnet zone, though abrupt, are complete. With advancing metamorphism, the schistosity becomes more and more pronounced, and the phenocrysts, so prominent in the lower rank rocks, disappear completely. Obliteration of the porphyritic texture is brought about, not only by decrease in size of the phenocrysts, but also by an increase in grain size of the groundmass minerals to such an extent that the rock in the garnet zone is essentially equigranular.

The feldspar in a specimen collected in the narrow transition belt between the biotite and garnet zones is albite-oligoclase (An_{11}); in specimens collected well within the garnet zone it is oligoclase-andesine (An_{28-32}). The crystals tend to be idioblastic. Albite twins are very scarce, but pericline twins are common; cleavage is poor. Quartz, abundant greenish-blue amphibole, some biotite, a very little epidote and chlorite, apatite, and grains of iron ore complete the mineral assemblage.

Mineral Paragenesis.

IN QUARTZ DIORITE AND FELDSPAR PORPHYRY.

The mineral paragenesis of these rocks, diagrammatically illustrated in the accompanying figure, although similar in some respects to that of the greywackes, has some striking and important differences. The main mineral constituents in the chlorite zone are albite, quartz, epidote, and chlorite. White mica is very scarce or absent, and this deficiency is the main factor in producing the differences in mineral composition in comparison to the greywacke with advancing metamorphism. All the white mica is used up to produce biotite in the lowest grades of the biotite zone, and the excess chlorite, ten to fifteen per cent of the rock, persists far into this zone. However, in intermediate grade rocks, a new mineral, greenish-blue amphibole, makes its appearance, and as it increases in amount, chlorite and epidote decrease. At the garnet isograd, no chlorite is left, and the rock enters the garnet zone as an quartz-albite-epidote-amphibole schist.

The amphibole is a distinctive variety, apparently identical with one described by Rice (1935, pp. 307-309). Its most characteristic features are strong greenish-blue pleochroism parallel to Z, and absorption Y>Z>X. Pleochroism parallel to Y is deep olive-green and parallel to X, pale yellow. The analyses given by Rice show, as he has pointed out, the unusual relationships $Fe_2O_3 > FeO > Al_2O_3$ in one type, and $Fe_2O_3 > Al_2O_3 > FeO$ in another. The pleochroism is increased slightly in the garnet zone, but what change in composition this may indicate, if any, is unknown. Apparently it is formed by a reaction between chlorite and epidote that begins about the middle of the biotite zone and is practically completed, by exhaustion of the chlorite, in the vicinity of the garnet isograd.

The plagioclase in these rocks undergoes changes identical with those in the greywackes. From albite in the biotite zone it changes abruptly as before to oligoclase-andesine in the garnet zone. No micas are present to supply additional alumina to form anorthite from all the calcium of epidote, and it may be that here too some calcium is freed, in this case to enter into the composition of the amphibole. At any rate, epidote diminishes in amount and virtually disappears from the suite within the lower grades of the garnet zone, and the rock becomes a quartz-oligoclase-amphibole schist.

METAMORPHISM OF WEKUSKOAN GREENSTONES.

The Wekuskoan volcanics, the oldest rocks in the area, are metamorphosed in a manner similar to the overlying Missi series. However, since the volcanic group includes several different types of rock, it is impossible to be certain that specimens collected from the different zones had the same composition, thus making it impossible to discover the sequence of mineral changes that may have occurred during metamorphism. In spite of this, the fact that the mineral suites developed in the more basic volcanics in the various zones are both characteristic and predictable is perhaps of sufficient interest to merit brief description.

Character in the chlorite zone.

The greenstones in the chlorite zone as defined from the greywackes are mostly massive rocks, with local zones of intense shearing. The massive rocks are composed principally of albite, colorless to very pale-green amphibole, chlorite, epidote, and some calcite. Specimens of basic dikes intrusive into the greenstones in this part of the area have well preserved ophitic texture, although all the minerals are secondary. The rocks have suffered complete degradation, and are in equilibrium in the zone of their occurrence.

Character in the biotite zone.

A specimen of dark-green volcanic rock collected at the north end of Cliff lake, i.e., just within the biotite zone of the greywackes, is made up of chlorite, epidote, calcite, albite, quartz, and biotite, named in order of decreasing abundance. White mica is absent; if there was any in the rock before metamorphism, it has all been used up to produce the minor amount of biotite.

Basic volcanics that have reached a more advanced stage of metamorphism in this zone, indicated by much better development of the schistosity, contain porphyroblasts of greenish-blue amphibole in addition to the minerals listed above. This amphibole is identical to that formed in quartz diorite gneiss and feldspar porphyry in the biotite zone.

Character in the garnet zone.

The volcanics in the garnet zone have been altered to crenulated schists. The crenulations are sharp folds, in general less than an inch in height from crest to trough; cleavage faces are glistening greenish black.

In thin section, one of the darker schists is seen to be made up mainly of a felt of tiny greenish-blue amphibole needles that sweep around the crenulations in perfect "flow" lines; some of the larger crystals, that lie across the lengths of the small needles, are slightly broken, but on the whole it is evident that they were formed after movements ceased. Small undeformed flakes of brown biotite and grains of quartz and plagioclase make up the remainder of the rock. The plagioclase grains, clear, untwinned, and practically devoid of cleavage, are too small for accurate determination, but are more calcic than albite. Epidote and chlorite are absent. In short, the rock is a quartz-plagioclase-biotite-amphibole schist. It would correspond to a quartz-albite-sericite-chlorite rock in the chlorite zone, with chlorite in excess over sericite; to a quartz-albite-biotite-chlorite-epidote schist in the low grades of the biotite zone; and to a quartz-albite-epidote-biotite-amphibole schist in the higher grades of the same zone.

METAMORPHISM OF THE CONGLOMERATES.

Metamorphism of the conglomerates produced drastic changes in the pebbles, both with regard to their shapes and their mineral composition. The changes in shape are described in a later section; with regard to the changes in mineral composition, a word here will suffice. All the pebbles examined from the conglomerates within the chlorite zone, including quartzites, felsites, quartz-feldspar porphyry, feldspar porphyry, and granitic types are made up, excluding minor constituents, of quartz, white mica, chlorite, epidote, and albite, in different proportions according to the original composition of the rocks. A granitic pebble collected from the more advanced portion of the biotite zone is made up of quartz, greenish-blue amphibole, epidote, and albite; other pebbles show the same degree of adjustment. In short, the present mineral assemblages of the pebbles correspond exactly to the zone in which they occur. It is a convincing illustration of the perfect adaptation of a variety of rocks to the pressure-temperature environment during metamorphism.

CAUSE OF METAMORPHISM.

The mineralogical changes recorded in all the different rocks described above, for example the formation of oligoclase from albite and epidote, could only have been brought about by increases in pressure and temperature.

Elevation of the temperature might be brought about by intrusion of an igneous body, by deeper burial of the more highly metamorphosed portions, or by an accumulation of mechanical heat developed by and during shearing. None of the igneous bodies now exposed in the area can be regarded as probable sources of heat, for not only are the highest rank rocks those which are farthest from the large intrusives, but all the intrusive bodies in the region are themselves kinetically metamorphosed. Furthermore, with the exception of a very few late, small bodies of ultrabasic rocks, the degree of their metamorphism depends solely upon their location relative to the zones. The heat may have been supplied by an igneous mass not yet exposed, but neither the outline of the zones nor the type of metamorphism favors such a possibility.

Depth of burial cannot be called upon to account for the rise in temperature in these rocks. The base of the Missi series is exposed in the chlorite zone in the southern, in the biotite zone in the middle, and in the garnet zone in the northern parts of the area. Roughly speaking, the metamorphic rank along any section perpendicular to the main fold axis is the same throughout, whereas a series of such sections from south to northwest show a progressive rise in rank. It is apparent that metamorphism took place after the series was folded, and there is no structural evidence whatever that the northern end of the fold was ever more deeply buried than the southern.

Frictional heat must be generated on planes of shear when differential movements take place. If the shearing is restricted to that along faults, the relatively small amount of heat generated, though possibly of local importance, is obviously insufficient to cause any significant rise in the temperature of a large body of rocks. On the other hand, if the rocks are intimately sheared, although the amount of frictional heat developed along any given plane may be small, the total amount must be enormous. Provided it is generated faster than it can be conducted away, the temperature of the whole body of sheared rock must rise.

Evidence of internal shearing in greywackes and conglomerates.

All the pebbles, with the exception of those of granitic rocks, in the conglomerates and scattered through the greywacke over the whole area are greatly elongated. In the chlorite and biotite zones they are drawn out into triaxial ellipsoids, with the greatest and least diameters in the planes of schistosity. The dimensions of the average pebble in the conglomerate east of Little Cliff lake are approximately 18 x 2 x 1 inches; in general the greatest diameter may be eight or nine times as long as the intermediate; the intermediate diameter is ordinarily not over four times the least and the two may be nearly equal. Granitic pebbles remained practically undeformed in these zones; they are well rounded, and the smaller ones are almost perfect spheres. The diameter

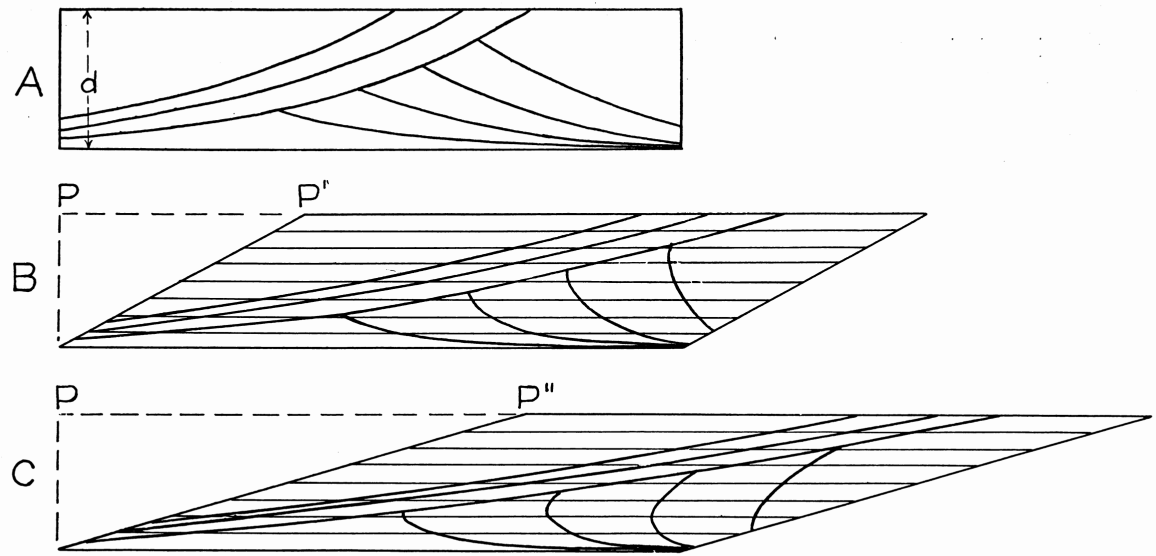

Fig. 4: A: Normal cross-bedding. B: Cross-bedding deformed by uniform shear, P-P′, the integration of a number of small movements on thin lamellae. Movement P-P′ is sufficient to deform a sphere of diameter 'd' into an ellipsoid with a major axis equal to 3xd. C: Cross-bedding deformed by shear, P-P″, sufficient to deform a sphere of diameter 'd' into an ellipsoid with a major axis equal to 4xd.

NOTE: Maximum deformation of cross-bedding will occur when the direction of movement is parallel to the direction of the current, and will approach zero when the two are perpendicular. Other variations are introduced if the planes of shear cut across the main bedding planes, but the net effect on the shape of the cross-bedding will be the same, regardless of the angle of intersection.

of a sphere of volume equal to the average elongated pebble is 3.3 inches. A pebble of feldspar porphyry that was completely freed from the matrix measures, along its principal axes, 16 x 1½ x 1½ inches; the diameter of a sphere of equal volume is 2.3 inches. If it can be assumed that all the pebbles were as well rounded before deformation as the granitic now are, it is apparent that they have been stretched four or five times their original diameter. Simultaneously, parallel to the intermediate and least axes of the ellipsoids, the original diameter decreased about one-third and two-thirds respectively.

The major axes of the elongated pebbles show a remarkable uniformity of orientation, not only with respect to one another within one bed of conglomerate, but throughout all the conglomerates and including the pebbles scattered through the greywackes as well (Fig. 1). It goes without saying that the pebbles could only have been elongated by the action of tremendous pressures, and the uniformity of orientation proves that these pressures were part of a single process that was operative over the whole area. The precise mechanism by which the pebbles were elongated is an intriguing question that cannot as yet be answered, but obviously pressures were sufficient to produce plastic deformation in the chlorite zone at temperatures not high enough to raise the greywackes above the greenschist facies.

As the pebbles were being elongated, the greywackes too were undergoing important mechanical changes. The sedimentary rocks are extensively cross-bedded, and the fine-grained greywacke interbedded with conglomerates whose pebbles are stretched as described has as well preserved cross-bedding as any in the area. However, the cross-bedding, though well enough preserved to leave no doubt as to the tops of the beds, has clearly been deformed.

In ordinary cross-bedding, the curving beds are concave up, asymptotic to the lower continuous bedding plane, and steepened upwards to a slope of 35° or less to meet the upper transecting bedding plane (Fig. 4, A). Cross-bedding in the Missi series is rarely of this simple form. In some places the cross-beds have been steepened to vertical or even overturned; in others they have been flattened practically to a straight line. Where two sets of cross-beds face in opposite directions as exposed in the outcrop, a not uncommon case, the effect of steepening and flattening is quite striking (Fig. 4, B and C).

Simple compression perpendicular to the bedding will produce flattening of the cross-bedding, and shortening of the beds in the plane of bedding will produce steepening, but if either of these processes had acted alone, all the cross-bedding would be affected in the same way, regardless of which direction it faced. Where both flattening and steepening occur together, as in Fig. 4, B, and as observed repeatedly in the field, there seems to be no escape from the conclusion that the deformation is the result of small differential movements all in the same direction on thin lamellae. In other words, differential movements took place on every plane of schistosity in the greywackes; furthermore, as proved by the pebbles, the whole series was affected by the same type of movement.

Heat by mechanical deformation.

The amount of heat that will be generated by friction between two surfaces depends upon the coefficient of friction, the pressure normal to the surfaces, and the amount of movement. The rate at which heat will be generated in any given case depends on the rate of movement, and the temperature will rise only if heat is generated faster than it can be conducted away. Therefore, when rocks over a considerable area are intimately sheared, assuming a uniform coefficient of friction and rate of conduction throughout, if any of the remaining factors is increased in one part of the area while the others are constant, the temperature in that part will rise correspondingly higher.

Little can be said regarding the comparative rates of movement from one part of the area to another; it was slow enough everywhere to allow the pebbles to flow rather than fracture. The amount of movement and probably the normal pressure were both greater in the biotite zone than the chlorite zone, as shown by the progressive milling out of the quartz grains and the perfection of the schistose cleavage. Within the garnet zone there can be little doubt that both normal pressure and the amount of movement were much greater than in either of the other zones. The high-rank rocks lie entirely within the important west-striking structural belt described in the first part of the paper, within which the schistosity is highly developed, and the schistose planes are sharply crenulated and drag-folded. Pebbles are drawn out to such incredible lengths[3] that they must be seen to be believed. In addition, whereas they are invariably straight in the biotite and chlorite zones (Fig. 5, A) here they are twisted and contorted into amazing drag folds (Fig. 5, B). Recognizable cross-bedding is entirely obliterated and only long straight or highly

Fig. 5A: Deformed conglomerate, Missi Series. Dark elongated pebbles are of chloritic schists; light colored elongated pebbles are of more salic fine-grained igneous rocks. Two round pebbles, one in the extreme middle right and the other eight inches to the left of this, are coarse-grained quartz-bearing phanerites. The six inch pencil gives the scale.

contorted lines of magnetite grains remain to show that it once existed. With such profound and thoroughgoing mechanical deformation, much greater amounts of heat must have been generated and the rocks raised to a correspondingly higher temperature in the garnet zone than in either of the other two zones. There are no other visible sources of heat. Therefore, it is reasonable to suppose that frictional heat alone

[3] Pebbles of felsite and greenstone 40 inches long and one to two inches wide are common. One apparently continuous pebble of a light colored volcanic measured over 70 inches long by one-half inch wide.

raised the temperature sufficiently to allow evolution of the high-rank minerals.

Growth of most of the minerals did not take place until after shearing ceased. The sequence of events during metamorphism is thus, intimate shearing of the rocks with generation of heat and elevation of the temperature of the rock body to a degree corresponding to the amount of shearing; practical cessation of movements over most of the area; crystallization of most of the minerals. Post-crystallization movements are restricted to the garnet zone; there they deformed some of the early minerals, and where most severe, produced partial retrogression of staurolite, garnet, and possibly biotite to chlorite.

Fig. 5B: Contorted conglomerate within the northern, intensely deformed belt. An elongated and contorted pebble may be distinguished near the right side of the picture. Thinning along the limbs of some of the folds is evident. The overlying contorted curly quartz-mica schist, derived from greywackes, is exposed in the upper left. The direction of drag folding indicates that the north (bottom) side moved west (right) with respect to the south.

Irregular undeformed cross-cutting patches of white quartz are later than the folding, and appear to have replaced the conglomerate to some extent.

SUMMARY.

The early Precambrian rocks near Flinflon, including the Wekuskoan volcanics, the Missi series of greywackes and conglomerates, and quartz diorite and feldspar porphyry intrusive into the older rocks, have been kinetically metamorphosed. The degree of metamorphism increases from south to north and reaches its highest rank in a belt across the northern part of the area where the rocks have been bent from a north-south strike to an east-west strike; the bend was caused by regional movements of the northern rocks towards the west relative to those on the south.

There zones of metamorphic intensity—the chlorite, biotite, and garnet zones—are recognized in the Missi greywackes. In the chlorite zone the greywackes are composed of quartz, albite, epidote, chlorite, and white mica; in the biotite zone they become quartz-albite-epidote-mica schists, and in the garnet zone, garnetiferous quartz-oligoclase-mica schists. Quartz diorite and feldspar porphyry in the chlorite zone are composed of quartz, albite, chlorite, and epidote. The smallness of the white mica content, in contrast to the greywackes, produces important differences in the mineral paragenesis, for in the biotite zone these rocks become quartz-albite-epidote-amphibole schists, and these in turn, in the garnet zone, become quartz-oligoclase-amphibole schists. The Wekuskoan volcanics and the pebbles of the conglomerates, regardless of their character, have a mineral composition that is in complete equilibrium within the zone of their occurrence.

The most highly metamorphosed rocks are those farthest removed from the large igneous bodies of the area; furthermore, all the igneous rocks, with the exception of small, late, ultrabasic and aplite dikes, are metamorphosed to the same degree as the other rocks. Clearly they did not supply the necessary heat. There is no structural evidence whatever to indicate that the high-rank rocks were ever more deeply buried than those of lower rank. However, all the rocks have been intimately sheared, and the rocks of highest metamorphic rank, i.e., those within the garnet zone, lie in a belt of most intense shearing and deformation. It seems reasonable to suppose, therefore, that the metamorphism was purely kinetic; that is, it was brought about by heat and pressures developed by and during intimate regional shearing.

ACKNOWLEDGEMENTS.

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