PALEOECOLOGY OF THE CAMBRIAN AND ORDOVICIAN STRATA OF MINNESOTA

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Minnesota was the site of at least three marine transgressions during Cambrian and Ordovician times. The epeiric sea was generally confined to the southern and southeastern parts of the state, with the shoreline trending southwestward from an area north of the Twin Cities to the southwestern corner of the state. Sediments accumulated in a shallow depression, which rapidly shoaled to the north, between the northeast-trending Transcontinental Arch in Minnesota and the northwest-trending Wisconsin Arch (figure VI-22). This sedimentary basin has been called the Hollandale embayment of the Ancestral Forest City basin (Austin, 1970b). Although the position of the shoreline during Late Cambrian time undoubtedly varied, isopach maps of the Upper Cambrian strata (Berg and others, 1956; Ostrom, 1964; Austin, this volume) suggest that the maximum transgressive shoreline was roughly parallel to the present boundary of the Paleozoic rocks.

Probably, the sea remained in Minnesota within the Hollandale embayment continuously from Late Cambrian through Early Ordovician time, and then retreated. The sea returned during the Middle and Late Ordovician, and at this time also covered extensive parts of northwestern Minnesota, and possibly the entire state, as the seas encroached on either side of the Transcontinental Arch (figure VI-22).

The thicknesses and to some extent the character of the Ordovician formations were affected by growing intrabasinal flexures such as the Twin City basin and the Rochester-Red Wing anticline.

ST. CROIXAN SERIES

Cambrian sedimentary rocks in Minnesota were deposited during two major transgressive-regressive cycles. During the first cycle, the Dresbachian sequence, including the Mt. Simon Sandstone, the Eau Claire Formation, and the Galesville Sandstone, was deposited; during the second, the Franconian-Trempealeauan sequence, including the Ironton Sandstone, the Franconia and St. Lawrence formations, and the Jordan Sandstone, was laid down (figure VI-23).

Nearly all the sedimentary rocks consist of clastic material, with four principal lithotopes being repeated many times (modified after Berg and others, 1956; Austin, this volume). These consist of a (1) coarse-grained lithotope of cross-bedded, fine- to coarse-grained sandstone, (2) fine clastic lithotope of thin-bedded, fine-grained to shaly sandstone and siltstone, (3) greensand lithotope of fine-grained, moderately glauconitic worm-burrowed sandstone, and (4) sandy dolomite lithotope.

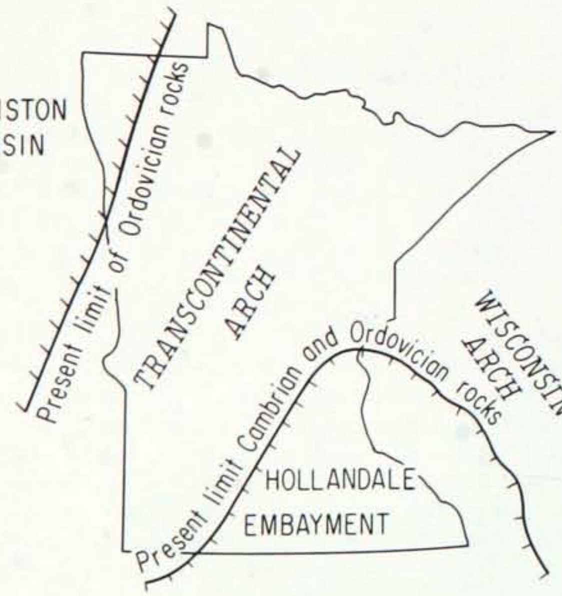


Figure VI-22. Regional setting of Paleozoic rocks in southeastern and northwestern Minnesota.

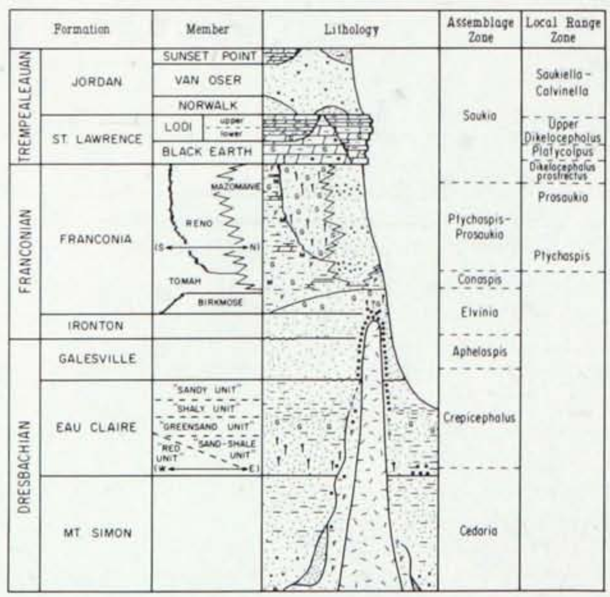


Figure VI-23. Croixan Series in southeastern Minnesota (modified from Austin, 1969).

The area inundated by the transgressing sea was one of low relief. A series of low bluffs composed of Upper Precambrian basalt formed islands near Taylors Falls, along the northeastern shoreline of the Hollandale embayment. These islands were not completely inundated until Franconian time. Islands also were present in the Baraboo region of Wisconsin, on the flanks of the Wisconsin Arch.

Fossil remains generally are not abundant in rocks of Late Cambrian age. Bottom communities consist mostly of trilobites, inarticulate brachiopods, and burrowing soft-bodied organisms that presumably were annelids. Inarticulate brachiopods are most abundant numerically, but are represented by only a few species. Trilobites show great diversity of form and are abundant within certain beds. However, even so-called fossiliferous formations commonly show gaps of tens of feet where fossils are rare or absent. Examination of a section of Upper Cambrian sedimentary rocks, which shows both columnar sections and fossil zones (Berg and others, 1956, figure 5), indicates that the fossil zones are rare and comprise about one bed per forty feet of strata.

THE DRESBACHIAN SEQUENCE

The shallow epicontinental seas that flooded Minnesota in Late Cambrian time initially deposited the Mt. Simon Sandstone over mostly Upper Precambrian clastic sedimentary rocks. The Mt. Simon typically is composed of white or gray, medium-grained, cross-bedded quartzose sandstone. Inarticulate brachiopods are the most common fossils, and their macerated remains in the coarser, cross-bedded units indicate a nearshore, high-energy marine environment of normal salinity. Scattered interbeds of fine-grained sandstone and shale are present locally, and indicate a quieter offshore environment. The transgressive-regressive coarse-grained sandstone lithotopes in the St. Croixan Series appear to have rather restricted bottom communities. Trilobite communities representing the *Cedaria* Assemblage Zone* are present but not abundant, and are incompletely known.

The Eau Claire Formation marks the maximum transgression of the Dresbachian sequence and includes several clastic rock types. The most atypical of these is a red, silty, fine-grained sandstone and red shale that is found near the western border of the Hollandale embayment. The relatively silty units are commonly worm-burrowed, and the entire section may represent normal marine offshore conditions where sedimentation was sufficiently rapid to prevent reduction of the hematitic pigment. The bulk of the Eau Claire Formation consists of fine- to medium-grained quartzose sandstone with interbeds of green shale and glauconitic, fine- to medium-grained sandstone. The Eau Claire is the most fossiliferous unit of the Dresbachian sequence, and contains inarticulate brachiopods and worm burrows as well as a diversity of trilobites. The Eau Claire is characterized by the *Crepicephalus* Assemblage Zone, although the zone actually continues into the bottom of the Galesville Sandstone. Trilobites are invariably disarticulated and crowd the bedding surfaces at many intervals. Over 30 species have been described from the *Crepicephalus* Assemblage Zone.

The Galesville Sandstone, representing the regressive phase of the Dresbachian sequence, is a white, cross-bedded, medium-grained quartzose sandstone similar to the Mt. Simon, except that it has somewhat less shale, is finer grained, and generally is better sorted. The sandstone at the base of the Galesville is moderately well-sorted and becomes well-sorted at the top (Austin, 1970b, and this volume). The Galesville and Mt. Simon Sandstones represent similar depositional environments. Fossils are sparse and trilobites predominate. Most of the Galesville is in the *Aphelaspis* Assemblage Zone although *Aphelaspis* itself has not been reported in Minnesota. The top of the *Aphelaspis* Assemblage Zone is probably identical to the top of the Galesville that is marked by an unconformity; the *Dunderbergia* Assemblage Zone, described from a continuously-evolving fauna in western United States (Palmer, 1965; Lochman and Wilson, 1958) is absent in Minnesota. The unconformity at the base of the Galesville Sandstone in Wisconsin is absent in Minnesota.

FRANCONIAN-TREMPEALEAUAN SEQUENCE

The second Upper Cambrian transgressive-regressive cycle was not greatly different from the first. A shallow epicontinental sea again transgressed an area of low relief. Again, islands of Upper Precambrian basalt were present at Taylors Falls although they eventually were inundated during this interval. The transgressive Ironton Sandstone, however, records a lower energy environment than that of its older analog, the Mt. Simon Sandstone of the Dresbachian sequence. The Ironton is not as well sorted and the proportion of silt is significantly higher. The Ironton is a white, medium-grained, well- to poorly-sorted quartzose sandstone that has a significant amount of admixed silt. The fauna is dominated by trilobites of the *Elvinia* Assemblage Zone, but is neither diverse nor abundant. Apparently, life was sparse in the nearshore phase of sedimentation represented by the Ironton. In the Taylors Falls area, coarse conglomerates of Ironton age have yielded an unusual molluscan fauna that includes monoplacophoran species. These fossils are found in sandstone pockets among coarse basalt boulders (as large as two feet in diameter), and lived in an intertidal environment at the shoreline. Included among the monoplacophorans are hypseloconids, which are high-coned, septate, and probably representative of the group from which cephalopods evolved.

The Franconia Formation was deposited in a wide variety of sedimentary environments that varied from shallow littoral to offshore marine. The Birkmose, the lowermost member, is a glauconitic, worm-burrowed, fine-grained sandstone representing a widespread offshore lithotope. It is characterized by trilobites of the *Elvinia* Assemblage Zone. The Mazomanie, Reno, and Tomah Members represent progressively basinward rocks that are laterally equivalent facies rather than vertical divisions, although the Reno overlies the Tomah where the two are found together in the same stratigraphic section (Austin, this volume). The Mazomanie Member, a thin- or cross-bedded, dolomitic, fine- to coarse-grained quartzose sandstone, is present in the

* Fossil Zones are shown in Figure VI-23.

northern part of the Hollandale embayment, and represents a shallow, perhaps sublittoral environment. The Mazomanie interfingers with and replaces both the Reno and Tomah Members in a shoreward direction. The Reno Member, a fine-grained, glauconitic, worm-burrowed, quartzose sandstone, is rather similar to the Birkmose Member, and probably represents deposition under similar environmental conditions. The Tomah Member, a very fine-grained silty sandstone with interbedded micaceous shale and minor dolomite probably represents the deepest depositional environment during the time of Franconian deposition. The Tomah thickens basinward and constitutes the entire Franconia in the center of the Hollandale embayment, in southeastern Minnesota (Austin, this volume).

The Franconia Formation is subdivided on the basis of trilobites, which constitute part of the North American Standard for the Late Cambrian, into several fossil zones. The trilobites are both diverse and abundant at several intervals. It is not known whether the upper trilobite zones, which are based primarily on the fauna in the Reno Member, extend or continue into the shoreward facies of the Mazomanie or into the center of the Hollandale embayment, where the Tomah Member probably constitutes the entire section.

The St. Lawrence Formation overlies the Franconia and is composed of sandy dolomite, dolomitic siltstone, and fine-grained dolomitic sandstone. Dolomite content decreases, whereas the clastic content and its grain size increase, as the formation thins toward the northeast. Trilobites of the *Saukia* Assemblage Zone dominate the fauna; inarticulate brachiopods and dendritic graptolites comprise the remainder of the fauna. The Black Earth Member, which represents an offshore depositional environment, is the lowermost member of the St. Lawrence Formation in Minnesota and is composed almost entirely of dolomite. The upper Lodi Member consists of dolomitic siltstones and sandstones, and comprises the entire formation in the nearshore areas to the east.

The Jordan Sandstone, the regressive phase of the Franconian-Trempealeauan sequence, and probably also the transgressive phase for Lower Ordovician sedimentation in Minnesota, is a white to yellow, fine- to coarse-grained quartzose sandstone. The Van Oser Member (see fig. VI-2) constitutes the bulk of the Jordan Sandstone and is a white to yellow, fine- to coarse-grained quartzose sandstone. It represents a shallow, nearshore, high-energy environment, and is generally unfossiliferous. The Norwalk and the Sunset Point Members are finer grained offshore equivalents of the Van Oser. Bottom communities of the Jordan are poorly known. Trilobites of the *Saukia* Assemblage Zone characterize the meager fauna.

CANADIAN SERIES

The Lower Ordovician Series in Minnesota is composed of a thick succession of predominantly dolomitic rocks, and represents a relatively thin segment of a very extensive dolomitic sequence found in central, southern, and eastern United States (fig. VI-24). The strata in this series are referred to as the Prairie du Chien Group, which is divided into the Oneota Dolomite and the Shakopee Formation;

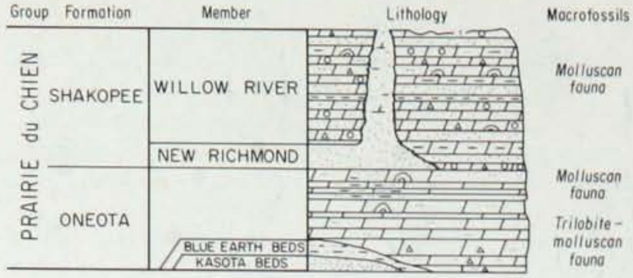


Figure VI-24. Lower Ordovician Series in Minnesota (modified from Austin, 1969).

these formations are further subdivided into several formal members and informal beds of restricted geographic extent (Austin, this volume).

Although several investigators indicate an unconformity at the base of the Lower Ordovician Series (Graham, 1933; Powell, 1935; Stauffer, 1925), recent studies have shown that the systematic boundary was the site of continuous sedimentation marked only by the gradual deposition of carbonate (Sardeson, 1936; Kraft, 1956; Berg and others, 1956). Indeed, only at Stillwater and Vermillion, Minnesota, is the contact between Cambrian and Ordovician sediments considered "sharp," and even here there is some gradation. The environment during the Early Ordovician in Minnesota was one of a shallow marine epicontinental sea, with extensive development of carbonate banks. Commonly, these banks were dotted with algal stromatolites, which formed biohermal structures. Shoaling is indicated by oolitic chert and dolomite, and periodic exposure is indicated by mudcracks and desiccated algal structures. Large, low-amplitude sand waves composed of cross-bedded dolo-arenites possibly represent deposits formed by tidal currents. Oscillation and current ripples and flat-pebble conglomerates are common.

Occasionally, the carbonate banks were affected on a regional scale by moderate to extreme quartz-sand sedimentation. This influx of sand resulted locally in the deposition of sandy dolomites and supermature quartz-sandstone beds as much as several feet thick. At places, the rapid influx of quartz sand inundated and killed the algal flora involved in reef building. The New Richmond Sandstone Member of the Shakopee Formation overlies an unconformity and represents a regionally important tongue of sandstone from nearby source areas to the east, west, and north.

Davis (1966b) has compared the environment of the Willow River Member of the Shakopee Formation with modern algal reefs of Shark Bay, Australia, as described by Logan and others (1964). Davis considered the Willow River environment to be one of a warm epeiric sea, with both shallow marine and intertidal regimes extensively represented. The three environmental regimes he recognized are: (1) a shallow, open marine area with oolitic beds and relatively abundant fossils; (2) an intertidal or near intertidal, high-energy zone where stromatolitic algae flourished, as represented by algal bioliths; and (3) a locally intertidal and hypersaline, low-energy zone represented by algal mats and abundant desiccation features.

A predominantly molluscan fauna is associated with the algal stromatolites. Gastropods are the most abundant and cephalopods rank second. Monoplacophorans, pelecypods, and crinoids are present but not abundant. Brachiopods and trilobites are rare in the lower part of the Oneota Dolomite and are generally absent throughout the remainder of the Prairie du Chien Group. Three somewhat overlapping faunas are found in the Prairie du Chien: a lower trilobite-molluscan and an upper molluscan fauna in the Oneota, and a molluscan fauna in the Willow River Member of the Shakopee. Simple-cone conodonts occur sporadically throughout the Prairie du Chien and have been described by Furnish (1938).

The Prairie du Chien Group is sparingly fossiliferous and preservation is generally poor. Internal molds are the most common form of preservation. Chert nodules crowded with well-preserved specimens are described by Stauffer (1937 a and b) from the Shakopee Formation. The fauna contains monoplacophorans, gastropods, cephalopods, minor pelmatozoan fragments, and a trilobite. Except for rare crinoids, bottom-dwelling filter-feeders are absent. Perhaps more important, the fauna is dwarfed; larger equivalents of the same species are found elsewhere in the group. Apparently, conditions were far from optimum for normal marine life, and some environments may have been relatively barren of life. These conditions may represent hypersaline environments. Because of the predominance of gastropods, the virtual lack of filter-feeders, and occasional dwarfed faunas, I do not believe that abundant organic remains were destroyed by penecontemporaneous dolomitization. Rather, I think that the environment was too rigorous to support a diverse or abundant fauna.

CHAMPLAINIAN AND CINCINNATIAN SERIES

It is not known whether the seas retreated from the Hollandale embayment after the deposition of the Prairie du Chien Group. The presence of an unconformity between the Canadian and Champlainian Series in Wisconsin and elsewhere cannot be demonstrated at the rather poor exposures of the contact of these series in Minnesota. Rather, the contact appears to be conformable, but weathering has obscured critical evidence (fig. VI-25). Regardless, sedimentation in the Hollandale embayment was probably continuous from the deposition of the St. Peter Sandstone in early Middle Ordovician (Chazyan) time to the deposition of the Maquoketa Formation in early Late Ordovician time. In northwestern Minnesota, sedimentation began somewhat later, with the deposition of the Winnipeg Formation in Middle Ordovician time (Black Riveran), and ended with the deposition of the Red Rock Formation along the eastern edge of the Williston basin in Late Ordovician time. Rocks of Cambrian age are not known in this area, and the Ordovician strata lie directly on the Precambrian basement.

Southeastern Minnesota

St. Peter Sandstone

The St. Peter Sandstone is an atypical formation in many ways, including its uniformity of grain size, mineralogy, and sorting, its lack of sedimentary structures, and its

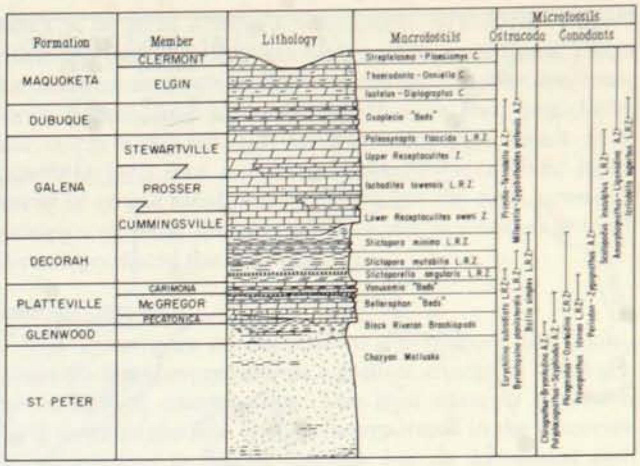


Figure VI-25. Champlainian and Cincinnatian Series in Minnesota (modified from Austin, 1969).

wide areal extent. Dapples (1955) estimated the present areal extent of the St. Peter Sandstone to be 225,000 square miles. In Minnesota, the formation averages about 80 feet thick and, as elsewhere, consists of medium- to fine-grained sand, more than 99 percent of which is quartz. Sorting is typically very high, resulting in a general lack of sedimentary structures. The sand grains are well rounded and the larger grains show a frosted and pitted surface. Normal and cross-bedding are known but rather rare. Shale is very rare and reported only in the subsurface of the Twin City basin.

Dapples (1955) interpreted the St. Peter Sandstone as having been deposited in an extensive area of low relief by a sea that gradually inundated the area from the southeast toward the northwest. The absence of shale is attributed by him to shoreline currents moving the fine clastics far to the southern and southwestern parts of the basin. He suggested that toward the end of St. Peter deposition the transgressing shorelines covered the Wisconsin and Transcontinental Arches.

That the St. Peter was indeed deposited by a shallow marine environment is indicated by the sparse fauna recovered from the Twin City region. The fauna is entirely molluscan, consisting of pelecypods, gastropods, and cephalopods. Possibly, pitting and frosting of the sand grains indicate the presence of shoreline dunes that were reworked by the transgressing sea.

Glenwood Formation

The Glenwood Formation, a thin unit of argillaceous sand and shale, marks the transition from the shoreward environment of the St. Peter Sandstone to the offshore carbonate bank environment of the Platteville Formation, as the sea continued its westward transgression onto the Transcontinental Arch. In Minnesota, the unit ranges in thickness from about 2 to 16 feet and averages about 5 feet. The lower boundary of the formation is difficult to define as there is only a slight gradational change in sediment size from the St. Peter into the Glenwood. The lower part of the Glenwood is sandy and barren of fossils. The upper part is

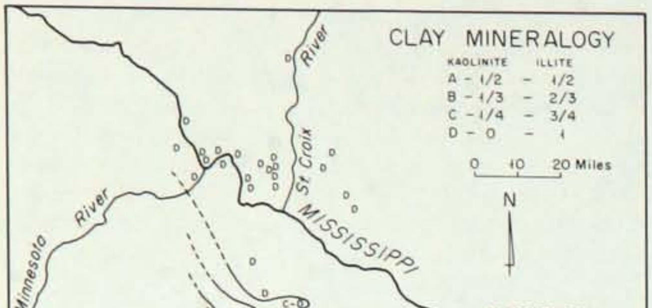


Figure VI-26. Relative abundance and distribution of kaolinite and illite in the Glenwood Formation (modified by Parham from Parham and Austin, 1967).

easily recognizable in that it consists of gray-green shales overlain by shaly sandstone. These shales are generally barren of macrofossils but yield abundant conodonts of the *Chirognathus-Bryantodina* Assemblage Zone.

Variations in the clay mineralogy of the Glenwood Formation indicate a source area to the southwest or west, presumably on the Transcontinental Arch (Parham and Austin, 1967). A geographic plot of the clay mineral variations, based primarily on the relative abundance of kaolinite and illite, is shown in Figure VI-26. Figure VI-26 shows belts of clay facies which presumably follow the shoreline in a general way. As might be expected, the general orientation is northwestward, paralleling the Transcontinental Arch. In detail, however, the facies pattern shows marked departures from a linear array. The marked northeasterly trend is modified by an east-northeast projection north and east of Faribault and by a second more subdued projection somewhat to the southeast. These two features are separated by an area characterized by a strong, broad, westerly shift in facies pattern. It is interesting to speculate on the possible significance of these variations. One might imagine that they indicate actual shoreline features, including well-developed headlands separated by a wide bay. More probably, however, these eastward projections were areas that received an abnormally large supply of fine clastic material, whereas the intervening area received sparse amounts of clastic material. Perhaps major river systems on the Transcontinental Arch entered the sea to the west of the promontories. An isopach map of the Glenwood lends support to this idea; the thicker parts of the formation coincide with the eastward-trending projections. Near Cannon Falls, Minnesota, the Glenwood is more than three times as thick as the average for the formation (Parham, oral comm.).

Judged from the character of the Decorah, Galena, Dubuque, and Maquoketa formations, the western source area of the Transcontinental Arch probably continued to supply clastics to southeastern Minnesota throughout the remainder of Ordovician time.

Platteville Formation

The Platteville Formation, which overlies the Glenwood Formation and is about 30 feet thick in Minnesota, is dominantly a carbonate unit comprised mainly of limestone. Dolomite is common, however, especially in the lowermost member, and shales alternate with the limestone beds in the upper parts of the formation. In the same way as the St. Peter Sandstone, the Platteville is a thin, widespread unit throughout the midwest.

A shallow marine carbonate bank environment existed during deposition of the Platteville Formation. The first widespread bottom communities of a sessile benthonic nature developed in this environment. These communities were dominated by articulate brachiopods, but also included diverse other invertebrates, especially mollusks. Species of *Pionodema*, *Protozyga*, and *Strophomena* are the most important of the articulate brachiopods, and commonly form coquinoid layers within the limestone beds. Cephalopods reached their greatest diversity and were most abundant during this time. They range in size from the one-inch *Zitteloceras* to the large *Endoceras*, which grew to lengths of as much as 15 feet. Gastropods, especially the bellerophontids, also were quite common in the formation.

The rate of development of the sessile-dominated bottom communities was slow. The lowermost Pecatonica Member, like most other dolomites or dolomitic limestones in the Ordovician of Minnesota, records a relatively low development of bottom communities. However, these communities were well established during the deposition of the remaining members, and they persisted, with minor exceptions, in the younger Middle and Upper Ordovician strata of Minnesota, and changed only by the gradual replacement of individual species. Only in the Decorah Shale and in the Stewartville Member of the Galena Formation do we find significant differences.

Seafloor conditions were generally stable and quiet in Platteville time. Scattered coquinoid layers within beds attest to periodic intervals of gentle current activity. Although crinoids, brachiopods, and other invertebrates are commonly disarticulated, they do not show evidence of abrasion. The seafloor itself must have been carpeted by a soft calcareous ooze. The large (as much as 10 inches in diameter) endoceroid cephalopods sank in the ooze to depths of two-thirds of their shell diameters. The exposed parts of the shells apparently dissolved before deposition of the next bed, inasmuch as the tops of the cephalopod shells are planed off at the upper bedding surface. Thus, long time intervals probably are recorded by the thin clay partings between carbonate beds.

In the Carimona, the uppermost member of the Platteville, limestone beds alternate with shales, and ultimately, these beds pass into strata assigned to the Decorah Shale.

The widespread addition of bryozoans to the bottom communities characterizes the shale beds. The addition of abundant fine-grained clastic material to the upper part of the Platteville Formation marks an uplift of the Transcontinental Arch in western Minnesota, and is a precursor to deposition of the Decorah Shale.

Local variations in the environmental regime of the Platteville Formation can be distinguished. For example, where the formation is only 12 feet thick in the Cannon Falls area, specimens of *Lingula* occur upright in their burrows in the Carimona Member. However, at this locality the Glenwood Formation is as much as 16 feet thick, and may represent a large, local clastic supply.

Decorah Shale

The Decorah Shale is characterized by gray-green shale and scattered, thin coquinoid limestone beds. It is as much as 80 feet thick in the Twin City region, but thins both eastward and southeastward; in southeastern Minnesota it is only about 45 feet thick. On the basis of variations in clay mineralogy, Parham and Austin (1969) have shown that the Decorah detritus was derived from a westerly or southwesterly source area, presumably the Transcontinental Arch.

The Decorah Shale was deposited in a shallow, nearshore marine environment, and the diversity of species indicates that the waters were warm and of normal marine salinity. Nearly optimum conditions for marine life must have prevailed, inasmuch as a large number of marine macro- and microorganisms, as represented by nearly all phyla of marine invertebrates found in the Ordovician of Minnesota, attain maximum abundance and diversity in the Decorah Shale.

Bottom communities are dominated by bryozoans and brachiopods, and include species of *Rhynidictya, Pionodema, Batostoma, Hallopora, Strophomena,* and *Sowerbyella* as well as many others. The fossil remains are typically broken or disarticulated but lack size sorting, and thus probably are representative of a biocoenose. Microfossils are abundantly represented by ostracodes, conodonts, scolecodonts, and chitinozoans.

The bottom itself must have been a soft muddy ooze. As in the Platteville Formation, large endoceroid cephalopods settled to two-thirds of their shell diameter, leaving the upper parts exposed. Also indicative of soft bottom conditions are species of the bryozoan *Prasopora*. These individuals initiated colonies on bits of debris on the bed surface, and then to obtain maximum support, developed low, flat-bottomed colonies. Although one might expect highly turbid conditions resulting from an influx of fine clastic material from nearby source areas, the high percentage of filter-feeders among the marine invertebrates indicates relatively clear water.

Except for periods of episodic high-energy conditions, the environment was relatively quiet. Layers of coquinoid limestone as much as 2 inches thick are common in the Decorah Shale, and represent winnowings of the bottom material by currents and/or waves. Weiss (1957) believed they are caused by large amplitude storm waves. These coquinoid layers can be found at some places with ripple marks as large as 2 feet in wavelength. Bryozoan colonies within the coquinas indicate high energy conditions. Twig-like colonies as much as half an inch in diameter are commonly broken into segments 2 inches or less long. Periodic storm waves should, however, not only result in a coquinoid layer but also in a relatively thick shale layer devoid of coarse fossil debris. This is not the case, however. Perhaps a periodic shoaling of the water or changes in tidal currents produced the coquinas.

Galena Formation

The lower part of the Galena Formation, the Cummingsville Member, represents a gradual change in lithology and in bottom communities. The land mass to the west which supplied the fine clastics incorporated in the Decorah Shale continued to supply detritus, but on a reduced and intermittent basis, and the alternating limestones and shales of the Cummingsville Member gave way gradually upward to limestones of the Prosser Member. As the clastic supply diminished, the bottom environment became rather quiet and a carbonate bank was established. Evidence of current activity is lacking, and the fossils in the fine-grained limestones are neither abraded nor broken.

Bryozoans, including species of *Prasopora, Batostoma,* and *Rhynidictya,* are well established in the shales and shaly limestones in the lowermost beds of the Cummingsville. Apparently adapted to the muddy substrate, they gradually disappeared from the bottom communities and were virtually absent by the beginning of Prosser sedimentation. The problematic *Receptaculites* is a common member of the bottom communities, which are dominated in upper Cummingsville and Prosser strata by articulate brachiopods. Some of the more common brachiopods include species of *Sowerbyella, Resserella, Rafinesquina,* and, to a lesser extent, species of *Parastrophina* and *Plectorthis.* Conodonts record an abrupt change in faunal composition at the Decorah-Galena contact. No less than 15 form species of conodonts which ranged throughout the Decorah Shale are absent in the lowermost limestone of the Galena. A possible climatic change is indicated by the gradual disappearance of the "midcontinent" conodont fauna and the appearance of species characteristic of the Appalachian and Scandinavian faunas. This trend continues throughout Galena time, and reaches a maximum in the alternating limestones and shales of the Dubuque Formation. The upper Dubuque and the Maquoketa Formation record the gradual return of the midcontinent fauna. Assuming that the Ordovician North Pole was to the west of North America, the relationship between the two faunas would be north-south rather than the present east-west. With this relationship, a temperature change seems a likely possibility.

The Prosser Member is predominantly limestone, which records a quiet offshore carbonate bank environment. Bottom communities continued to be dominated by filter-feeders, and indicate quiet, clear water of normal marine salinity.

Strata of the Stewartville Member record a profound environmental change. Bottom communities consisting of brachiopods and other filter-feeders are nearly absent, and they are replaced by faunas dominated by gastropods and cephalopods. The large gastropod *Maclurites* dominates the fauna; *Receptaculites* is also a common constituent. The environment may have been one of a shallow carbonate bank with restricted circulation. Hypersaline conditions in such an environment would account for the lack of filter-feeders and would limit the sparse populations to those with higher ecological valences—the cephalopods and gastropods. The presence of *Climactograptus* in the highly dolomitic strata of the Stewartville tends to support this possibility. Although not poorly preserved, the conodont faunas also record adverse conditions inasmuch as the fauna is sparse in numbers and in diversity of forms. The overall lithology, faunal diversity, and composition is similar to that of the Prairie du Chien Group.

Dubuque Formation

The Dubuque Formation, which averages about 35 feet in thickness in Minnesota, consists of intercalated, buff, medium-grained limestones and gray shales. The interlayered limestone and shale beds appear to indicate cyclical sedimentation. Two feldspathized bentonites are present, and aid in tracing individual limestone beds for many miles. The limestone beds thicken to the west as a result of increased amounts of fine clastic detritus within individual beds. The interbedded shales also thicken to the west, as they change from a dominantly calcilutite to an argillaceous shale. The cyclical nature of the limestones and shales as yet is not adequately explained. The Transcontinental Arch apparently was slightly uplifted at the beginning of deposition of the Dubuque Formation.

Bottom communities in the Dubuque Formation are dominated by filter-feeders, and differ markedly from the restricted faunas found in the Stewartville Member of the Galena Formation. However, this change was gradual. The lowermost Dubuque beds are highly dolomitic and sparsely fossiliferous. Conodonts are the most common fossils and even they are sparse. Limestones with abundant pelmatozoan remains are found slightly higher in the formation and filter-feeder bottom communities become well established in still higher beds. These filter-feeding bottom communities are dominated by brachiopods, and with some evolutionary modification are not unlike those of the Prosser Member of the Galena Formation. Common brachiopod species include *Resserella corpulenta* and *Sowerbyella minnesotensis*. Microfossils include abundant conodonts, characteristic of Appalachian and Scandinavian faunal areas, and ostracodes. These bottom communities indicate a return to normal marine conditions with open circulation.

Maquoketa Formation

The transition from the Dubuque Formation to the Maquoketa Formation is marked by a decrease in fine clastics and an increase in dolomitic strata, apparently without any break in sedimentation.

Environments represented by Maquoketa strata are complex and Bayer (1965, unpub. Ph.D. thesis, Univ. Minn.) recognized four lithosomes in Minnesota, each of which is associated with variations in bottom communities. The phosphatic "depauperate" bed, widespread in Iowa, is absent. A possible local equivalent is found in the lower 10 to 15 feet of strata at the base of the formation near Granger, Minnesota. Filter-feeding organisms are not represented in the meager fauna consisting of abundant graptolites and trilobites, with minor amounts of conodonts and cephalopods. Bayer referred to this fauna as the *Isotelus-Diplograptus* Community, and considers it indicative of periodic stagnation in an offshore area with associated toxic bottom conditions. The abundant concentrations of organic material and the fetid odor of some of the strata within this interval support his conclusion.

Most of the Maquoketa Formation is composed of alternating limestone, shaly dolomite, and dolomitic limestone. The limestone is typically sublithographic and sufficiently fossiliferous to form shell beds at many intervals. Filter-feeding organisms dominate the fauna. Most common are articulate brachiopods, with less abundant pelecypods, cephalopods, and graptolites, and a few gastropods and crinoids. Bayer referred to this as the *Thaerodonta-Onniella* Community, after two of the most common brachiopods. The fossiliferous limestone alternates abruptly with barren shaly dolomite or dolomitic limestone. Such abrupt changes in lithology and fossil content would argue for a sudden change from a normal marine environment with open circulation to one with restricted circulation and hypersaline waters. Bayer considered the strata to be cyclic, and prefers periodic epeirogenic oscillations to account for the observed variations in fossils and lithology. With this explanation, the dolomite would represent shallow near-shore conditions. The higher percentage of argillaceous clastic material in the dolomites supports this conclusion.

A local and thin facies of the above unit contains an atypical fossil assemblage. The bottom community is almost entirely composed of a single species of rugose coral—*Streptelasma corniculum*. A few brachiopods and sparse crinoids complete the fauna of the *Streptelasma-Plaesiomys* Community. The rugose corals are typically oriented and are surrounded by a matrix of fossil debris. Lithologies within the strata containing the corals are dominantly dolomitized limestones. Crude cross-bedding is present in some of the beds. Possibly, the deposition was in a shallow marine environment near or at wave base.

The beds in the upper part of the Maquoketa Formation are barren of fossils, with the exception of conodonts, and consist of sandy and shaly dolomitic limestone. Coarseness and abundance of clastic material in the strata increase in a westerly direction. Apparently the major source of sediments for all the Ordovician rocks in Minnesota was from the west, where gentle epeirogenic uplifts periodically increased the influx of clastic material. Apparently, the Wisconsin Arch served as a barrier to clastic deposits from the east that were associated with the Taconic orogeny.

Northwestern Minnesota

Sedimentation in northwestern Minnesota began in Champlainian time (probably late Chazyan) and continued into the Cincinnatian. These sedimentary rocks mark the eastern featheredge of the Williston basin. Natural exposures of these rocks are lacking in Minnesota, and the data given here are based on rotary drill cuttings from a few holes. Bayer (1959, unpublished M.S. thesis, Univ. Minn.) has described the available subsurface information. Probably none of the drill holes contains a complete section of the Ordovician rocks. One of the most complete drill logs is given below.

Summary of Florance No. 1 well
(Modified after Bayer, 1959, *op. cit.*)

Location: Sec. 6, T. 162 N., R. 49 W., Kittson County, Minnesota

Glacial Drift

Red River Formation	Cathead Member—135 ft. Buff Limestone
	Doghead Member—100 ft. Gray-brown limestone, lower 81 ft. argillaceous
Winnipeg Formation	70 ft. Sandstone, white, medium-grained, friable, thin shales in upper 10 ft.
	85 ft. Shale, gray-green to brown, minor limestone beds
	10 ft. Sandstone, white, medium-grained, friable

Hole Bottomed in Precambrian Schist

The sedimentary rocks penetrated in the Florance No. 1 well are believed to correlate with the Winnipeg and Red River Formations of Manitoba, as indicated above. Correlations are based on lithologic criteria, as paleontologic information is lacking.

Winnipeg Formation

The Winnipeg Formation is a clastic unit that has a maximum recorded thickness in Minnesota of 185 feet and a range in thickness from 135 to 185 feet. The lowermost white friable sandstone marks the beginning of a transgressive sequence. This is overlain by gray-green fissile shale with minor amounts of dense, coarse-grained, argillaceous limestone near the base. A few conodont faunules have been recovered from the shales. The upper part of the Winnipeg is a white, medium-grained, friable sandstone, not unlike the lowermost 10 feet; the two intercepts probably represent a similar depositional environment. The upper sandstone unit marks a regressive phase of sedimentation, and probably correlates with similar sandstone in North Dakota, South Dakota, and Manitoba. Shale beds in the upper part of the sandstone mark a second transgressive phase. The regressive-transgressive nature of the sandstone is probably a result of a slight uplift of the Transcontinental Arch, followed by either subsidence or erosional leveling.

Red River Formation

In Minnesota, the Red River Formation is a carbonate unit having a maximum thickness of 235 feet. The true thickness of the formation is greater, however, inasmuch as the top of the section showing maximum thickness is an erosional surface. Bayer (1959, *op. cit.*) correlated these carbonate units with the Doghead and the Cathead Members of the Red River Formation of Manitoba. The lower 100 feet, correlated with the Doghead Member, consists of dark gray to purple, silty and argillaceous limestones that have a microgranular texture. The lower part of the unit is more dolomitic than the upper. The contact with the underlying Winnipeg is gradational, and the overlying Cathead Member indicates a continuation of the transgressive sequence initiated during deposition of the upper Winnipeg sandstones. The upper 135 feet, correlated with the Cathead Member, consists of yellow to buff, slightly dolomitic limestones that have a sucrosic texture. This unit is indicative of an offshore carbonate bank receiving little clastic material, and represents a continuation of the transgressive sequence. It is not known whether the younger Selkirk Member of the Red River Formation or the Stony Mountain Formation were deposited in Minnesota.

Sequences of Sedimentary Environments and Correlation of the Ordovician Strata in Northwestern and Southeastern Minnesota

Primarily on the basis of faunas reported from adjacent regions of Manitoba, North Dakota, and South Dakota, Bayer (1959, *op. cit.*) correlated the Ordovician strata of northwestern Minnesota with the St. Peter-to-Galena sequence in southeastern Minnesota (table VI-1). I concur with this correlation, and believe it can be documented. The two areas have had a similar tectonic history, which is reflected in comparable sequences of sedimentary rocks.

Although the two basins lie on opposite sides of the Transcontinental Arch and their eroded edges are presently as much as 300 miles apart at the closest point (figure VI-22), the erosionally truncated edge of the sedimentary rocks of northwestern Minnesota is almost 400 feet thick and accordingly the maximum transgressive shoreline must have been considerably farther southeast. A similar thickness of sedimentary rocks occurs at the eroded edge of the Hollandale embayment, and the maximum transgressive shoreline of this basin must have been considerably farther northwest. In Early Paleozoic time, the Transcontinental Arch must have been very narrow or possibly even inundated as the transgressive seas lapped in from either side. Regardless, any tectonic activity associated with the arch must have affected both basins and produced similar sequences of sedimentary environments.

The first and most important event that affected both basins occurred at the time the Middle Ordovician seas transgressed midwestern United States. This transgression yielded a single major transgressive sequence of Middle and Upper Ordovician strata in southeastern and northwestern Minnesota. Superimposed on this regional pattern are periodic epeirogenic fluctuations of the Transcontinental Arch. During the initial transgression, the St. Peter Sandstone, the Glenwood Formation, and the lower part of the Platteville Formation were deposited in southeastern Minnesota, and the lower Winnipeg sandstone and the Winnipeg shales were deposited in northwestern Minnesota (figure VI-27). Evidence that the above strata correlate across the Transcontinental Arch is given by the meager conodont fauna recovered by Bayer (1959, *op. cit.*) from the Winnipeg shales. These conodonts were identified by William Furnish of the University of Iowa, and tentatively corre-

Table VI-1 Correlation of Middle and Upper Ordovician sedimentary rocks of northwestern and southeastern Minnesota.

Epeirogenic Activity Transcontinental Arch	Sedimentary Sequence	Rock Types	Northwestern Minnesota	Southeastern Minnesota
Mild uplift	Regressive	Sandy Carbonates	—	Maquoketa Fm Clermont Mbr
Mild subsidence	Transgressive	Carbonates	—	Maquoketa Fm Elgin Mbr
Mild uplift	Regressive	Limestones & Shales	—	Dubuque Fm
Mild subsidence	Transgressive	"Clean" Carbonates	Red River Fm Cathead & upper Doghead Mbr	Galena Fm Prosser & Stewartville Mbr
		Shaly Carbonates	Red River Fm lower Doghead Mbr	Galena Fm Cummingsville Mbr
Uplift	Regressive	Sandstone & Limestone	Winnipeg Fm upper sandstones	Decorah Shale upper Platteville
Subsidence	Transgressive	Limestone & Shale	Winnipeg Fm Shale beds	Lower Platteville & Glenwood shales
		Sandstone	Winnipeg Fm lower sandstone	Upper St. Peter Sandstone

lated with conodont faunas in the Glenwood-Platteville-Decorah sequence in southeastern Minnesota. Subsequently, a more comprehensive study of the conodont faunas of southeastern Minnesota has been completed (Webers, 1966), and it can be shown that the Winnipeg shales correlated with the Glenwood Formation and the lower half of the Platteville Formation. It is inferred that later uplift of the Transcontinental Arch increased the amount of clastic detritus entering the basins, which was deposited under regressive conditions. In southeastern Minnesota, this is recorded as shale interbeds in the upper Platteville Formation, and as uplift increased, by thick shale intervals in the Decorah Shale. Similarly in northwestern Minnesota, uplift and increased clastic deposition are recorded by the sandstones in the upper part of the Winnipeg Formation. Continued erosion and/or subsidence of the arch ultimately resulted in a diminished clastic influx, as recorded in southeastern Minnesota by the shaly limestones of the Cummingsville Member of the Galena Formation, and in northwestern Minnesota by the shaly limestones in the lower part of the Doghead Member of the Red River Formation. Continued subsidence of the arch and a concomitant lack of clastic material is recorded in the pure carbonates of the Prosser and Stewartville Members of the Galena Formation in southeastern Minnesota and in the pure carbonates of the Upper Doghead and Cathead Members of the Red River Formation in northwestern Minnesota. Either the Selkirk Member of the Red River Formation and the overlying Stony Mountain Formation of Manitoba were eroded or were never deposited in Minnesota. In southeastern Minnesota, the Dubuque Formation and the Clermont Member of the Maquoketa Formation represent short periods of renewed mild uplift of the arch.

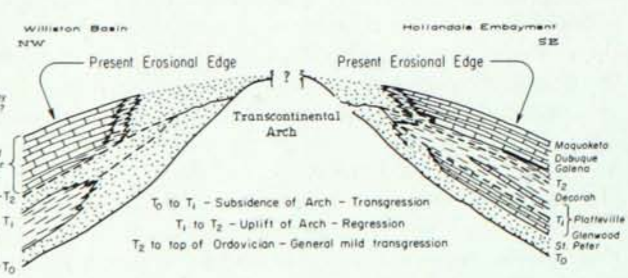


Figure VI-27. Diagrammatic sketch of the correlation of the Middle and Upper Ordovician formations in southeastern and northwestern Minnesota.

DIVERSITY OF SPECIES IN THE ORDOVICIAN OF MINNESOTA

The Ordovician strata of Minnesota reflect a variety of environments, which are indicated by marked variations in species diversity of the major invertebrate groups. Data presented here summarizing the faunal composition of the various stratigraphic units were taken principally from the work of Stauffer and Thiel (1941). Although several faunal studies have been completed since this publication, they are either detailed studies of only a part of the Ordovician strata or are restricted to a particular fossil group. Undoubtedly, the species names used in Stauffer and Thiel are out of date, but the relative number of species recognized by them in each of the various invertebrate groups probably will not change markedly with additional work.

The major groups of invertebrates selected for this study of species diversity include trilobites, gastropods and monoplacophorans, cephalopods, pelecypods, bryozoans and brachiopods. Information on these groups is tabulated in Figures VI-28-31.

It can be seen from Figure VI-28 that there is considerable variation in the total number of species between given stratigraphic intervals. Nearly optimum environmental conditions must have prevailed during Platteville through Prosser deposition, inasmuch as these rocks are characterized by a maximum in the total number of species as well as a maximum in the diversity of species (figure VI-29). Most maxima occur in the Cummingsville and Prosser Members of the Galena Formation. The stratigraphic interval between the Platteville Formation and Prosser Member is marked by limestone and shale that probably reflect deposition in warm shallow waters that had normal marine salinity. In general, the abundance of individuals on bedding plane surfaces also closely follows their diversity. With the exception of the Dubuque Formation, which consists of limestone and shale, all the remaining stratigraphic intervals having low species diversity have dominant sandstone or dolomite lithologies. Inasmuch as the limestones within the Dubuque Formation are commonly criquinas, I believe further investigation would reveal a much greater diversity of species than shown in Figure VI-29. In contrast, dolomites or highly dolomitic limestones within the section are almost devoid of bottom organisms. However, the general lack of both micro- and macrofossils appears not to result from destruction of the fossils by dolomitization, but rather to reflect the rigorous nature of the environment. There is good evidence that the environment was probably hypersaline at the time of deposition of the Prairie du Chien Group, and this was probably also true at the time of de-

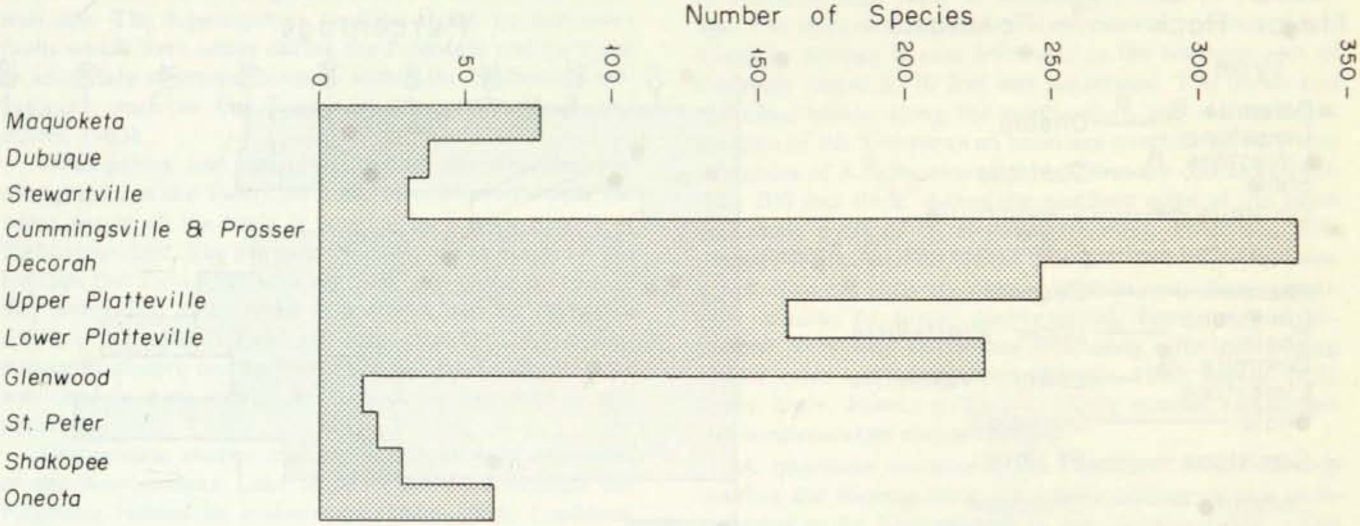


Figure VI-28. Total number of invertebrate macrofossil species in the Ordovician of Minnesota.

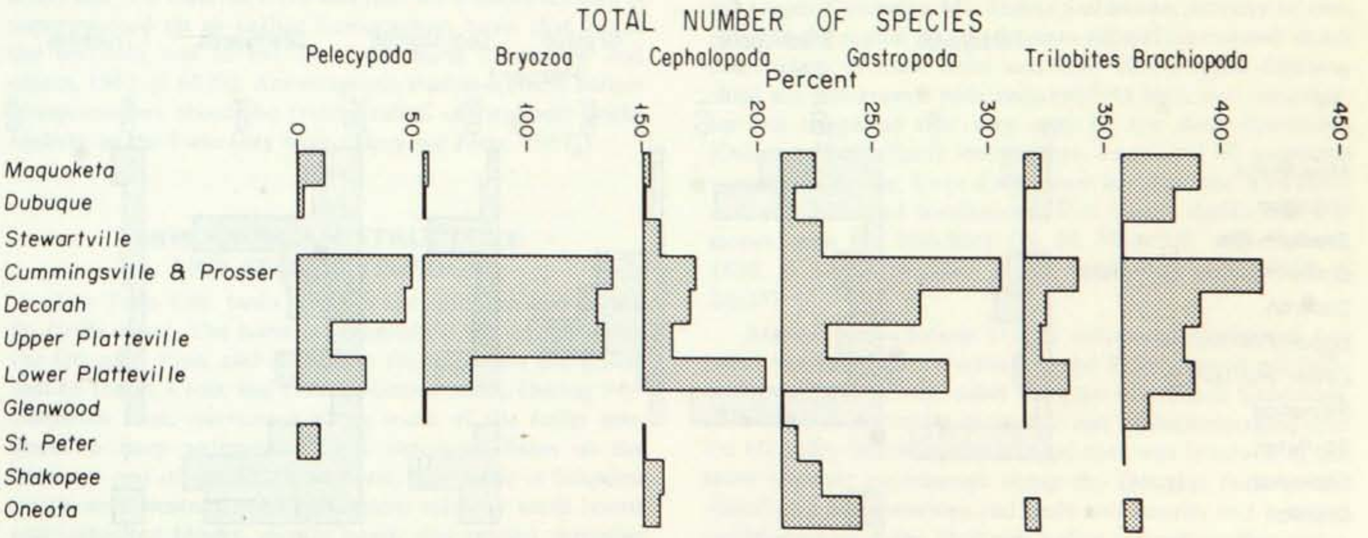


Figure VI-29. Number of species of major phyla in the Ordovician of Minnesota.

position of the Stewartville Member of the Galena Formation. For example the conodonts show a continual decline in diversity and abundance in the Galena Formation, and this is especially marked in the Stewartville Member (Webers, 1966). However, the conodonts were not destroyed through recrystallization in the dolomites; they simply were sparse to begin with.

It appears that only those organisms with wide ecological tolerances were able to tolerate the environment represented by the dolomites. Gastropods are known for their high tolerance of diverse conditions. A study of the fauna of the Permian Reef Complex (Newell and others, 1953) indicated that the fauna of the hypersaline backreef environment was dominated by gastropods. Virtually the only other organisms in this environment were various kinds of blue-green algae. Filter-feeding organisms, especially attached forms, were practically non existent inasmuch as these tend to be stenohaline in nature. This ecological picture generally fits the faunal picture of the dolomitic environments of the Minnesota Ordovician. From Figure VI-30 it can be seen that filter-feeding organisms dominated the fauna at all stratigraphic intervals in the Ordovician of Minnesota except those that are marked by abundant dolomite.

Figure VI-31 indicates that there is a marked variation in ecological tolerance among major groups of marine invertebrates. Bryozoans appear to be most restricted and to have the narrowest tolerances. These are followed in order by the brachiopods and pelecypods. The non filter-feeders, whether bottom dwelling or not, appear to have the greatest tolerance for changes in environment. This is especially true of gastropods and cephalopods, whose patterns of diversity and abundance are similar. One might have predicted the pattern of the gastropods, but their similarity to the cephalopods is quite surprising.

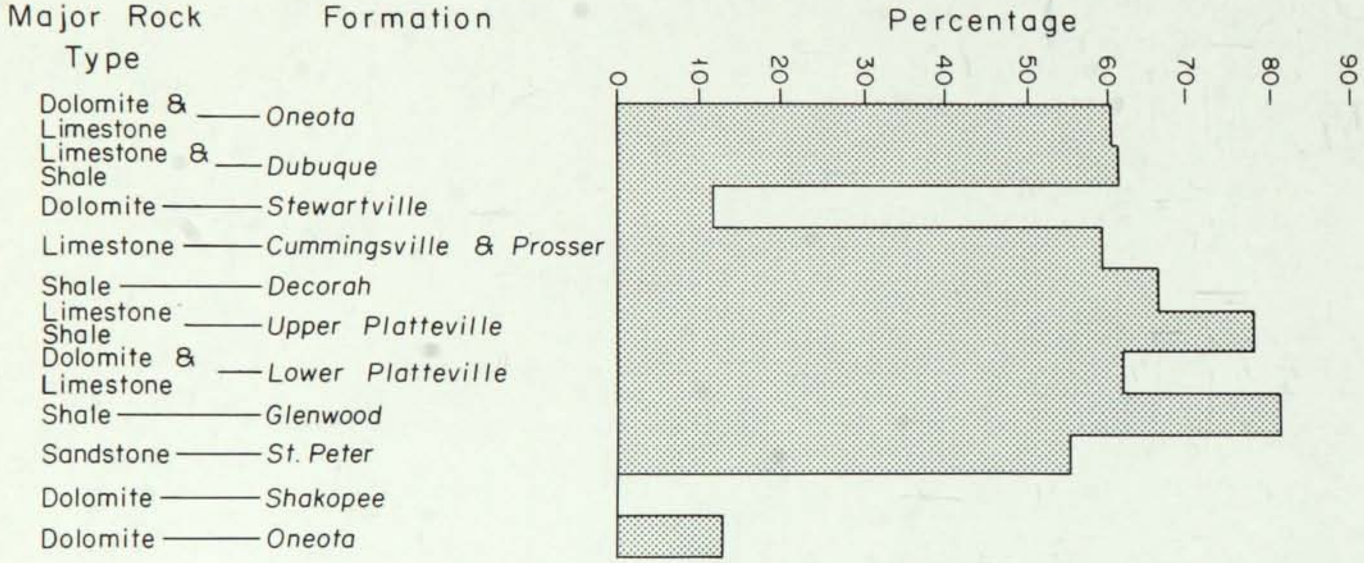


Figure VI-30. Percentage of filler-feeding macrofossil species in the Ordovician of Minnesota.

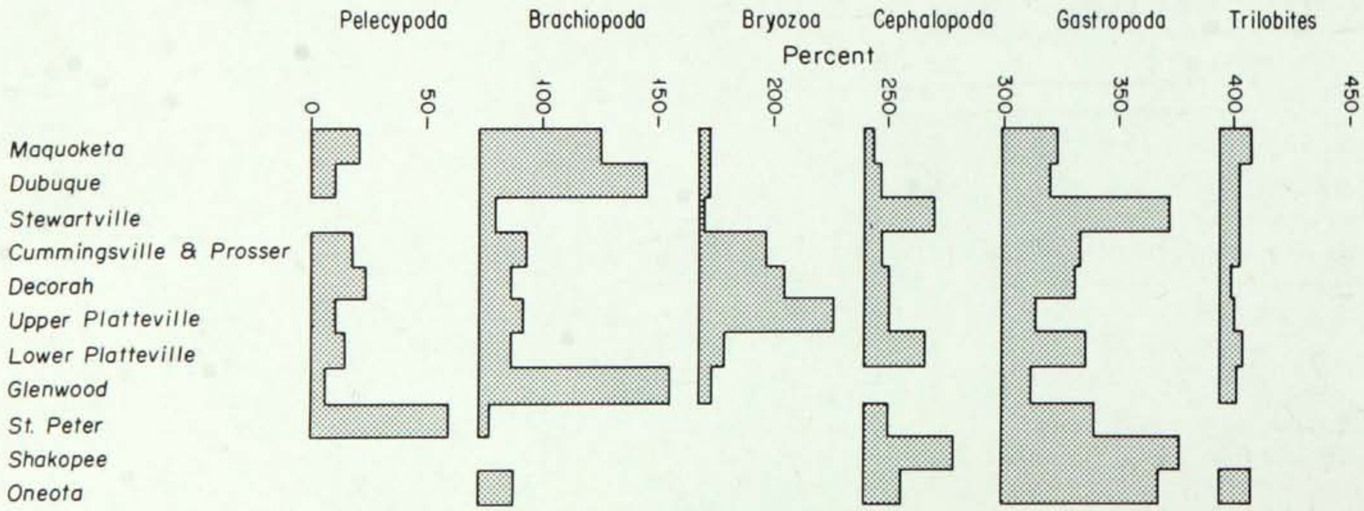


Figure VI-31. Percentage of major invertebrate phyla in the Ordovician of Minnesota.

PALEOZOIC STRUCTURE AND STRATIGRAPHY OF THE TWIN CITY REGION

John H. Mossler

The Twin City basin overlies part of a Paleozoic depositional lowland, named the Hollandale embayment (Austin, 1969), that extended from the Ancestral Forest City basin in Iowa into southeastern Minnesota and western Wisconsin (see Austin, section on Paleozoic lithostratigraphy, this volume, fig. VI-3). The Hollandale embayment is bounded by the Transcontinental Arch on the west and the Wisconsin Dome and arch on the east. It developed over part of a southwesterly extension of the Keweenawan Lake Superior syncline, a basin that contains a thick succession of basalts of Middle Keweenawan age and red beds of Late Keweenawan age. The Keweenawan syncline is cut by numerous faults which were active during the Paleozoic and gave rise to secondary structural features within the Hollandale embayment, such as the Twin City basin (Craddock and others, 1963).

Descriptions and interpretations of the structure and stratigraphy of the Twin City basin have evolved slowly because nearly all the basin is covered by a thick mantle of Pleistocene drift. The presence of a separate structural basin beneath the Twin City area north of the major portion of the Hollandale embayment was first noted by Schwartz (1936, p. 89; also in Thiel and Schwartz, 1941). Apparent structural closure on the Platteville and Jordan formations was cited in these reports as evidence for existence of this basin.

Gravimetric studies indicate southwestward extension of the Keweenawan Lake Superior syncline beneath the Paleozoic Hollandale embayment (Thiel, 1956; Craddock and others, 1963). The Twin City basin is superimposed on a gravity low at the southern end of the St. Croix horst, a positive structure which lies within the Keweenawan syncline, and it is inferred from this that the Paleozoic basin is superimposed on an earlier Keweenawan basin that lay at the southern end of the St. Croix horst (Craddock and others, 1963, p. 6029). Aeromagnetic studies confirm earlier interpretations about the configuration of basement rocks underlying the Twin City basin (Sims and Zietz, 1967).

PRECAMBRIAN STRUCTURE AND SEDIMENTOLOGY

The Twin City basin overlies the southern end of the St. Croix horst. The horst is bounded on the northwest by the Douglas, Pine, and subsidiary faults and on the southeast by the St. Croix and Cottage Grove faults. During Precambrian time, movement along some of the faults produced a deep sedimentary and structural basin at the southern end of the St. Croix horst. The basin is bounded on its northwestern and southeastern sides by small horsts and upfaulted blocks, mostly basalt and related extrusive rocks, which form the margins of the St. Croix horst (Sims and Zietz, 1967). Subsequent movement along these same fault zones during the Paleozoic produced the Twin City basin.

The Keweenawan sedimentary basin is filled with rocks that are primarily red beds of Late Keweenawan age. These rocks have not been completely penetrated by drilling in the center of the basin. Calculations indicate that the Keweenawan red beds and Paleozoic sedimentary rocks at the center of the basin have a combined thickness of nearly a mile (Sims and Zietz, 1967). Only the top 900 to 1,000 feet of the mile-thick layer is sedimentary rock of Paleozoic age. The thickest sequence of Keweenawan red beds penetrated by drilling is near Stillwater in the northern part of the basin where 2,470 feet was penetrated. The horsts and upfaulted blocks along the southeastern and northwestern margins of the Keweenawan basin are overlain by very thin sequences of Keweenawan red beds that are commonly less than 200 feet thick. Along the northern edge of the basin and along parts of its southeastern edge, Paleozoic sedimentary rocks are in direct contact with Keweenawan basalt. The red beds are dark reddish-brown, characteristically arkosic to lithic, poorly-sorted, fine- to medium-grained, very silty sandstones and sandy siltstones having similar color and composition (Kirwin, 1963, unpub. M.S. thesis, Univ. Minn., p. 22-27). Minor coarser sandstones and conglomerates also are present.

A quartzose sandstone, the Hinckley, which directly overlies the Keweenawan red clastic sediments and is interpreted to be Keweenawan in age (Austin, 1969), crops out in Pine County in east-central Minnesota, northwest of the Douglas fault and the Twin City basin. Because of close lithologic similarity between this sandstone and the overlying Upper Cambrian Mt. Simon Sandstone, scarcity of outcrops in the region and inadequate subsurface control, exact distribution of these units and their stratigraphic relationships are not known with certainty. At least one investigator has suggested that they may be the same formation (Ostrom, 1967). Early investigators correlated all quartzose sandstones in the Twin City basin between the Keweenawan red beds and fossiliferous Eau Claire shales and siltstones with the Hinckley (N. H. Winchell and Upham, 1888, p. 31-32; Stauffer, 1927a and b; Schwartz, 1936, p. 24-25).

Atwater and Clement (1935) correlated sandstones between the Eau Claire Formation and Keweenawan red beds in the upper St. Croix valley with the Mt. Simon Sandstone of Wisconsin primarily because it can be demonstrated that the Hinckley Sandstone in its type area was involved in the same tectonic movements along the Douglas fault as the underlying Keweenawan red beds and basalts and because sandstones along the St. Croix valley unconformably onlap faulted Keweenawan red beds and basalts.

Comparison of the quantity of detrital feldspar at the type sections of the Hinckley and Mt. Simon formations with feldspar contents of subsurface samples from the Twin City basin suggested to Thiel and Crowley (1940) that systematic, but minor, changes in quantity of feldspar in the basin sandstones indicated the presence of both the Hinckley and Mt Simon.

Because the Mt. Simon and other Paleozoic formations recently have been shown to be involved in considerable tectonic movement in the basin (Sloan and Danes, 1962; see also figs. VI-32, 33, and 34, this paper) and because lithologic differences between the Mt. Simon and Hinckley are slight and possibly may be due to facies changes in the same formation (Ostrom, 1967), the relationship of the Mt. Simon to the type Hinckley is still considered to be unresolved.

Most of the quartzitic sandstones between the Eau Claire Formation and Keweenawan red clastics in the Twin City basin are tentatively assigned to the Upper Cambrian Mt. Simon Sandstone for three reasons. First, correlations made from subsurface samples indicate that the Mt. Simon is a continuous blanket sand between its type section in Eau Claire, Wisconsin, and the Twin City basin. In addition, in the Twin City basin, the sandstone exhibits a gradational boundary with the overlying Eau Claire Formation and an unconformable boundary with the underlying Fond du Lac that generally is marked by a persistent basal pebble conglomerate or a very coarse sandstone. Finally, no similar persistent conglomerates or weathering horizons are present in the intervening part of the sandstone unit that would indicate an unconformity and justify separating the unit into two formations.

The Hinckley Sandstone is better preserved on the western, downthrown side of the Douglas fault (figs. VI-33 and 34), west of the Twin City basin; however, thin remnants also are present sporadically in the Twin City basin.

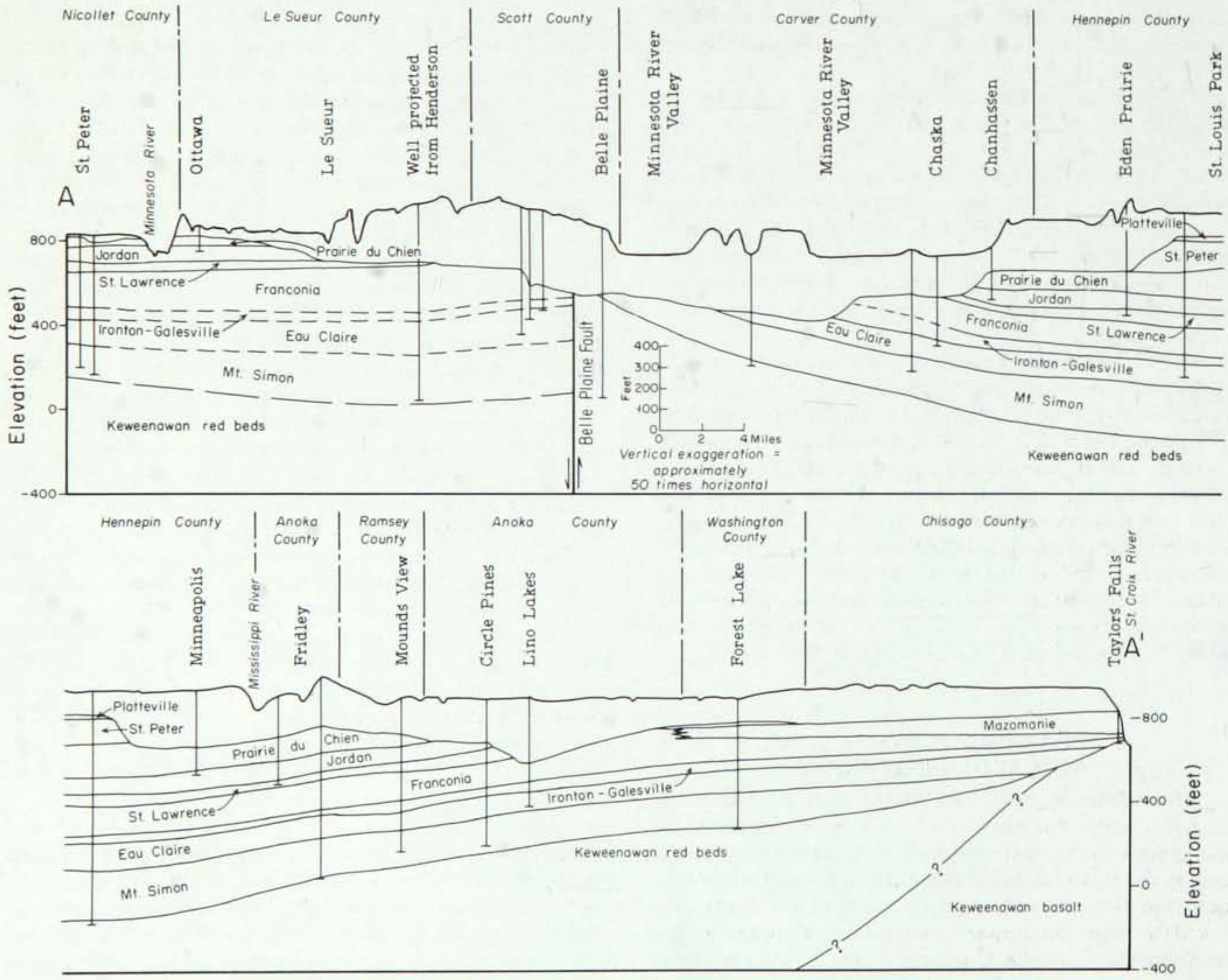


Figure VI-32. Section across the Twin City basin from Taylors Falls toward St. Peter. Line of section shown on Eau Claire structure map (fig. VI-35) as line A-A'.

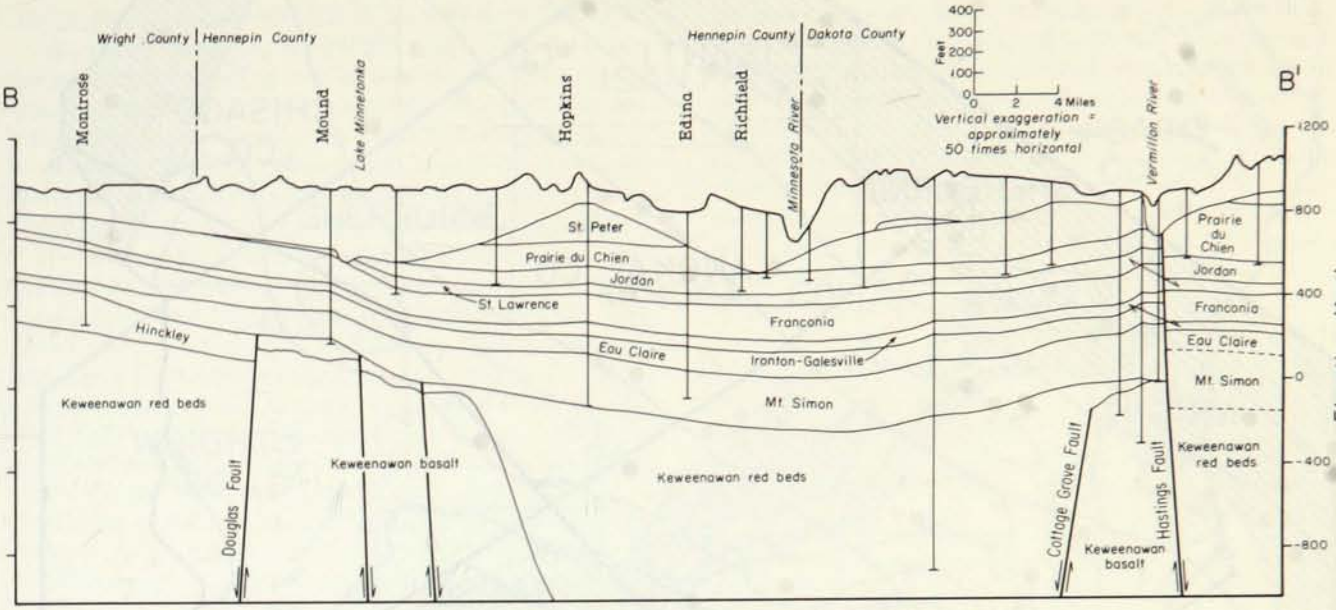


Figure VI-33. Section across the Twin City basin from region of Hastings toward Montrose. Section shown on Eau Claire structure map (fig. VI-35) as line B-B′.

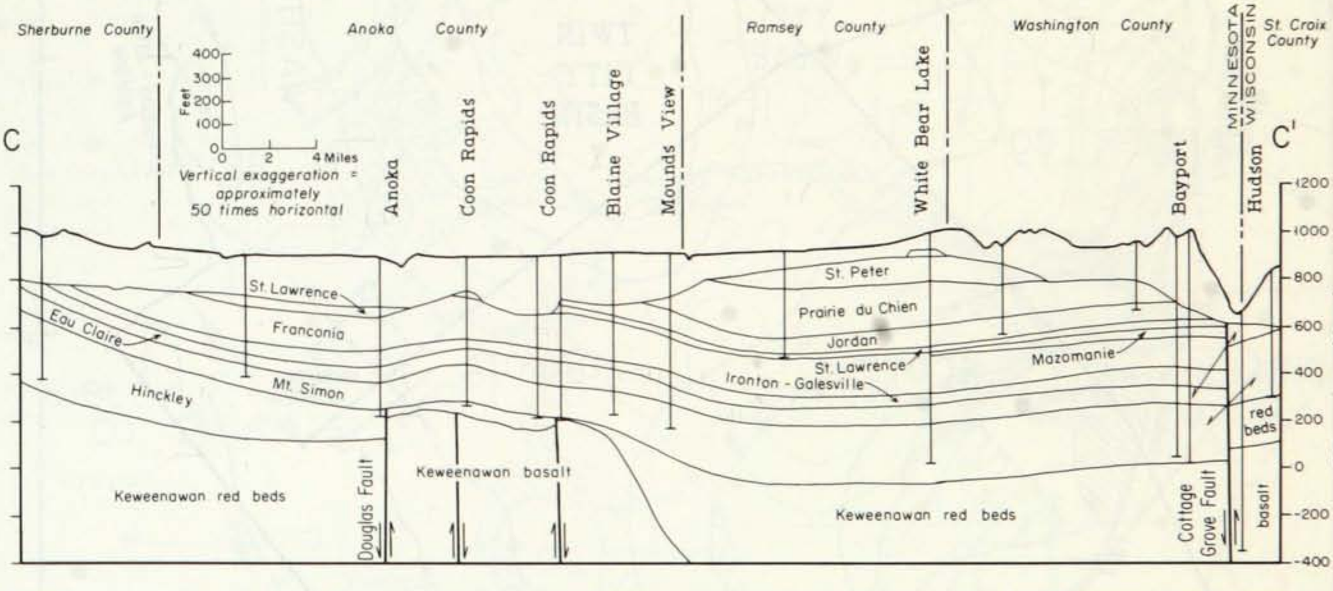


Figure VI-34. Section across the Twin City basin from region of Hudson, Wisconsin, toward northeastern Sherburne County. Section shown on Eau Claire structure map (fig. VI-35) as line C-C′.

STRUCTURAL LIMITS OF THE TWIN CITY BASIN

The eastern margin of the Twin City basin is bounded by at least three southward-plunging, en echelon anticlines (figs. VI-35 and 36). The northernmost anticline, the Hudson-Afton, crops out along the St. Croix valley and was mapped and described by Schwartz (1936, p. 94; also Thiel and Schwartz, 1941, p. 55). The anticline has a closure of more than 200 feet on the Jordan Sandstone (Schwartz, 1936, p. 94). Faults associated with this structure are exposed along the Mississippi River near Hastings (Thiel and Schwartz, 1941, p. 57), and similar faults are inferred to lie buried beneath glacial drift elsewhere on the limbs of the anticline (Thiel and Schwartz, 1941, p. 57; Tyler, 1958, unpub. M.S. thesis, Univ. Minn., p. 13).

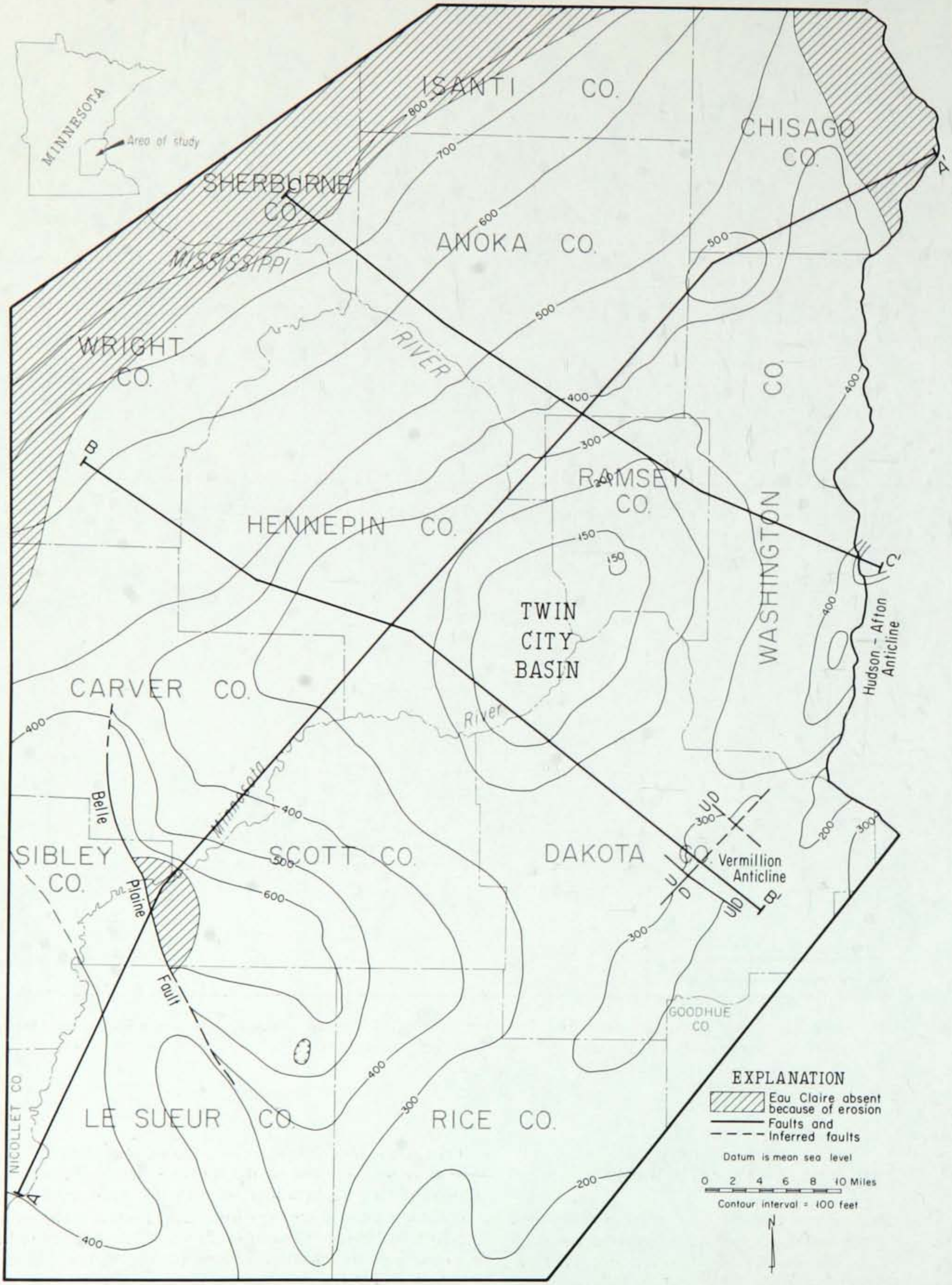


Figure VI-35. Structure map at the top of the Eau Claire Formation showing the general configuration of the Twin City basin. Lines A-A′, B-B′ and C-C′ are lines of sections across the basin.

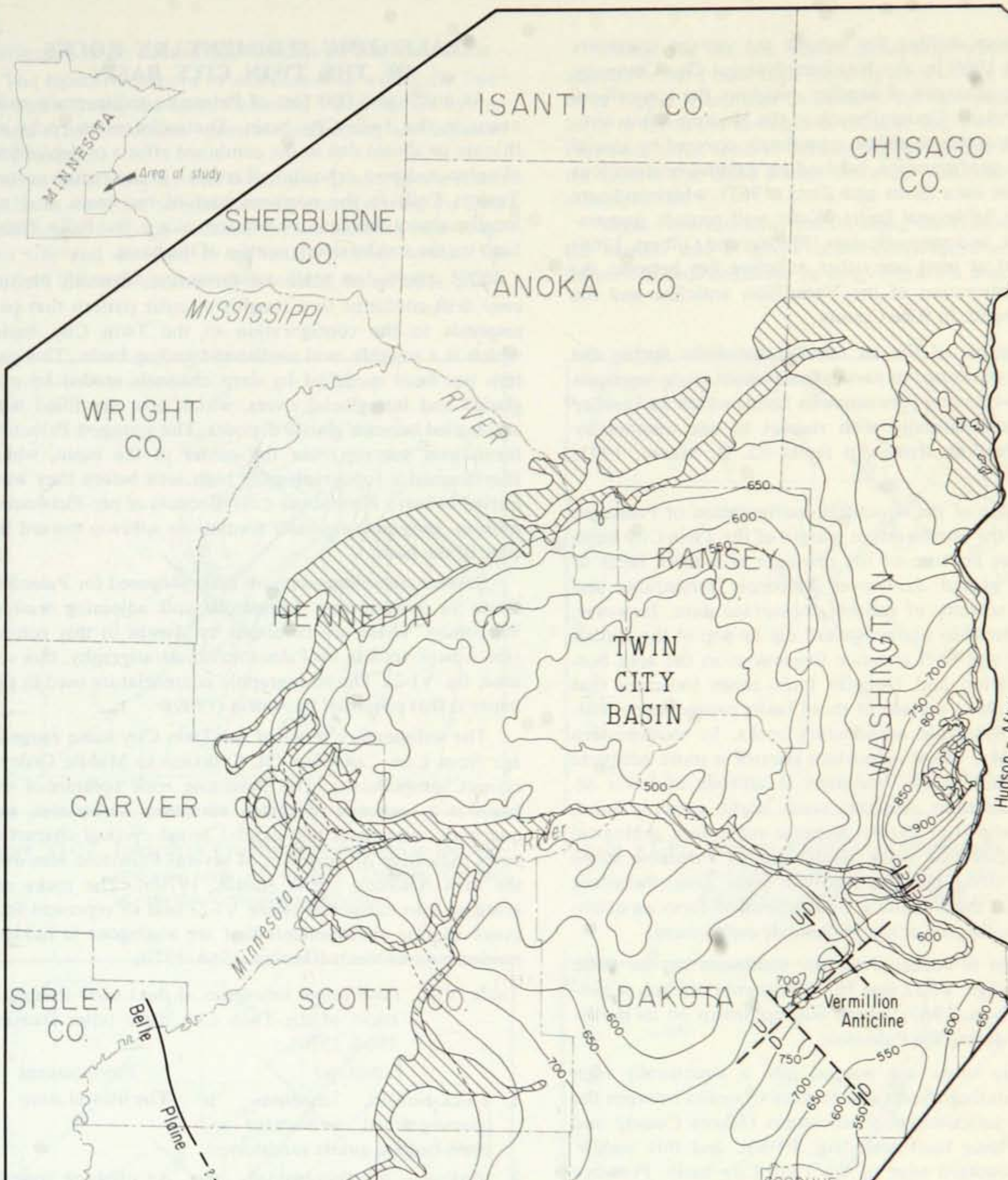


Figure VI-36. Structure map on top of the Jordan Sandstone showing the configuration of the Twin City basin.

Exploratory drilling for natural gas storage reservoirs in 1965 and 1966 by the Northern Natural Gas Company revealed the existence of another anticline, the Vermillion, in eastern Dakota County, south of the Hudson-Afton anticline. The structure is almost completely covered by glacial drift and its configuration is based on gamma-neutron logs and magnetic data (Sims and Zietz, 1967), which indicate that it is cut by several faults. Water well records, gamma-neutron logs, and magnetic data (Philbin and Gilbert, 1966) indicate that at least one other anticline lies between the southern termination of the Vermillion anticline and the city of Faribault, in Rice County.

Development of the en echelon anticlines during the Paleozoic is attributed to vertical movement along segments of northeast-trending Keweenawan horsts which had earlier been displaced laterally with respect to one another by northwest-trending strike-slip faults (G. B. Morey, 1971, oral comm.).

Knowledge of the structural configuration of Paleozoic strata along the northwestern margin of the Twin City basin is incomplete because of the presence of a thick layer of Pleistocene glacial drift over Paleozoic formations and because of scarcity of reliable subsurface data. However, marked increase in southeastward dip on top of the Jordan (fig. VI-36) and St. Lawrence formations in the area bordering the Pine and Douglas fault zones indicates that movements along certain of these faults caused some folding in the Paleozoic sedimentary rocks. In southwestern Anoka County, where subsurface control is more adequate across the fault zones, variation in altitude of lower St. Croixan formations indicates some slight, possibly local, reversal of dip (fig. VI-34). Because subsurface geological studies indicate that some disturbance of Paleozoic strata took place along the Douglas-Pine fault zone, Paleozoic strata west of these faults are considered to form an extension of the main part of the Hollandale embayment.

The basin is bordered on the southwest by the Belle Plaine fault (figs. VI-32 and 36), a Precambrian fault (Craddock and others, 1963), which was upthrown on its northeastern side during the Paleozoic.

Paleozoic strata are warped into a structurally high 'saddle' in southern Scott and Dakota Counties between the en echelon anticlines in southeastern Dakota County and the Belle Plaine fault zone (fig. VI-36), and this 'saddle' forms the southern edge of the Twin City basin. Presence of this 'saddle' was first noted by Thiel and Schwartz (1941), and has been confirmed by more recent subsurface studies.

The northern margin of the Twin City basin is partially controlled by depositional onlap of progressively younger St. Croixan sedimentary rocks onto the basalts that form the northern edge of the Precambrian sedimentary basin (fig. VI-32). However, structural contour lines on the Jordan Sandstone (fig. VI-36) and faulting in Paleozoic formations in extreme northern Washington County (Quaschnick, 1959, unpub. M.S. thesis, Univ. Minn.) indicate that structural movements also deformed the Paleozoic strata along this margin of the basin.

PALEOZOIC SEDIMENTARY ROCKS OF THE TWIN CITY BASIN

As much as 1,000 feet of Paleozoic sedimentary rocks occur in the Twin City basin. The sedimentary rocks are thinner or absent due to the combined effects of depositional onlap and post-depositional erosion in the region around Taylors Falls in the northern part of the basin, and are locally absent because of erosion along the Belle Plaine fault on the southwestern margin of the basin.

The subcrop of Paleozoic formations beneath Pleistocene drift conforms to a roughly circular pattern that corresponds to the configuration of the Twin City basin, which is a roughly oval northeast-trending basin. This pattern has been modified by deep channels eroded by preglacial and interglacial rivers, which now are filled with and buried beneath glacial deposits. The youngest Paleozoic formations subcrop near the center of the basin, where they formed a topographically high area before they were buried beneath Pleistocene drift. Because of pre-Pleistocene erosion, progressively older formations subcrop toward the edge of the basin.

Several classifications have been proposed for Paleozoic strata in southeastern Minnesota and adjoining western Wisconsin. These are tabulated by Austin in this volume (see Austin, section on Paleozoic lithostratigraphy, this volume, fig. VI-2). The stratigraphic nomenclature used in this paper is that proposed by Austin (1969).

The sedimentary rocks of the Twin City basin range in age from Late Cambrian (St. Croixan) to Middle Ordovician (Champlainian). The Paleozoic rock column of the basin is a sequence of shales, siltstones, carbonates, and quartzitic sandstones that exhibit broad cyclical characteristics indicative of incursion of several Paleozoic seas over the area (Ostrom, 1964; Austin, 1970b). The rocks are grouped into categories (table VI-2) said to represent four major marine environments that are analogous to modern marine environments (Ostrom, 1964, 1970).

Table VI-2. Four major lithotopes of the Lower Paleozoic rocks of the Twin City basin (after Ostrom, 1964, 1970).

Lithotope	Environment
1. Thick-bedded, medium- to coarse-grained, well-sorted and cross-bedded quartz sandstone	The littoral zone
2. Medium- to thin-bedded, reworked quartz sandstone characterized by alternating poorly-sorted sandstone which is commonly burrowed, calcareous, slightly glauconitic and shaly and well-sorted, medium- to coarse-grained sandstone	An offshore zone of little or no deposition and reworked littoral sediments
3. Shale or argillaceous, thin-bedded sandstone that is fine-grained, glauconitic, and shaly with minor carbonate	Shelf depositional zone of fine-grained clastics
4. Carbonate or sandy or silty carbonate or calcareous siltstone	Biogenic zone of calcareous reefs

Influence of Structural Movements on Sedimentation

The repetitive pattern of sedimentation within the Twin City basin during the Early Paleozoic was modified by major and minor structural elements underlying and adjacent to the basin.

Close proximity of the Wisconsin Dome to the Twin City basin modified sedimentation patterns during the Late Cambrian, and this is reflected by an increase in average grain size and decrease in carbonate content toward the dome on the north and northeast (Berg and others, 1956).

Except for the Mt. Simon Sandstone, whose thickness appears to have been influenced by pre-existing topography, most Upper Cambrian formations are comparatively uniform in thickness throughout most of the Twin City basin. However, some seem to thicken slightly toward the Wisconsin Dome, the chief source for their sediment (Berg and others, 1956).

Slight recurrent movements along the Hastings fault on the eastern side of the Precambrian Hudson-Afton horst influenced sedimentation in the Vermillion area of Dakota

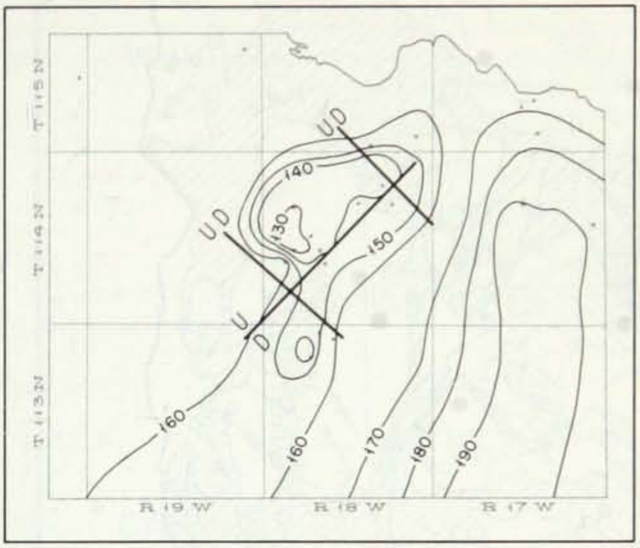


Figure VI-37. Isopach of Franconia Formation in the Vermillion area of eastern Dakota County across the Vermillion anticline.

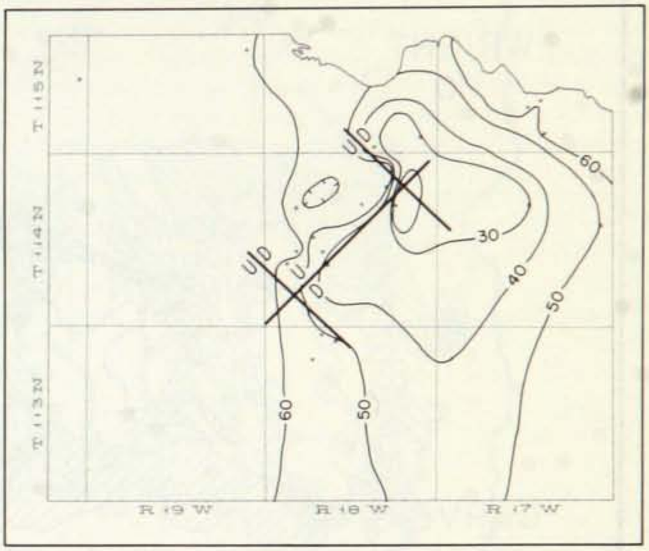


Figure VI-39. Isopach of combined thicknesses of Ironton and Galesville formations in eastern Dakota County across the Vermillion anticline.

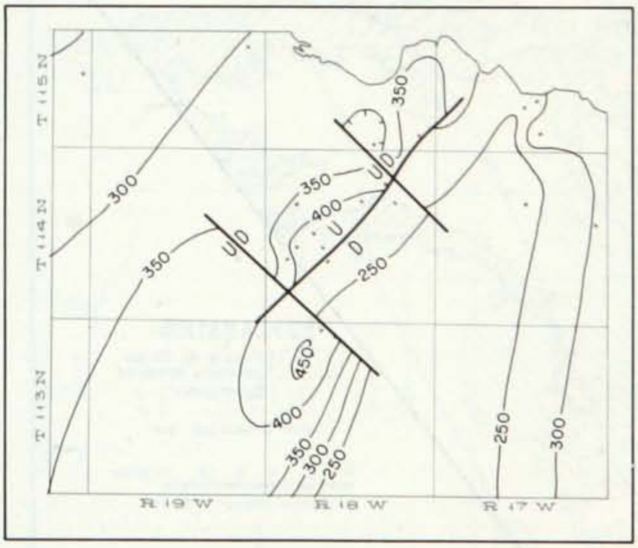


Figure VI-38. Structure map on top of the Ironton Sandstone in eastern Dakota County across the Vermillion anticline.

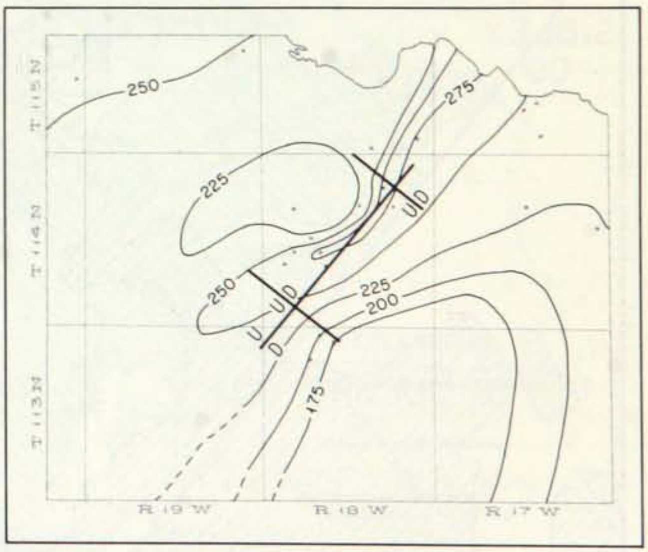


Figure VI-40. Isopach of the Mt. Simon Sandstone in eastern Dakota County across the Vermillion anticline.

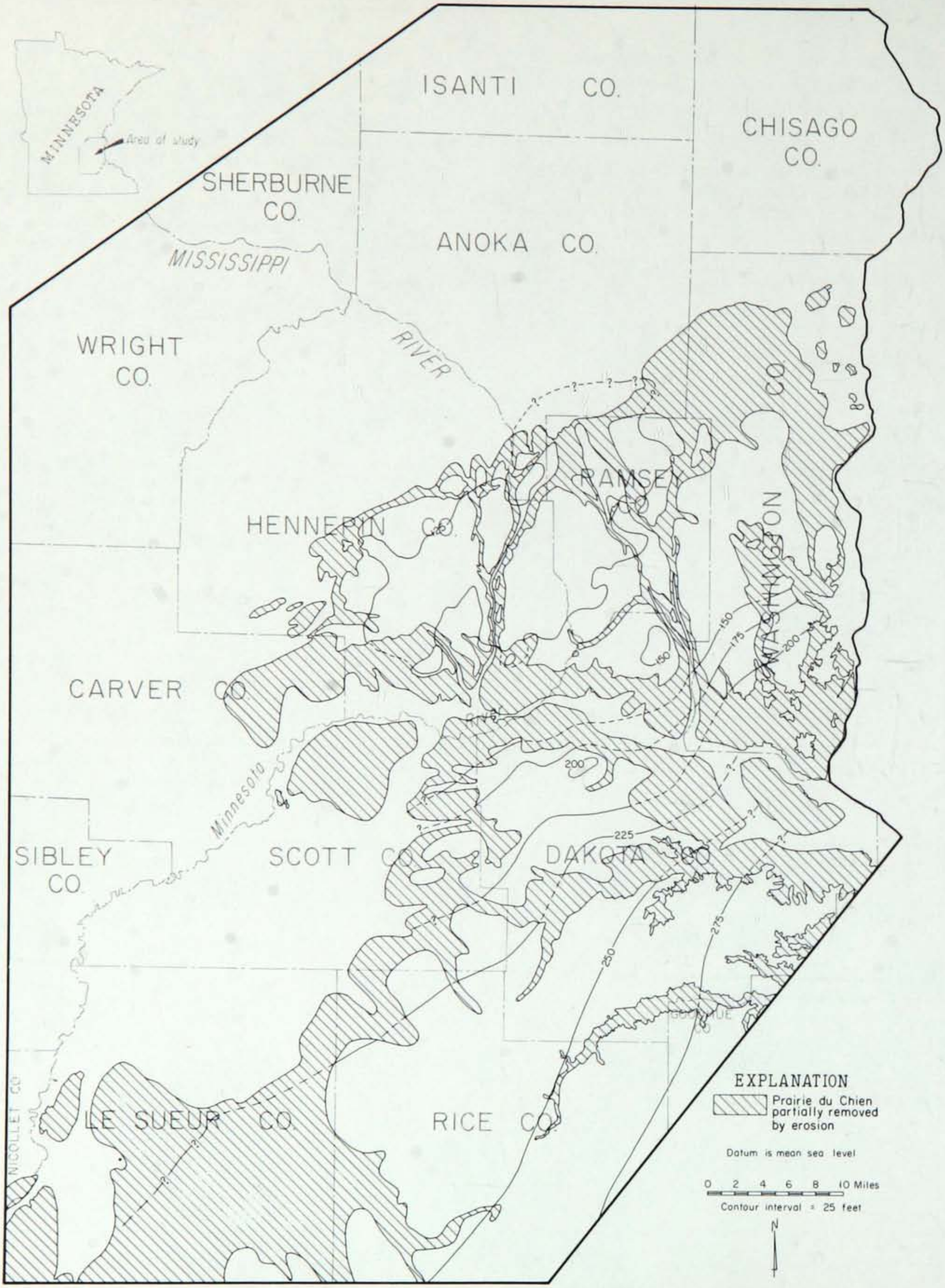


Figure VI-41. Isopach of the Prairie du Chien Group in the Twin City region. Based on wells in which the group has not been subjected to post-St. Peter erosion.

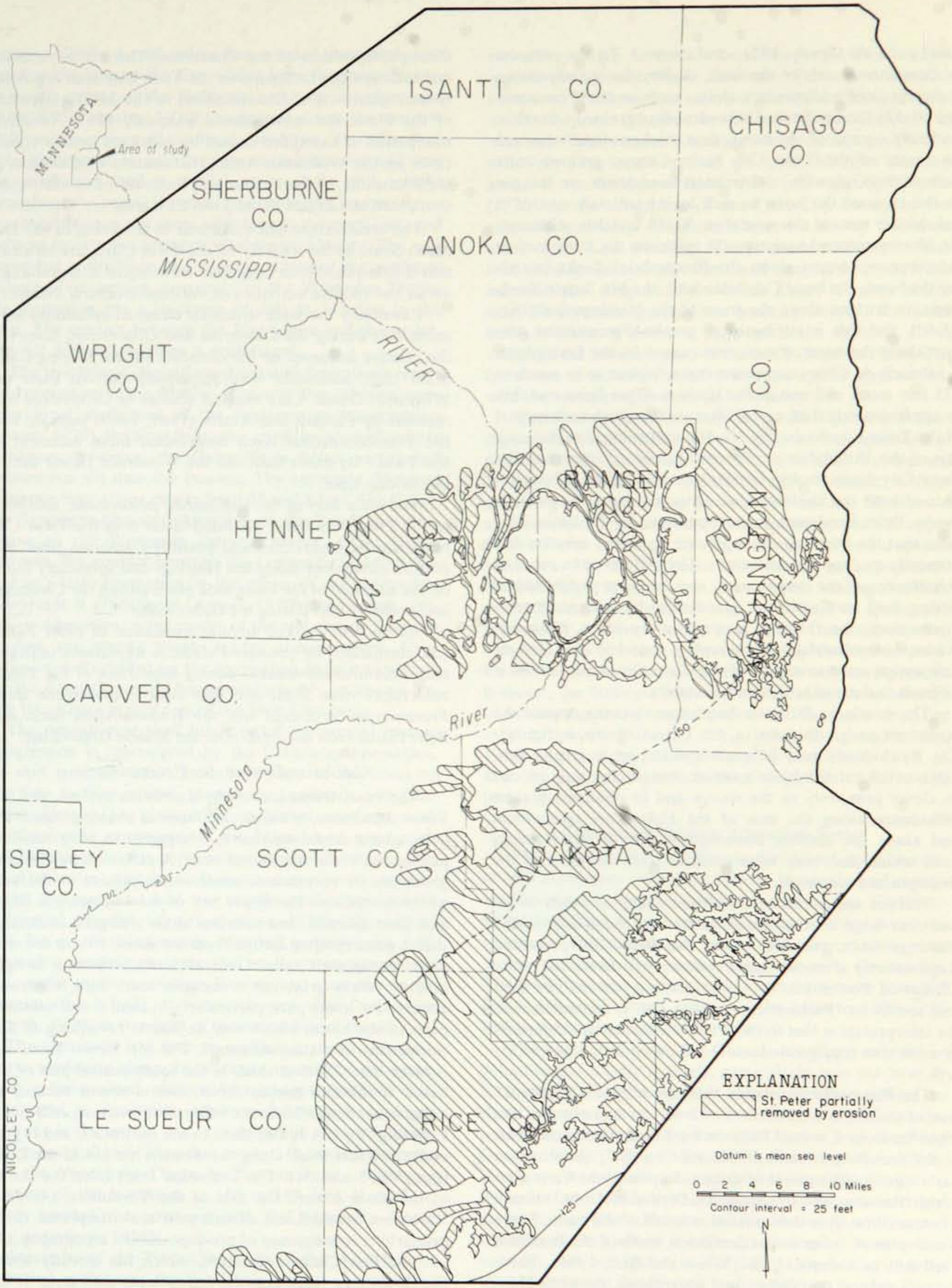


Figure VI-42. Isopach of the St. Peter Sandstone in the Twin City region. Based on wells in which the formation has not been subjected to erosion.

County (G. B. Morey, 1971, oral comm.). To the east, on the downthrown side of the fault, slightly thicker sequences of fine-grained sedimentary units, such as the Franconia (fig. VI-37) accumulated. These same fine-grained units thin markedly on top of the horst, then thicken slightly toward the center of the Twin City basin. Coarser-grained units such as the Galesville and Ironton Sandstones are thinner on the crest of the horst as well as immediately east of it and thicker toward the west (figs. VI-38 and 39). Although the aforementioned variations in thickness are attributed to recurrent movement along the Precambrian faults bounding the horst, the basal Cambrian unit, the Mt. Simon Sandstone, is thickest along the trace of the Hastings fault (fig. VI-40), and this is attributed to greater Precambrian erosion along the zone of weakness caused by the fault plane.

Prairie du Chien carbonate rocks thicken to as much as 275 feet south and east of the Hudson-Afton horst and thin to approximately 125 to 140 feet in the northwestern part of the Twin City basin (fig. VI-41), indicating that the main axis of the Hollandale embayment subsided faster than the Twin City basin during Early Ordovician time. The presence of both the Shakopee and Oneota formations beneath the St. Peter Sandstone throughout most of the basin indicates that the thinning of the Prairie du Chien must be due primarily to depositional causes. The difference in rates of subsidence of the embayment and basin probably began during the Late Cambrian, as indicated by recurrent movements along the Hastings and subsidiary faults. However, inadequate subsurface stratigraphic contol in the Hollandale embayment southeast of the Twin City basin makes it difficult to determine this with certainty.

The overlying St. Peter Sandstone thickens toward the northwest and is thickest in the Twin City basin (fig. VI-42). Its thickness may be attributed in part to greater subsidence toward the basin's center, but it also may be due to closer proximity to the source and to relatively greater subsidence along the axis of the Hollandale embayment and along the eastern, downthrown side of the Hastings fault which then may have acted as traps for most of the finer-grained sediments.

Thicker sequences of Glenwood shale, as much as 16 feet near Sogn in Goodhue County, are found east of the Hastings fault, consistent with the latter interpretation. Regional clay mineral studies indicate that the source of the Glenwood Formation in Minnesota was toward the west and southwest (Parham and Austin, 1967), consistent with the interpretation that the St. Peter Sandstone was deposited in a sea that transgressed toward the west in the Twin City area.

The Platteville Formation thins from approximately 35 feet in the Twin City basin to 15 feet or less along a line trending from Cannon Falls to Faribault, Minnesota, east of the anticlines in eastern Dakota County. The formation has a significant amount of dolomitization in the Twin City basin (Rassam, 1967, unpub. Ph.D. thesis, Univ. Minn.), whereas it has little dolomitization south of the basin. There is a change of facies in the formation south of the line from Faribault to Cannon Falls (Weiss and Bell, 1956). These features suggest that differential movements along the Hudson-Afton horst during Late Cambrian time continued through deposition of the Platteville. The extensive dolomitization of the formation in the Twin City area is possibly analogous to the dolomitization of the Mifflin Member of the Platteville in Wisconsin (Asquith, 1967), which is interpreted to have been caused by restricted marine conditions on the Wisconsin Arch. Distribution of dolomite in the Platteville of Minnesota may indicate more restricted, shoreward conditions in the Twin City area.

The areal extent and variations in thickness of the Decorah Shale in the vicinity of the Twin Cities are so small that it is impossible to make any meaningful interpretations about the relative activities of various structural elements.

Essentially the same structural elements influenced sedimentation during the Cambrian and Ordovician; however, the relative influence of the elements varied. During Cambrian time, sediments were principally derived from the Wisconsin Dome. Clay mineral studies of Ordovician formations by Parham and Austin (1967, 1969) indicate that the Transcontinental Arch contributed more sediment to the Twin City basin than did the Wisconsin Dome during Ordovician time.

The main axis of the Hollandale embayment, southeast of the Twin City basin, subsided faster than the Twin City basin during Ordovician and possibly Cambrian time. Recurrent movement along the Hastings and subsidiary faults on the east side of the basin took place during the Cambrian and probably also during the Ordovician.

Absence of marked angular truncation of older Paleozoic formations by younger Paleozoic formations indicates that structural movements during deposition of the Paleozoic rocks were slight and that most displacement along Precambrian structures like the Hudson-Afton horst and Belle Plaine fault occurred after the Middle Ordovician.

Cambrian System, St. Croixan Series

The basal formation of the Cambrian System, the Mt. Simon Sandstone, is mostly medium- to coarse-grained, friable, quartz sandstone that is interpreted to have been deposited mainly in the littoral zone. A persistent pebble conglomerate or very coarse sandstone is present at the base of the formation. The upper half of the formation is siltier and finer grained, and contains shale stringers, indicating that it was deposited farther from the strand line in the non-depositional shelf zone. Generally, the formation is light gray or yellow gray, but it contains some dark red-brown zones in its lower part, particularly in central and southern parts of the basin, which may be due to reworking of Keweenawan red clastic sediments. The Mt. Simon Sandstone is more than 250 feet thick in the south-central part of the basin. It thins to the northeast, and is absent because of onlap in the upper St. Croix valley in the vicinity of Taylors Falls (fig. VI-32). It also thins to the northwest, and is generally 150 feet thick along the western margin of the basin (figs. VI-33 and 34). The formation thins from the center of the basin toward the axis of the Vermillion anticline. Variations in thickness of this unit are interpreted to be primarily a consequence of pre-depositional topography.

The Eau Claire Formation, which lies directly above the Mt. Simon, represents the depositional shelf zone. It has been divided into five informal rock units in Minnesota

(Austin, 1969). A red, silty, fine-grained sandstone unit found at the base of the Eau Claire from the western margin to the center of the Hollandale embayment (Berg and others, 1956; Austin, 1970b) is present along the extreme southwestern margin of the Twin City basin. The other four units, glauconitic to slightly glauconitic siltstones and very fine sandstones and a medial green shale unit, persist throughout the basin. However, the shale unit becomes progressively thinner and more silty toward the northeast, and is absent in extreme northern Washington County. The unit diminishes in thickness and disappears toward the northeast because of increasing proximity to the Wisconsin Dome. The formation has a thickness of 100 to 125 feet in the basin. The contact between the Eau Claire and the underlying Mt. Simon Sandstone is gradational.

The overlying Galesville and Ironton formations represent, respectively, the littoral zone and the offshore nondepositional shelf zone of the next marine transgression (Ostrom, 1970). Both are light-gray, medium-grained, quartz sandstones; however, the Galesville is better sorted and contains less silt than the Ironton. The aggregate thickness of the two formations ranges from 35 to 65 feet. The Galesville Sandstone lies unconformably on the Eau Claire Formation on the Wisconsin Dome (Ostrom, 1964, 1970). However, the contact between the Galesville and underlying Eau Claire Formation in the center of the Hollandale embayment is gradational (Austin, 1970b), indicating continuous deposition in the center of the embayment between the first and second major marine transgressions. Core studies of wells drilled on the Vermillion anticline indicate that the basal Galesville lies unconformably on the underlying Eau Claire in that part of the Twin City basin.

The depositional shelf lithotope of the second marine transgression is represented by the Franconia Formation. The unit is represented throughout most of the basin by very fine- to fine-grained, glauconitic sandstone and glauconitic siltstone. The glauconitic sandstones interfinger with non-glauconitic sandstones of the littoral and nondepositinal lithotopes in the northeastern part of the basin. The non-glauconitic sandstones become progressively coarser toward the northeast and are mostly medium-grained in east-central Chisago County (Nelson, 1956). This shoreward facies, the Mazomanie Member, intertongues with glauconitic sandstones along the northern margin of the basin, as far south as northern Ramsey and southern Anoka Counties (fig. VI-32). It may represent slight regression (Ostrom, 1970) or standstill of the strand line of the sea.

The formation is a fine- to medium-grained, non-glauconitic sandstone along the northern margin of the basin, mostly a fine-grained, glauconitic sandstone along the eastern margin, a glauconitic siltstone and very fine-grained sandstone in the center, and very dolomitic along the southwestern margin, indicating that most of the sediments came from the Wisconsin Dome to the northeast. Aggregate thickness of the formation is 150 to 165 feet. It thickens slightly toward the northeast and the Wisconsin Dome (Berg and others, 1956).

Proximity of the Twin City basin to the major source of detrital sediments, the Wisconsin Dome, continued to influence sedimentation during deposition of the St. Lawrence Formation. In the southern, deeper parts of the Hollandale embayment, the St. Lawrence Formation is mainly composed of a dolomitic member, the Black Earth, assigned to the biogenic lithotope by Ostrom (1970). Toward the northeast, this member interfingers with dolomitic siltstones and very fine sandstones of the Lodi Member that are assigned to the depositional shelf lithotope by Ostrom (1970). In the subsurface of the Twin City basin, the formation is primarily dolomitic siltstone and very fine sandstone of the Lodi Member. The presence of finely crystalline, resistant dolomite in some well samples from the basin indicates that tongues of the Black Earth Member extend into the southern part of the basin. However, the Black Earth Member is absent in outcrops in the upper St. Croix valley, in the northern part of the basin (Nelson, 1956). The formation is as much as 65 feet thick and is conformable with the underlying Franconia Formation.

The Jordan Sandstone represents the littoral and nondepositional shelf lithotopes of the next transgression of marine waters (Ostrom, 1970). The formation is composed of three members: the basal Norwalk Member, a fine-grained quartz sandstone; the middle Van Oser Member, a medium- to coarse-grained quartz sandstone; and the upper Sunset Point Member, an argillaceous, dolomitic quartz sandstone. The Sunset Point Member, the nondepositional shelf lithotope, is not known to be present in the subsurface of the Twin City basin, but may be difficult to distinguish from sandy dolomites at the base of the Oneota Dolomite. The Van Oser Member is the most widely-distributed member in the basin, and commonly is the only one present; however, the Norwalk Member is present in the subsurface of extreme southern Anoka County and has been identified in outcrops along the St. Croix valley (Berg and others, 1956) and in Scott County (Stauffer and Thiel, 1941, p. 49). The formation has a uniform thickness of 85 to 100 feet in the basin.

Ordovician System, Canadian Series

The Lower Ordovician Prairie du Chien Group consists of two formations: the basal Oneota Dolomite and overlying Shakopee Formation. Isopach contours (fig. VI-41) indicate that these formations thicken appreciably toward the southeast, indicating that greater subsidence took place within the Twin City basin during the early Ordovician. The total thickness of the group ranges from more than 200 feet at the southeastern margin of the basin to 125 to 140 feet in the northern part of the basin.

The basal formation, the Oneota, is a finely-crystalline dolomite that locally is sandy, particularly near the base. It is sparingly fossiliferous and rarely contains oolites or chert. Most recent workers believe that no significant depositional break exists between the Jordan and overlying Oneota (Kraft, 1956; Heller, 1956; Ostrom, 1964, 1965). However, well-developed transitional deposits similar to those further south in the Hollandale embayment, described as the Sunset Point Member of the Jordan Sandstone and the Blue Earth Beds of the Oneota Dolomite (Austin, this volume), are not found in the Twin City basin; and the contact between the Oneota and underlying Jordan Sandstone is diastemic at Stillwater (Kraft, 1956). The absence of well-developed transitional deposits between the Oneota and Jordan in the Twin City area is attributed to proximity of the area to the edge of the Hollandale embayment. The Oneota is interpreted to be the biogenic lithotope of the marine transgression that began with deposition of the Jordan Sandstone (Ostrom, 1970).

The overlying Shakopee Formation is composed of two members, a basal quartz sandstone, the New Richmond, and a dolomite, the Willow River. They are separated from the Oneota by a disconformity and are interpreted to be representative of another, more minor, transgression of marine waters (Ostrom, 1970).

The New Richmond is a discontinuous sandstone unit in the Twin City basin, as it is elsewhere in the Hollandale embayment (Austin, section on Paleozoic lithostratigraphy, this volume). Lithologically, its sandstones are well-sorted, fine- to medium-grained quartz sandstones that have abundant carbonate cement. Dolomitic siltstone at some places occupies the same interval as the sandstone, and commonly the member is not present and dolomites of the Willow River lie directly on the Oneota. The New Richmond is commonly from 0 to 25 feet thick; however, in a well at Minnetonka Mills on the western edge of the basin it is as much as 60 feet thick.

The overlying Willow River Member is finely crystalline dolomite which is commonly sandy or oolitic and in most places contains abundant chert. The unit generally contains discontinuous stringers of sandstone that are lithologically similar to the underlying New Richmond. Algal stromatolites are commonly present in outcrops. Davis (1966b) interpreted the environment of deposition of the Willow River to be similar to modern environments of carbonate deposition in Florida and Australia. The unit is very sandy in the Twin City basin, especially near the top, which probably is a reflection of the proximity of the area to a northern or northwestern strand line. Davis (1966b) depicted the Twin City region as an area of shoaling and very shallow water on his map of the paleogeography during Willow River time.

Ordovician System, Champlainian Series

The St. Peter Sandstone represents the littoral zone of the next marine transgression. It is a light yellow-gray to light-gray, fine- to medium-grained, generally well-sorted quartz sandstone. A very thin, basal gray-green shale is rarely present in well samples. There is a thin persistent siltstone unit about 50 feet above the base of the formation that is commonly 6 feet thick. The top 8 to 10 feet of sandstone is coarser than the underlying sandstone and has a higher clay content, and is considered to be a lower member of the Glenwood Formation by Templeton and Willman (1963, p. 51-52). This sandstone unit is transitional to the overlying Glenwood shales. The basal contact of the St. Peter on the Prairie du Chien is unconformable; however, it does not have large-scale relief similar to that described by Ostrom (1964, 1967) in Wisconsin. In the Twin City basin, well sample analyses indicate that the St. Peter Sandstone unconformably overlies the Shakopee Formation and the Oneota Dolomite. It is not known to directly overlie any St. Croixan formations. The St. Peter is as much as 155 feet thick in the center of the basin and thins toward the south (fig. VI-42). In southern Goodhue County it is less than 100 feet thick in some wells. The thickening in the Twin City area is interpreted as primarily due to proximity to a western or northern strand line and the thinning to the southeast as due to relatively greater subsidence and deeper marine waters along the axis of the Hollandale embayment.

The Glenwood Formation consists of nondepositional and depositional shelf sediments (Ostrom, 1964, 1970). The lower 8 to 10 feet of the unit, as defined by Templeton and Willman (1963), consists of the coarser, clayey sandstone at the top of the St. Peter Formation. The upper two and one-half to three feet is chiefly phosphatic, arenaceous green shale with minor sandstone layers. Southeast of the basin and the Hastings fault, in the main part of the Hollandale embayment, the upper shale unit thickens to as much as 16 feet and overlying Platteville carbonates thin and become extremely shaly near their base.

The Platteville Formation consists of carbonates representative of the biogenic lithotope. It is composed of three members, of which the basal unit, the Pecatonica, is arenaceous, phosphatic, finely-crystalline dolomite that is transitional between the underlying detrital sediments and overlying less arenaceous carbonates. The medial McGregor Member is subdivided into three submembers in the Twin City basin. The basal Mifflin submember is a thin, crinkly-bedded dolomitic limestone. The medial Hidden Falls submember is massive, very argillaceous, finely-crystalline dolomite. The upper Magnolia submember is medium-bedded, fossiliferous, finely-crystalline dolomite. These units are overlain by the Carimona Member, medium-bedded fossiliferous limestone that is interbedded with shale beds as much as 9 inches thick that are transitional to the overlying Decorah shales. The formation ranges in thickness from 35 feet in the center of the Twin City basin to approximately 15 feet at its southeastern margin.

The medial McGregor Member is composed of thin, crinkly-bedded, finely-crystalline, dolomitic limestone in the portion of the Hollandale embayment south of the Twin City basin. Lithologic changes in the member, thinning of the entire formation toward the southern margin of the Twin City basin, and the more intense dolomitization of the Platteville in the basin (Rassam, 1967, *op. cit.*), suggest that recurrent movements along the Hudson-Afton anticline possibly influenced depositional patterns in the basin. Intense dolomitization of the McGregor Member and overlying Carimona Member in the Twin City basin resembles dolomitization in shoreward facies of other ancient carbonates and may indicate that the Platteville of the Twin City basin formed closer to the ancient shoreline than the Platteville elsewhere in the Hollandale embayment.

The Decorah Shale is a greenish-gray to olive-gray, fossiliferous shale with minor coquinoidal limestone lenses. It attains a maximum thickness of 80 feet in St. Paul near the center of the basin and thins progressively to the south, in the main part of the Hollandale embayment, to 23 feet in southeastern Fillmore County (Weiss and Bell, 1956). The shale is considered part of the biogenic lithotope of the Champlainian transgression, a marine transgression that is considered to have continued into the Cincinnatian (Ostrom, 1964, 1970). The Decorah interfingers with and is replaced by carbonate units in extreme southern Minnesota and Iowa which reflect the prominence of biogenic carbonate-producing activities there. Greater thicknesses of shale in the Twin City area may indicate greater influx of detrital clay sediment into that area, which would have suppressed normal biogenic activities producing carbonate. The clay mineralogy of the formation (Parham and Austin, 1969), like that in the underlying Glenwood Formation, indicates that the source for the clays lay toward the southwest and west, indicating that the paleogeography was the same during the Decorah interval as during preceding Ordovician intervals.

A small outlier, as much as 30 feet thick, of the limestone of the lower part of the Galena Formation overlies the Decorah Shale along the Mississippi River in St. Paul (Stauffer and Thiel, 1941, p. 187-189; Schwartz, 1936, p. 54). Post-depositional erosion has removed all Cincinnatian and any other succeeding rock units from the Twin City area.

SUMMARY

The Twin City basin is part of a Paleozoic depositional lowland, the Hollandale embayment, that extended northeastward from the Ancestral Forest City basin between the Transcontinental Arch on the west and the Wisconsin Dome and arch on the east. The Hollandale embayment developed over part of a pre-existing Precambrian basin that was a site for deposition of thick red bed sequences during the Late Keweenawan and mafic extrusives during the Middle Keweenawan. During the Early Paleozoic it continued to be a locus for deposition of sediments. The Precambrian basin is cut by numerous faults which were active during Paleozoic time and produced smaller low and high areas within the Hollandale embayment. One of the major depressions that was formed by movement along these faults is the Twin City basin.

As much as 1,000 feet of Paleozoic sedimentary rocks are present in the Twin City basin. These rocks are composed of four recurrent lithologies that record at least five transgressions of marine waters across the area during the Late Cambrian (St. Croixan) and Early and Middle Ordovician (Canadian and Champlainian). During the Cambrian, the sediments were primarily silts and quartzitic sands derived from the Wisconsin Dome northeast and north of the basin. During the Early Ordovician, the Transcontinental Arch began to influence sedimentation patterns more strongly and contributed more sediments to the basin. Lower and Middle Ordovician rocks of the basin are primarily quartz sandstones, carbonates, and shales. Slight recurrent movements along the Hastings fault on the east side of the basin influenced depositional patterns during both the Cambrian and Ordovician. Structural contour maps indicate that much of the structural movement that gave rise to the Twin City basin came after the deposition of the Middle Ordovician (Champlainian) formations. The geologic history of the area after the Middle Ordovician and prior to Pleistocene glaciation can only be inferred from surrounding areas because erosion has removed any sedimentary rocks that once may have overlain the Middle Ordovician shales and carbonates.

THE IRON ORES OF SOUTHEASTERN MINNESOTA

Rodney L. Bleifuss

Concentrations of iron oxides are common in the pre-Pleistocene weathering mantle of southeastern Minnesota, and commercial iron ore mines have been developed from some of these deposits in western Fillmore, southern Olmsted, and eastern Mower Counties. Following extensive exploration work that was conducted in the 1930's, two companies carried out mining operations in the Fillmore County district from 1942 to 1968. Cumulative production is 8.1 million tons of iron ore.

The iron ore deposits were first mentioned in the Minnesota Geological and Natural History Survey Final Report for the years 1872-1882 (N. H. Winchell and Upham, 1884). Since then they have been studied by several geologists (Stauffer and Thiel, 1941, 1944, 1949; Sloan, 1964; Kohls, 1961, unpub. Ph.D. thesis, Univ. Minn.; Austin, 1963), who generally agreed on the following points relative to the origin of the ores and their relationship to the Cretaceous Windrow Formation:

(1) The ores were formed by weathering of the underlying limestone units;
(2) The development of the ore bodies required some supplementary process of concentration, involving migration and local concentration of iron during the weathering cycle;
(3) The age of the Windrow Formation is Cretaceous, and the deposits in the Fillmore County district are correlative with similar lithologic units of known Cretaceous age in other parts of the region;
(4) Fossil evidence that would positively date the Windrow Formation is absent in the district; and
(5) The most likely age of the iron-rich residuum and associated iron ores is Cretaceous.

Contrary to the conclusions above, I propose that the ores are Tertiary in age, and that they were developed from the oxidation of a primary marine siderite facies of the Cedar Valley Formation.

GENERAL GEOLOGY

Bedrock Formations

The bedrock exposed in the Fillmore County district ranges in age from Middle Ordovician (Platteville Formation) to Middle Devonian (Cedar Valley Formation).

Commercial iron ore bodies are nearly restricted to the outcrop areas of the Stewartville Member of the Galena Formation and to the Solon Member of the Cedar Valley Formation; there are no known ore deposits on either the intervening Maquoketa or Dubuque Formations. This distribution pattern of the ore leases is shown on the bedrock geologic map of the district (fig. VI-43). Although some of the leases fall within the cartographic limits of the Dubuque and Maquoketa Formations on Figure VI-44, neither formation has been identified definitely within any of the mines; the overlap reflects the uncertainties inherent in delineating the geologic units in detail, and the contacts shown on the original maps have been adjusted only where positive evidence was available. The bedrock dips gently to the southwest, and the ore is localized near the surface along the featheredges of the Cedar Valley and Galena Formations, as shown on the generalized cross-section (fig. VI-44).

Surficial Materials

Discontinuous ferruginous sands and gravels associated with the iron ores—considered part of the Windrow Formation—disconformably overlie the bedrock. The Windrow Formation consists of two members, a lower iron-rich regolith (Iron Hill Member), and an upper unconsolidated clastic unit (Ostrander Member).

In the past, the iron ores of the Fillmore County district have been considered part of the Iron Hill Member, which was defined by Andrews (1958) as the iron-rich regolith on the pre-Pleistocene erosional surface. My study shows, however, that the iron ores have a distinctly unique origin and are not part of the normal bedrock weathering residuum.

The Ostrander Member, a widely distributed unit in southeastern Minnesota, is exposed in all the mines and in several gravel pits in the area. It is composed of unconsolidated gravels, sands, silt, and clay, and varies in lithology widely over short distances. Many deposits show a distinctive yellow or orange color because of contamination by finely divided goethite; others are composed of clean sand and lack iron contamination. Silicified fossils of both Devonian and Ordovician species, such as corals, brachiopods, pelecypods and, commonly, stromatoporoid fragments are found in the deposits. The member is generally considered to be of fluvial origin (Andrews, 1958), and to represent terrestrial deposits equivalent to the upper Carlile and the basal Niobrara formations further west. The deposits are considered to have formed on a coastal plain during eastward transgression of the Cretaceous epicontinental sea (Sloan, 1964).

DESCRIPTION OF THE IRON ORES

Mineralogy

The ore is composed predominantly of the mineral goethite and has minor amounts of hematite. The major gangue constituents are silt-size quartz and minor amounts of illitic clay. Two types of ore material are readily identifiable in the field—"hard ore" and "soft ore." The term "hard ore" is applied to that material in which the principal ore mineral is dense, hard, crystalline goethite. Its most striking physical characteristic in place is its coarse, broken,

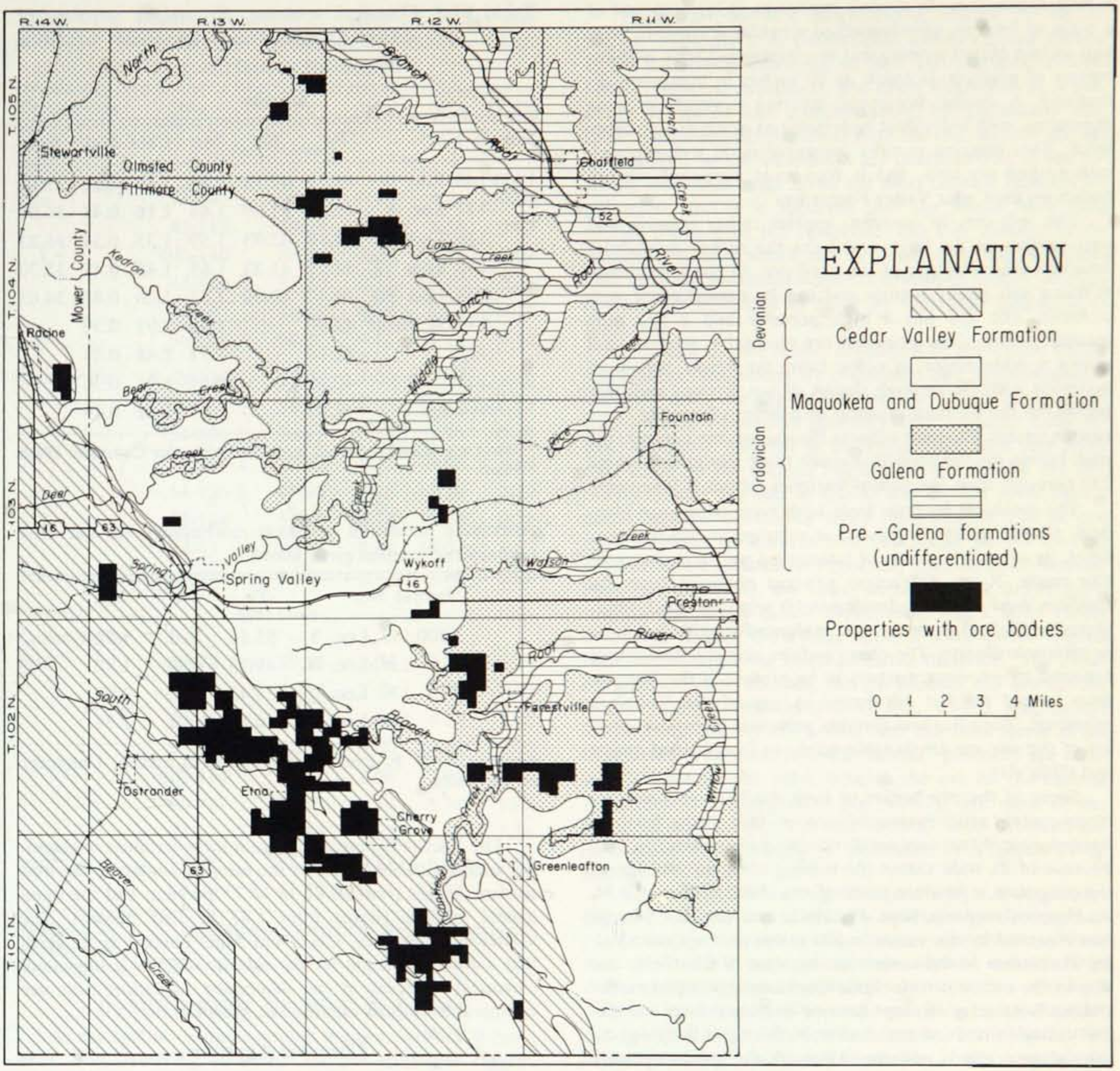


Figure VI-43. Bedrock geology of the Fillmore County district.

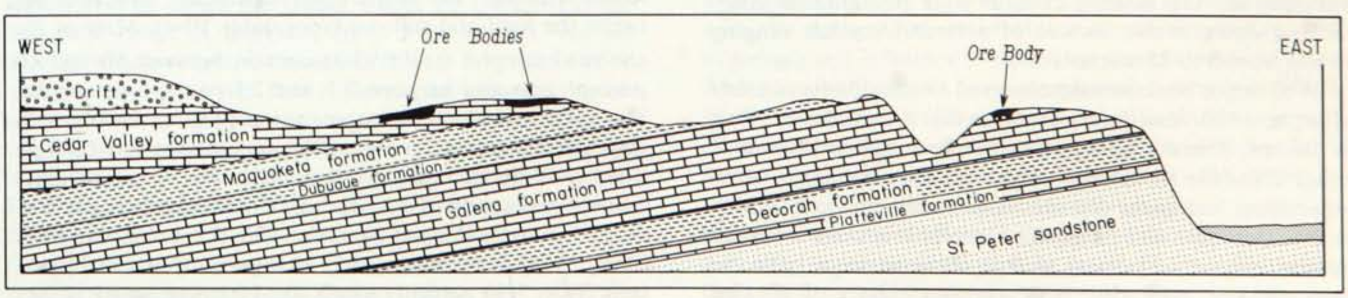


Figure VI-44. Generalized geologic section, Fillmore County district.

rubbly appearance. In typical exposures, it is composed of a mass of broken, closely-packed, angular fragments, one-half to two inches across, that are intermixed with nodular masses of goethite as much as 10 inches in maximum dimension. A distinct horizontal layering is visible in some exposures, with individual beds being as much as six inches thick. This layering is most conspicuous on well-exposed, rain-washed ore cuts, and is seen most commonly in ore bodies on the Cedar Valley Formation.

The soft ore, in contrast, appears rather massive and structureless in the field, and lacks the rubbly or nodular structure characteristic of the hard ore. In hand specimen, it has a soft punky texture and can be carved easily with a knife. The ore has a high porosity and a low bulk specific gravity. The principal ore mineral is goethite that shows a wide range in color from the bright yellow of ocherous goethite through shades of tan, brown, and dark brown, to the brilliant crimson of ocherous hematite. To a certain extent the color reflects the manganese content; the dark brown ore varieties have much more manganese (about 2.0 percent) than the yellow varieties (about 0.5 percent).

The insoluble residues from both types of ore are identical. Clastic quartz is present as subangular, silt-size particles, as are a few grains of intermixed pyrite in the same size range. X-ray diffraction patterns obtained from the residues show a poorly-developed 10 angstrom peak interpreted as illitic clay; there is no evidence for either kaolinite or montmorillonite. The chert nodules and silicified fossils reported by previous workers to be present in the iron ore were looked for but not found in any of the 50 mines examined. The silica and alumina present in chemical analyses of the ore are attributable solely to fine-grained quartz and illitic clay.

Some of the ore bodies of both the hard and soft ore types contain relict masses of siderite. One of the first pits opened near Etna contained several feet of siderite, and because of its wide extent the mining company considered shipping it as a separate grade of ore (John Owens, the M. A. Hanna Company, Sept. 15, 1962, oral comm.). Siderite was observed by the writer in two mines on the Cedar Valley Formation in the same vicinity, west of Chatfield, and also in the coarse ore stockpile from ores developed on the Galena Formation. Except for one specimen from the Cedar Valley Formation that had well-developed bedding, the siderite generally is massive. Although the specimens were examined carefully for evidence of replaced fossils or other structures which might be pertinent to the origin of the siderite, neither macrofossil nor chert remnants nor nodules were found. The siderite consists of a fine-grained interlocking equigranular mosaic of anhedral crystals ranging in size from 5 to 25 microns.

The major contaminant observed in the siderite consists of scattered silt-size quartz particles that are similar to those in the ore. There is no evidence for interstratified layers of either dolomite or calcite. Pyrite is rather common, often spherulitic, and generally occurs as small subhedral grains in the 10-micron size range. The insoluble residue from the siderite is nearly identical to that obtained from both the hard and soft ores; that is, it consists mainly of silt-size quartz grains, small pyrite aggregates, and illitic clay.

Table VI-3. Chemical analyses, in weight percent, of siderite samples and associated oxidation products.

Siderite

Sample No.		Fe	Fe^{++}	Mn	CaO	MgO	CO_2
1300.146	Loc. 8	44.27	42.14	1.42	1.41	0.53	35.62
.137	Loc. 8	42.20	37.09	1.64	1.16	0.45	31.96
.136	Loc. 48	43.98	42.93	1.59	1.35	0.51	36.83
.148	Loc. 37	44.75	41.22	1.65	1.45	0.47	35.70
.135	Loc. 50	40.82	40.09	1.55	1.20	0.82	34.65
.150	W.Plant*	42.66		1.61	2.01	0.39	
.151	"	42.18		1.57	2.48	0.37	
.152	"	42.25		1.55	1.91	0.36	
Average. .		42.88		1.57	1.62	0.49	

* Hand specimens collected from the Schroeder Company Wash Plant

Associated oxidation products correspond to the first four siderite samples given above

Sample No.		Fe	Fe^{++}	Mn
1300.145	Loc. 8	55.14	n.d.	1.72
.138	Loc. 8	49.73	0.12	1.42
.136	Loc. 48	54.43	0.05	2.08
.148	Loc. 37	56.78	n.d.	1.57

Analyses by V. E. Bye, Mines Experiment Station, University of Minnesota

The siderite is pure and there is little variation in chemical analyses (table VI-3). In the samples analyzed, the iron content ranges from 40.82 to 44.75 percent, and the manganese content ranges from 1.42 to 1.65 percent. The variation in Fe, Mn, CaO, and MgO content is so small that a common origin for the siderite samples is suggested. Chemical analyses of the associated oxidation rims are comparable to those obtained on the iron ores.

Chemical Analyses

The iron content of the ore bodies is remarkably consistent, and there is no significant difference between the ores on the Cedar Valley Formation and those on the Galena Formation, nor is there any significant difference between the hard and soft ore types (table VI-4). Most of the analyzed samples (table VI-4) contain between 50 and 60 percent iron and between 0.5 and 2.5 percent manganese. The silica and alumina content varies proportionately with iron and manganese content. The ore samples have a high silica to alumina ratio, which is indicative of a low clay-mineral content relative to quartz. The average composition of the iron ores shipped from the district for the year 1963 (Alm, 1964) is as follows: Fe, 53.6 percent; Mn, 0.98 percent; SiO_2, 7.56 percent; Al_2O_3, 0.57 percent; and P, 0.243 percent.

Table VI-4. Chemical analyses, in weight percent, of iron ore samples from mines on both the Cedar Valley and Galena Formations.

Description	No.	Fe	Mn	SiO_2	Al_2O_3
		Hard Ores			
Cedar Valley Fm.	51-94	39.74	0.63	4.41	0.24
	43-119	59.69	1.57	3.36	0.45
	60b-140	60.01	2.44	1.55	0.07
	43-120	51.70	0.22	9.70	2.21
	42-84	54.58	0.35	6.60	0.29
	36-68	56.97	0.67	4.26	0.31
Galena Fm.	2s-134	55.30	1.08	7.54	1.58
	2s-135	56.98	1.22	4.13	0.57
	54-106-1	56.17	0.14	3.90	0.49
	54-106-2	52.26	0.31	9.68	1.46
	54-106-3	49.95	2.35	7.39	2.19
	53-104	55.30	0.77	5.86	0.29
	24-43	44.53	0.86	18.02	3.35
	8-13	55.54	0.67	5.68	0.92
		Soft Ores			
Cedar Valley Fm.	39-72	59.53	2.08	2.31	0.24
	47-90	58.25	2.07	3.00	0.42
	49-91	51.94	1.87	4.23	1.01
	51-122	52.83	0.50	8.27	1.00
	60b-139	56.82	2.08	2.95	0.56
	40-73	56.50	1.59	3.71	1.04
	51-97	52.99	0.67	6.48	1.16
	37-69	55.22	0.33	6.60	0.68
	19-26	41.40	8.21	13.52	3.85

Analyses by V. E. Bye, Mines Experiment Station, University of Minnesota

DESCRIPTION OF ORE BODIES

As the spatial relations among the ore, bedrock formations, and overlying materials are not displayed clearly within any one mine, the postulated field relationships based on examination of more than 50 mining exposures are illustrated on the composite section (fig. VI-45). The ore bodies overlie either the Cedar Valley or Galena Formation, and range in thickness from 3 to 30 feet. An underclay which ranges in thickness from a few tenths of an inch to more than two feet is developed between the ore and the underlying carbonate rocks. The ore is locally overlain by decomposed Cedar Valley Formation, residual clays, or sediments of the Ostrander Member of the Windrow Formation. Both the Cedar Valley Formation and the Galena Formation beneath the ore generally are fresh, although they may have been changed to a sandy dolomite ranging in thickness from a fraction of an inch to several feet.

The underclay is a characteristic feature of the ore bodies. Typically it is thin-bedded and laminated parallel to the underlying limestone surface, and contains silicified fossil fragments, conodonts, and chert fragments. It is generally devoid of iron contamination, but commonly contains manganese-rich laminae. Both the microfossils and the macrofossils in the underclays are characteristic of the adjacent limestone units. Even the most delicate features are preserved, and it is apparent that the fossils have been neither transported nor reworked. The clay mineral in the underclay is illite, which also is found in the adjacent dolomite bedrock. Montmorillonite is developed locally, and appears to have formed in place by supergene alteration of illitic clay.

Although the ore bodies developed on the Cedar Valley and Galena Formations are chemically and physically similar, they differ in size and shape. The ore bodies on the Cedar Valley Formation generally have a greater areal extent, are more uniform in thickness, and have less relief than those on the Galena Formation; and several are distinctly tabular and can be traced laterally for more than a mile. Deposits containing more than 50,000 tons of ore were common. In contrast, the ore bodies on the Galena Formation are isolated and generally contain much smaller tonnages. Generally, the upper surface of the ore is quite smooth, has a few closed depressions, and a relief rarely exceeding 10 feet. On a large scale, it is somewhat convex beneath the overlying unconsolidated materials. The relief on the carbonate bedrock surface beneath the ore on the Cedar Valley Formation is small (fig. VI-46), and, except in the more easterly ore bodies near the Root River or its tributaries, prominent bedrock "horses" generally are lacking. In contrast, the relief beneath the ore on the Galena Formation is much greater (fig. VI-47), and most of the mines show prominent bedrock "horses," some of which are more than 30 feet high, as illustrated by Figures VI-48 and 49.

At a few localities along the western edge of the district (fig. VI-50), the ore is directly overlain by decomposed Cedar Valley Formation. The coarser size fractions from this decomposed material contain silt-size quartz, large drusy quartz aggregates, dolomite rhombohedra, and fragments of coarse chert. The sand- and silt-size fractions are composed of about 50 percent dolomite rhombohedra, the remainder being quartz and chert. The sharp crystal faces of the dolomite in these samples indicate that it has not been subjected to attrition, and suggest that it accumulated in place. The association of coarse aggregates of dolomite, aggregates of drusy quartz, and coarse fragments of chert with the dolomite rhombohedra indicates a complete lack of sorting and is further evidence for accumulation in place. The clay-size fractions from samples of decomposed Cedar Valley contain only illite. The lack of kaolinite is significant in view of the widespread occurrence of this mineral within the Cretaceous regolith in other parts of the state.

Residual clays are present on top of several of the ore bodies on the Cedar Valley Formation. The clay is thin, and generally is preserved in pockets and broad depressions on the upper surface of the ore. Except where contaminated by iron, it is a distinctive grayish green when wet and a

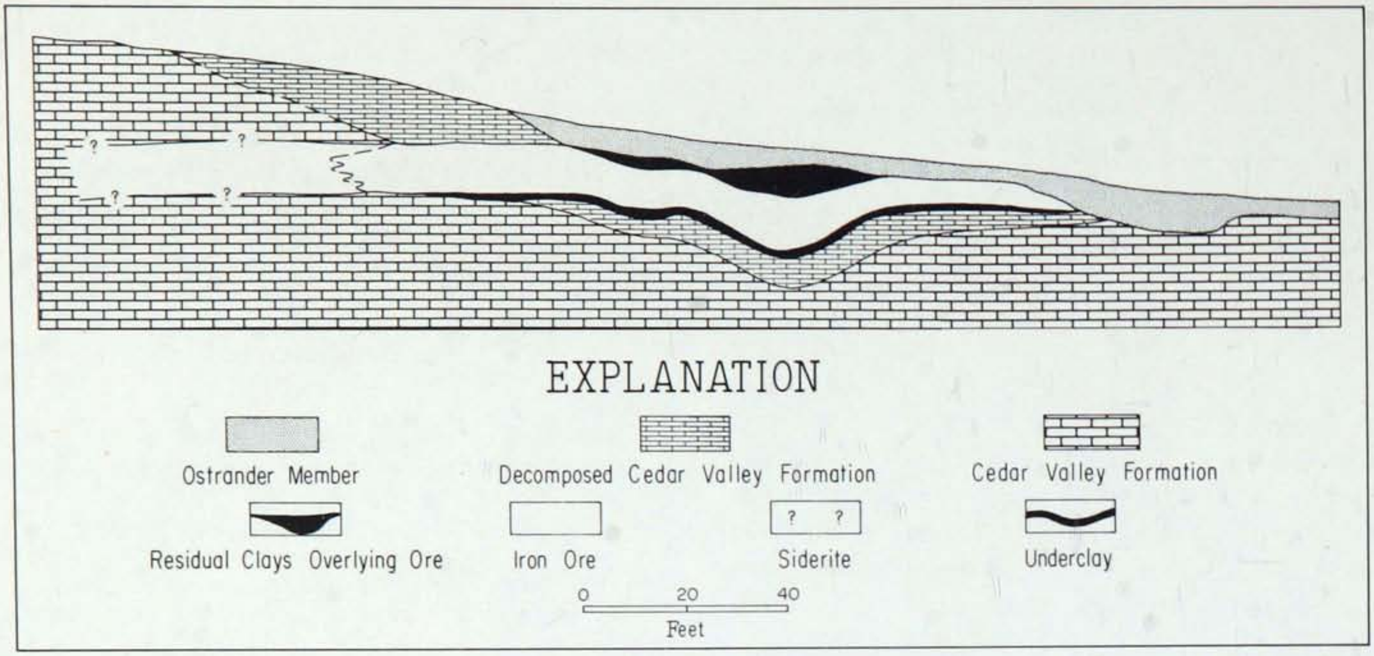


Figure VI-45. Composite section of a hypothetical iron ore body on the Cedar Valley Formation.

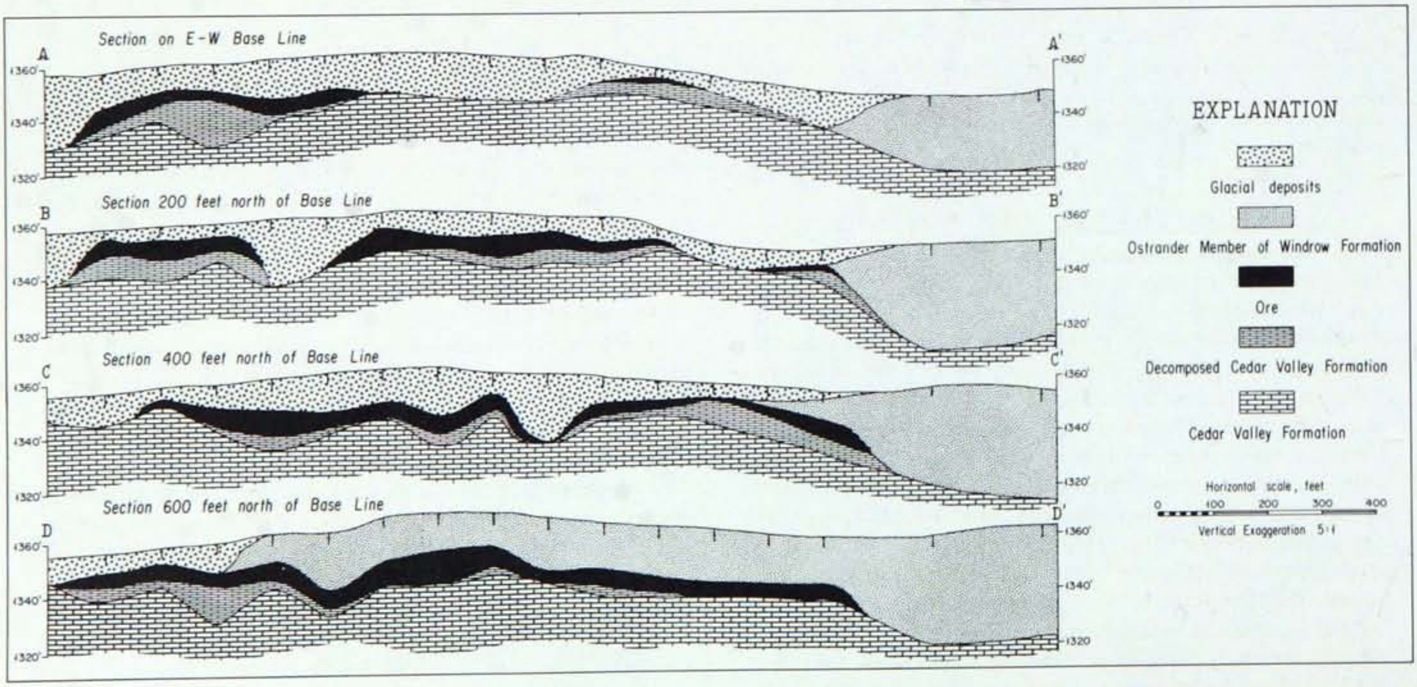


Figure VI-46. Section of main ore body in section 32, T. 102 N., R. 12 W.

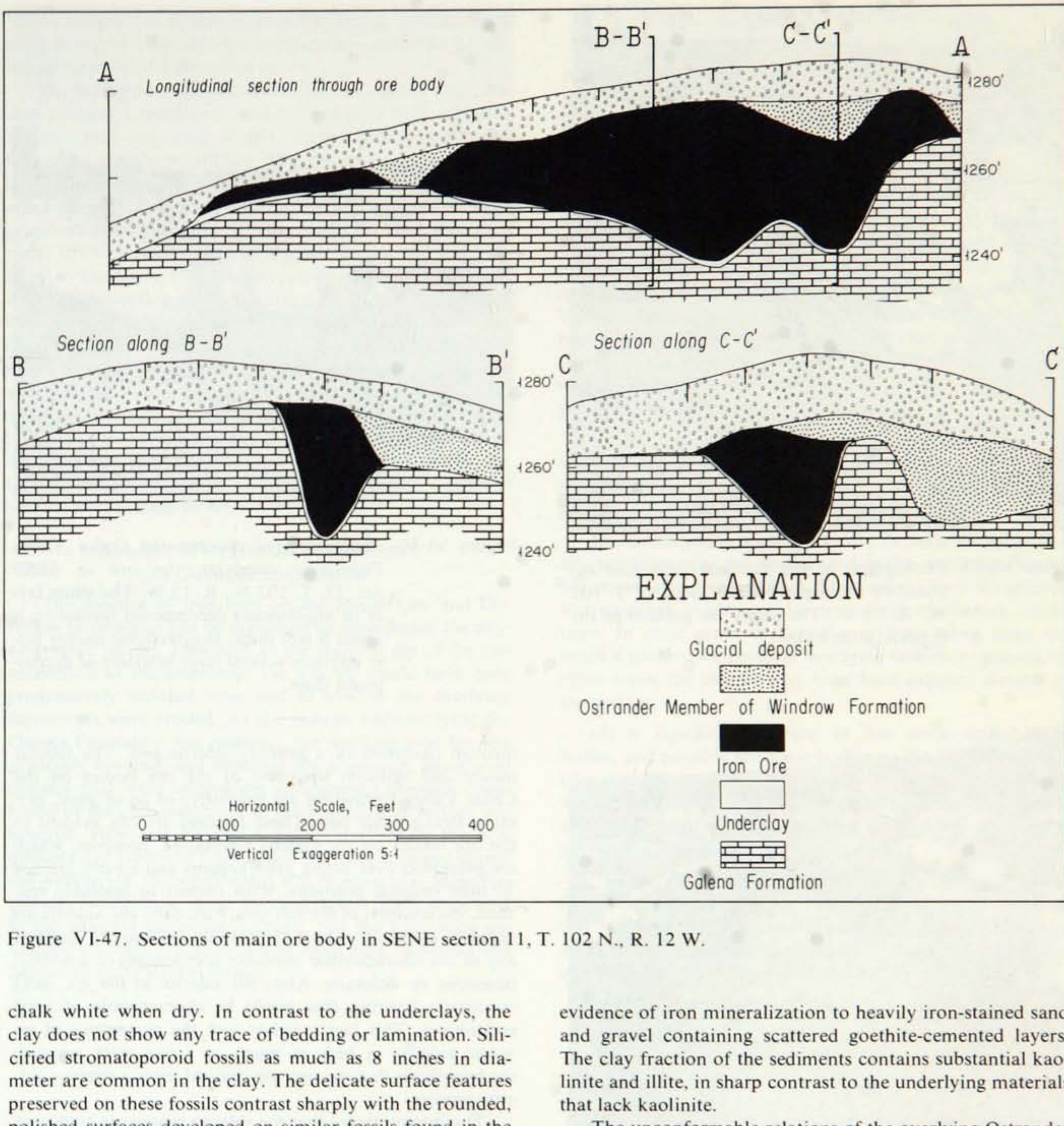


Figure VI-47. Sections of main ore body in SENE section 11, T. 102 N., R. 12 W.

chalk white when dry. In contrast to the underclays, the clay does not show any trace of bedding or lamination. Silicified stromatoporoid fossils as much as 8 inches in diameter are common in the clay. The delicate surface features preserved on these fossils contrast sharply with the rounded, polished surfaces developed on similar fossils found in the sediments of the Ostrander Member. Clearly, they are relict fossils left essentially in place during the weathering and removal of the original carbonate rock. The coarse fractions of the clay contain aggregates of drusy quartz, fine sand and silt particles, decomposed chert fragments and, rarely, a small amount of limonitic material. The clay-size fraction contains only illite.

Clays, sands, and gravels of the Ostrander Member of the Windrow Formation are widely distributed over the area, and traces of these sediments are found over every ore body. The sediments vary from white clay and silt that lack evidence of iron mineralization to heavily iron-stained sand and gravel containing scattered goethite-cemented layers. The clay fraction of the sediments contains substantial kaolinite and illite, in sharp contrast to the underlying materials that lack kaolinite.

The unconformable relations of the overlying Ostrander Member with the iron ore in a number of pits establish that the Ostrander is distinctly younger than the iron ores. The truncation of the ore bodies is illustrated by the cross sections shown in Figures VI-47 and 48. From the nature of the erosional surface, it is evident that deposition of the Ostrander was preceded by erosion of parts of the ore bodies.

ORIGIN

The field relations and chemical and physical characteristics of the iron ores are compatible with an origin

Figure VI-48. Photograph of the Galena Formation exposed by mining in NENW, sec. 33, T. 102 N., R. 11 W. The joints are parallel to the regional joint system.

Figure VI-49. Photograph of the Galena Formation exposed by mining in SENW sec. 9, T. 104 N., R. 12 W. The open cut shown parallels a joint and was originally 15 feet deep, but since has been partially refilled by rain wash.

Figure VI-50. Photograph of decomposed Cedar Valley Formation overlying the ore in SESE sec. 23, T. 102 N., R. 13 W. The white layer of argillaceous decomposed formation is about 6 feet thick; the overlying darker layer contains several large boulders of decomposed chert.

through oxidation of a primary siderite bed. The tabular nature and uniform thickness of the ore bodies on the Cedar Valley Formation are suggestive of an original, primary stratigraphic bed. These features are not evident in the ore bodies on the Galena Formation, however, which are preserved over active joint systems and clearly are not in their original positions. With respect to insoluble residues, the contents of the soft ores, hard ores, and siderite are virtually identical; neither the siderite nor the ore contains any of the characteristic insoluble components of a normal limestone or dolomite. Also, the siderite in the ore lacks concentric layering that would be characteristic of local concretions. This fact together with the occurrence of residual siderite in beds as much as 5 feet thick supports the hypothesis that the ore was derived from a massive siderite protore.

The interpretation that the siderite represents a primary sedimentary unit in the lower part (Solon Member) of the Cedar Valley Formation is based on the following observations: (1) decomposed Cedar Valley Formation locally overlies the ore; (2) residual clays derived from Cedar Valley locally overlie the ore; (3) underclays derived by weathering of the Cedar Valley lie beneath the ore; and (4) the ore bodies on the Cedar Valley Formation are tabular and are confined to the lower submember of the Solon Member.

The ore bodies developed on the Galena Formation are believed to be related to oxidation of a landward extension of the siderite bed responsible for the ore on the Cedar

Valley Formation. Probably, primary siderite was deposited directly on the truncated Ordovician formations during the transgression of the Devonian sea.

The textural features of the ore and associated clays, the clay mineral assemblages, and the regional field relations indicate that oxidation of the siderite took place under temperate climatic conditions. The ores and associated clays lack the features characteristic of humid subtropical weathering conditions, which are found in Lower and lower Upper Cretaceous deposits in other parts of the state (See Parham, 1970; Austin, 1963). If the ore bodies had been related to the Early Cretaceous episode of weathering, they would have developed the pisolitic textures characteristic of the other Cretaceous iron deposits. Also, the presence of the illite-montmorillonite clay mineral assemblage associated with the ores provides additional evidence that the iron ores were not exposed during the Early Cretaceous, for this assemblage is characteristic of temperate weathering conditions. Had they been exposed at that time, kaolinite would dominate the clay mineral assemblage, as it does in the Lower Cretaceous regolith of Minnesota (Parham and Hogberg, 1964; Parham, 1970).

Absence of Ore on the Maquoketa and Dubuque Formations

The absence of iron ores on the Maquoketa and Dubuque Formations can be explained by considering the physiographic factors. Inasmuch as the regional dip of the formations is to the southwest, the siderite would have been progressively exposed from east to west as the overlying formations were eroded. As the siderite bed overlying the Galena Formation was oxidized, the resulting iron ore was gradually lowered, and subsequently preserved, along joint systems that were undergoing enlargement through dissolution of the carbonate rock as the general land surface was lowered. However, because the joint systems in the Maquoketa and Dubuque Formations were not enlarged during the weathering, the siderite that was exposed over these formations would have stood as topographic highs compared to the adjacent carbonate bedrock, and the iron oxide developed through oxidation would have been dispersed and carried away by stream action.

CONCLUSIONS

The hypothesis presented for the origin of the iron ores in the Fillmore County district is based on a sequence of geologic events consistent with the known pattern of evolution of the midcontinent region from the early Paleozoic to the present. The hypothesis involves the following sequence of geologic events:

(1) The protore in the district was siderite, which originated as a primary sedimentary facies of the Solon Member of the Cedar Valley Formation. The siderite was deposited in a near-shore marine environment, either lagoonal or estuarine, which received sparse fine clastic materials. The siderite unit overlapped the truncated Maquoketa, Dubuque, and Galena Formations to the east, and the entire sequence was subsequently covered by the onlapping Devonian sea and buried beneath an unknown thickness of younger Devonian strata;

(2) During an intensive weathering interval associated with the Cretaceous marine transgression, the siderite remained buried beneath a substantial thickness of Upper Devonian and possibly younger rocks;

(3) Continued erosion during the Tertiary Period gradually uncovered the ore bodies, and as the overlying Devonian strata were stripped back from east to west the siderite was oxidized in place under temperate weathering conditions. In some areas, oxidation may have taken place beneath a continuous cover of overlying Ostrander gravels; in other areas, the siderite may have been exposed directly at the surface;

(4) A significant portion of the more easterly ore bodies, and possibly some ore bodies on the Maquoketa and Dubuque Formations, was removed by erosion, except where the ore was coincident with active joint systems that allowed it to be preserved with the gradual lowering of the land surface; and

(5) There is no evidence that the ore bodies were affected by glaciation in the Pleistocene, but it is probable that some of the more easterly ones were partially removed at that time by erosion related to the rapidly downward-cutting ancestral Mississippi River and its tributaries.

PRE-MT. SIMON REGOLITH

G. B. Morey

Rocks of Early or Middle Cambrian age have not been recognized in the upper Mississippi River valley, where the Mt. Simon Sandstone of Late Cambrian age directly overlies a variety of Precambrian rocks. Evidence for a prolonged period of weathering prior to Mt. Simon deposition is common in Wisconsin, where an apparently extensive regolith beneath the Mt. Simon Sandstone is exposed (Ostrom, 1966). Unfortunately, the basal part of the Mt. Simon Sandstone rarely crops out in Minnesota, but recent drilling penetrated a well-developed regolith on the subcrop granite surface near Monticello, in Wright County.

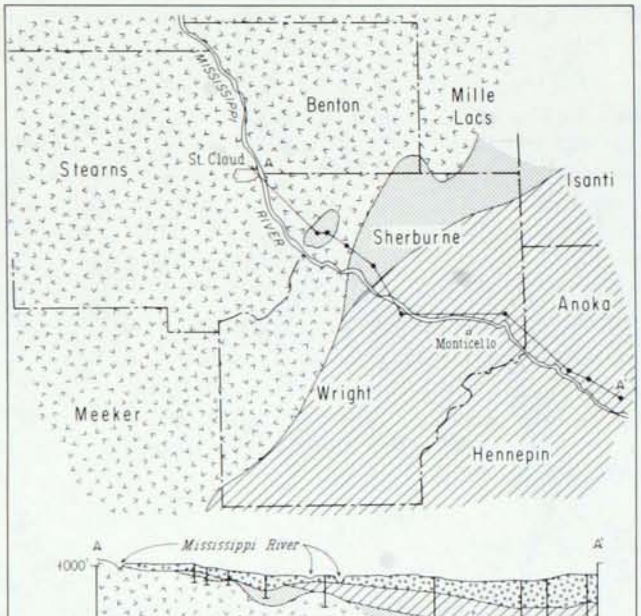


Figure VI-51. Generalized bedrock geologic map and cross section of Wright and Sherburne Counties and vicinity showing the distribution of underlying and overlying rocks to the regolith at Monticello.

In Wright County and adjoining Sherburne County, the Paleozoic strata overstep both the Keweenawan sandstone and an older granitic terrane (fig. VI-51). Although somewhat complicated by faulting, the strata dip southeastward into the Twin City basin but have been eroded near the edge of the basin, to expose Middle Precambrian granitic rocks near St. Cloud, in Stearns County. Numerous test borings near Monticello encountered as much as 60 feet of regolithic material beneath a white, poorly-sorted, medium- to coarse-grained, friable sandstone that contains fragmented remains of inarticulate brachiopods (R. E. Sloan, 1969, oral comm.); this and other similar sandstones in the subsurface of the area are assigned to the Mt. Simon Sandstone.

Regolithic material also overlies granite on the pre-Pleistocene subcrop surface in the vicinity of St. Cloud. However, this regolith is believed to be part of an extensive weathered surface that developed throughout much of Minnesota during early Late Cretaceous time (Parham, 1970, p. 46). Thus, within the area of Figure VI-51, there are two regoliths of distinctly different age developed on the same granite surface.

MINERALOGY OF THE REGOLITH AT MONTICELLO

Approximately 500 feet of core from 15 drill holes has been logged and sampled in detail. The sampled material includes the regolith itself, the underlying fresh or slightly weathered granite, and the overlying Mt. Simon Sandstone. A generalized stratigraphic section and associated clay mineral distribution is shown in Figure VI-52; in addition, several stratigraphically arranged X-ray diffractograms from one boring are shown in Figure VI-53. The less-than-2-micron clay fraction in the Mt. Simon Sandstone is characterized by kaolinite and well-ordered montmorillonite consisting entirely of expandable layers; in addition, the lower several feet of the formation contains a minor amount of illite, which most likely was incorporated during reworking of the underlying regolith. The upper 5 feet of the regolith consists of a soft, white, quartz-rich clay composed dominantly of kaolinite and well-ordered illite. This unit passes transitionally downward into less-weathered material that contains trace amounts of quartz, feldspar, and biotite in addition to kaolinite and illite. In hand specimens of this material, the feldspar and biotite show varying degrees of alteration; some grains appear to be nearly completely destroyed, whereas others appear fresh. The original granitic texture is preserved. The clay mineral assemblage is characterized by kaolinite, which decreases in abundance downward, and by a mineral having an X-ray diffraction pattern similar to that of mixed-layer illite/montmorillonite, described by Hower and Mowatt (1966). The abundance of

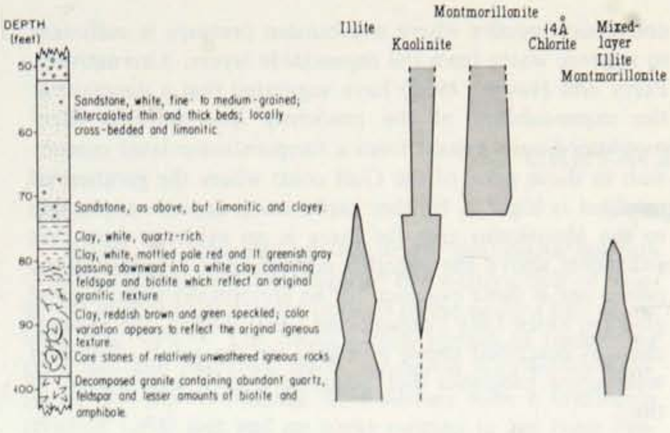
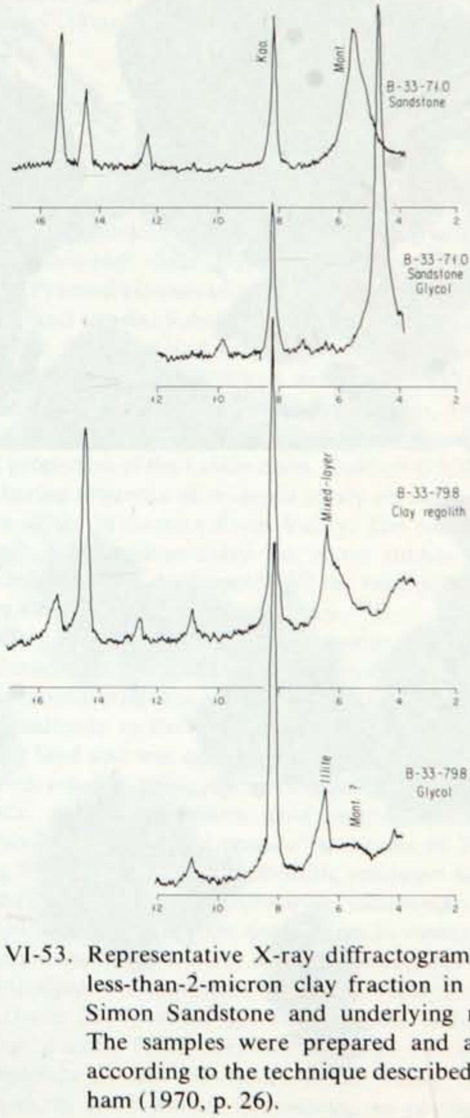


Figure VI-52. Generalized stratigraphic section and associated clay mineral distribution in one test boring from near Monticello.



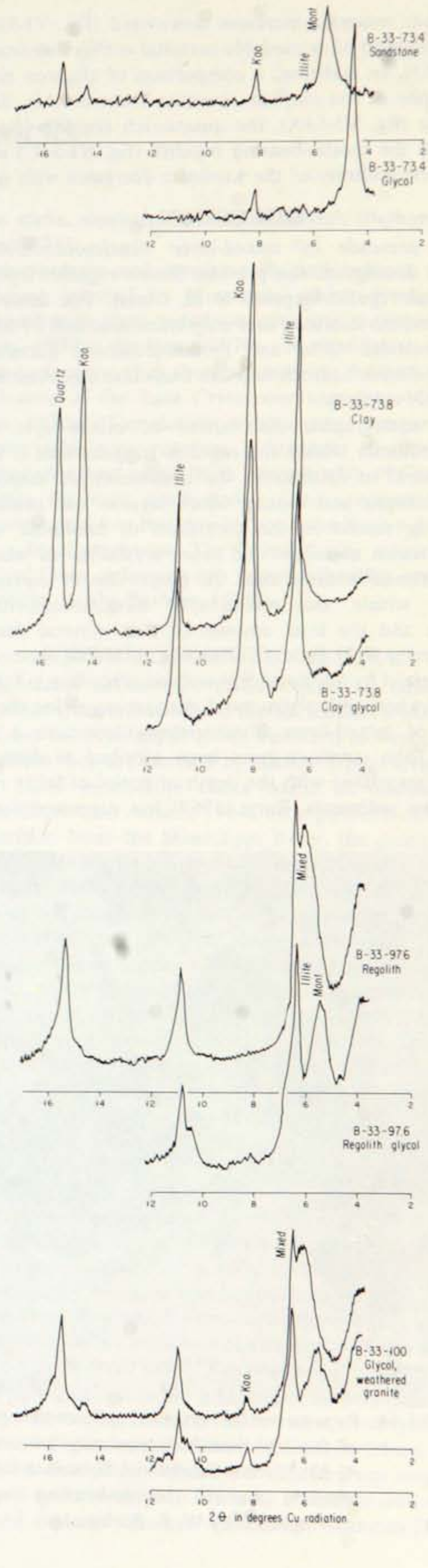


Figure VI-53. Representative X-ray diffractograms of the less-than-2-micron clay fraction in the Mt. Simon Sandstone and underlying regolith. The samples were prepared and analyzed according to the technique described by Parham (1970, p. 26).

mixed-layer material increases downward (fig. VI-53), as does the amount of expandable material within the structure (fig. VI-53). In addition, a comparison of electron microphotographs of the clay-size fraction from the Mt. Simon Sandstone (fig. VI-54A), the quartz-rich regolith (fig. VI-54B), and the biotite-bearing regolith (fig. VI-54C) shows that the crystallinity of the kaolinite decreases with depth.

DISCUSSION

The presence of mixed-layer illite/montmorillonite serves to distinguish the pre-Mt. Simon regolith from the Cretaceous regolith exposed at St. Cloud. The latter contains abundant kaolinite and only trace amounts of halloysite, muscovite, illite, and montmorillonite, particularly within the fresh bedrock-regolith transition (see Austin, fig. VI-56, this volume).

The stratigraphic distribution of mixed-layer illite/montmorillonite within the regolith suggests that it is the first mineral to form from the breakdown of amphibole and/or feldspar and biotite. More intense and prolonged weathering results in the formation of kaolinite, which becomes more abundant and more crystalline as weathering progresses, whereas both the proportion of expandable material within the mixed-layer illite/montmorillonite structure and the total amount of that mineral decrease progessively with time. Ultimately, a stable assemblage characterized by kaolinite and well-ordered illite is formed.

Somewhat similar structural changes involving the conversion of mixed-layer illite/montmorillonite to a more ordered illite structure have been ascribed to diagenetic changes associated with the depth of burial of fairly recent Gulf coast sediments. Burst (1969) has suggested that the conversion occurs where overburden pressure is sufficient to squeeze water from the expandable layers. Alternatively, Perry and Hower (1969) have suggested that a decrease in the expandability of the randomly interstratified illite/montmorillonite results from a temperature-related conversion in those parts of the Gulf coast where the geothermal gradient is highest. Neither explanation appears applicable to the Monticello area for there is no evidence here that sediments above the regolith accumulated to great thicknesses nor is there evidence for an abnormally high thermal gradient since Late Precambrian time. Rather, the mineral changes described above probably are the result of normal weathering processes that occurred over a long period of time.

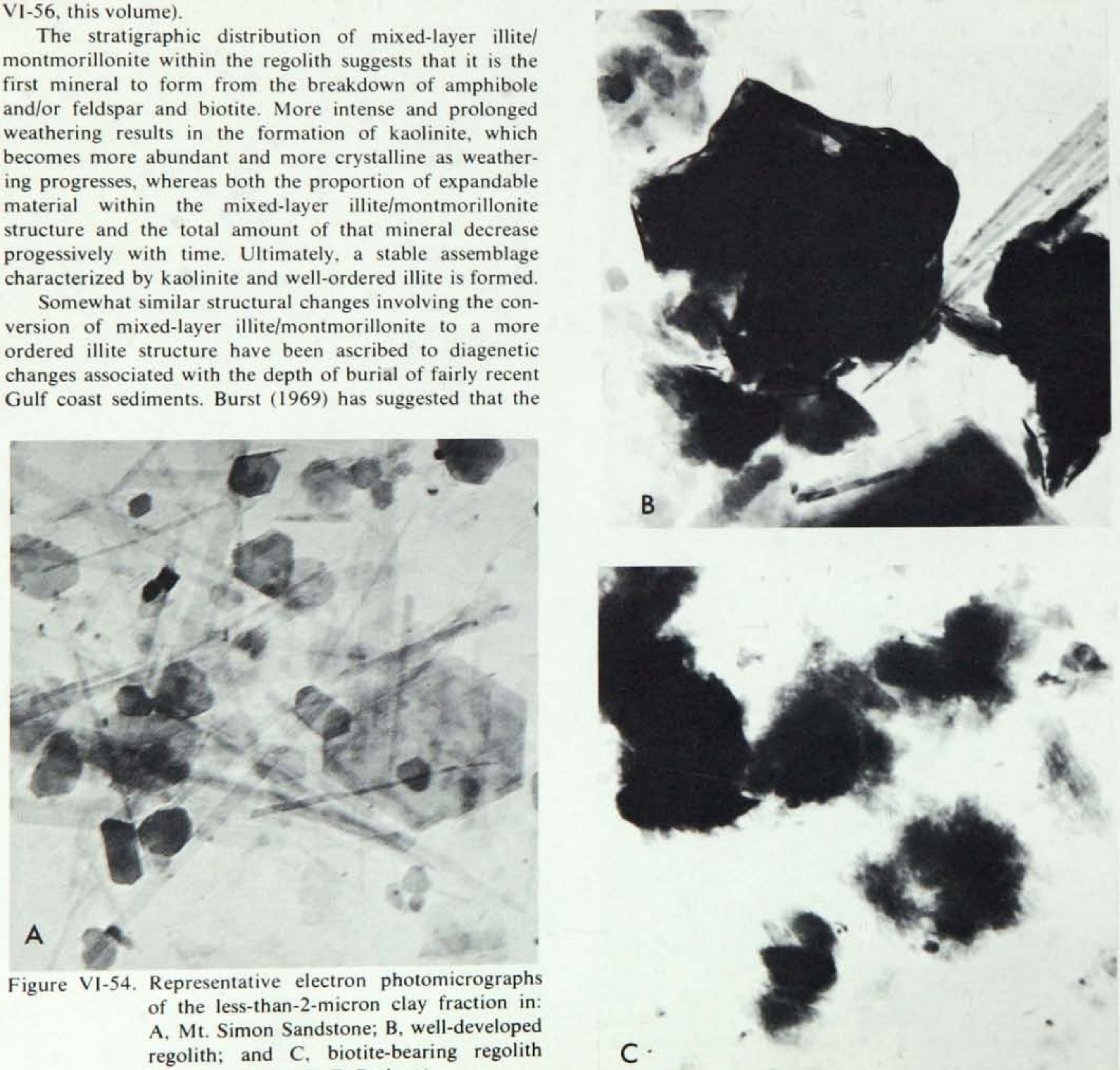


Figure VI-54. Representative electron photomicrographs of the less-than-2-micron clay fraction in: A, Mt. Simon Sandstone; B, well-developed regolith; and C, biotite-bearing regolith (photographs by W. E. Parham).

CRETACEOUS ROCKS

George S. Austin

Rocks of Cretaceous age are nearly continuous beneath thick Pleistocene drift throughout the western half of Minnesota and form numerous outliers in the eastern half of the state (fig. VI-55). The rocks, which consist of underlying residuum and overlying shale, sandstone, and minor limestone, rest unconformably on a surface with a maximum relief of 1,400 feet and on rocks ranging in age from Precambrian to Devonian. The basal residuum developed during a long interval of weathering that existed from sometime after Middle Devonian into earliest Late Cretaceous time. Upper Cretaceous marine and nonmarine sedimentary rocks overlie the residuum. In Minnesota, rocks of Cretaceous age are known principally from three areas: (1) the Minnesota River Valley in southwestern Minnesota; (2) southeastern Minnesota; and (3) northern and western Minnesota.

MINNESOTA RIVER VALLEY

The Cretaceous rocks exposed in the Minnesota River Valley, in southwestern Minnesota, include both residuum and overlying sedimentary rocks. As early as the 1890's, Winchell (1893) described several localities at which Cretaceous organic-rich shale, clay, and lignite overlie highly weathered Precambrian rocks composed primarily of a kaolinitic clay and quartz. Subsequent investigators (Grout and Soper, 1919; Emmons and Grout, 1943; Bergquist, 1943, written comm.; Prokopovich and Schwartz, 1957; Sloan, 1964; Parham and Hogberg, 1964; and Parham, 1970) described additional exposures and determined some of the physical properties of the kaolin clays. Goldich (1938) based his weathering sequence of minerals partly on studies of the residuum in the Minnesota River Valley. The recent paper by Parham (1970) summarized the above studies and described the sequential development of the residuum and the overlying kaolin-rich sedimentary rocks.

Probably no earlier than Late Jurassic time and definitely during Early Cretaceous time (Sloan, 1964), a widespread epicontinental sea existed in Minnesota and the states immediately to the west. The sea had a warming effect on the land and was a source for the precipitation necessary to develop a thick, chemically weathered zone. In Minnesota, water well drillers have reported penetrating thicknesses of "decomposed granite" in excess of 200 feet overlying "fresh granite." The kaolinitic residuum has been penetrated beneath Upper Cretaceous sedimentary rocks and glacial drift, or both, over a wide area in western Minnesota and eastern North and South Dakota, and is very likely equivalent to a weathered zone also known in Manitoba (Parham, 1970). Although the residuum is best preserved over granitic rocks, it is also found on Precambrian Sioux Quartzite in the Minnesota River Valley (Austin, 1970a) and, in south-central Minnesota, on younger carbonate rocks, sandstone, shale, and basalt (Parham, 1970; Austin, 1971).

Paleontologic and stratigraphic data indicate that the weathering interval in Minnesota which produced the abundant kaolin clays ended sometime in Cenomanian time (Austin, 1970a; Parham, 1970). A significant climatic change took place after the residuum was formed, prior to the advance of the Late Cretaceous seas into Minnesota (Austin, 1970a). The climate became more temperate and the water table rose, resulting in stagnant alkaline waters on the nearby land areas. With transgression of Cretaceous seas from the west, thick shales, some sandstone, minor limestone and one or two thin bentonite beds were deposited above a basal sandstone. The included clay minerals (fig. VI-56) of these sediments are dominantly illite and smectite (Austin, 1970a; Parham, 1970).

SOUTHEASTERN MINNESOTA

Nonmarine sedimentary rocks of Late Cretaceous age are exposed discontinuously in the southeastern part of the state, from the Mississippi River westward to west of Mankato, where they are covered by marine Upper Cretaceous rocks. These nonmarine rocks occur as discontinuous patches (Sloan and Austin, 1966) on a post-Devonian erosion surface. Near the Mississippi River, the glacial cover is relatively thin and patchy and the Cretaceous rocks lie at or near the surface. West of Mankato, the drift is thicker and, with few exceptions, the Cretaceous rocks are known only from those drill holes that penetrate bedrock.

The Cretaceous rocks of southeastern Minnesota have been included in the Windrow Formation, a formation originally proposed by Thwaites and Twenhofel (1921) to designate rocks lying above material of Paleozoic age in the Driftless Area of Wisconsin. Later, Andrews (1958) applied the name to all Cretaceous and pre-Cretaceous nonmarine rocks lying above Lower and Middle Paleozoic formations in the upper Mississippi valley. Andrews (1958) distinguished two members within the Windrow, a lower iron oxide-rich regolith, developed on pre-Cretaceous bedrock called the Iron Hill Member and an upper clastic unit subsequently identified as the Ostrander Member of the Windrow Formation (Austin, 1963; Sloan, 1964). The Iron Hill and Ostrander Members are equivalents of the kaolinitic residuum and the Dakota Formation, respectively, of western Minnesota. The Windrow Formation contains leaf imprints and carbonized wood. The presence of a shark's tooth found in place in a clay pit in Goodhue County (fig. VI-55) suggests that marine sediments at one time overlay the nonmarine sediments in southeastern Minnesota (Sloan, 1964).

Until recently, the Windrow Formation was considered important economically as a source of brown iron in Fillmore and adjacent Olmsted and Mower Counties (Stauffer

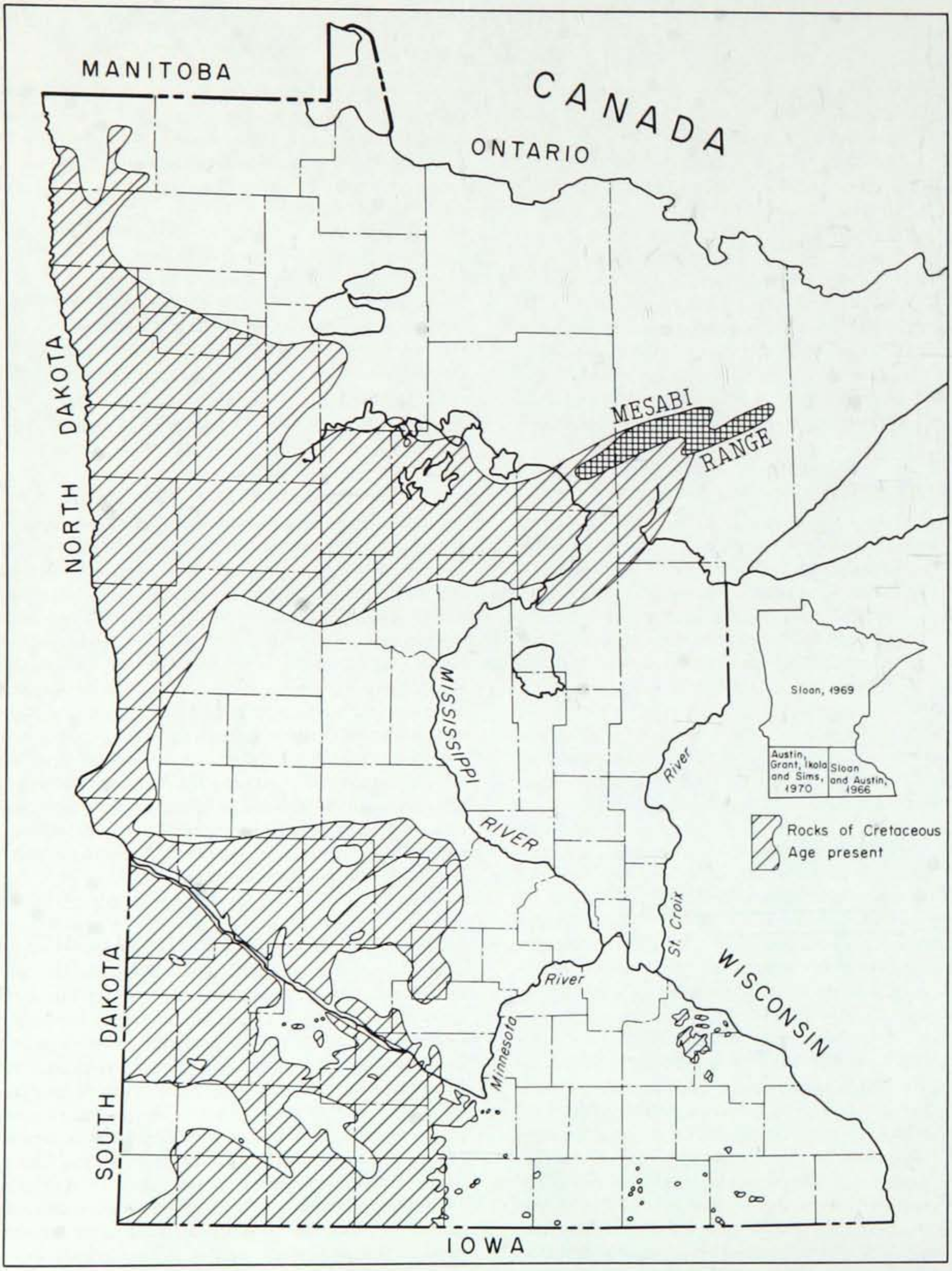


Figure VI-55. Generalized map showing the extent of Cretaceous rocks now present in Minnesota.

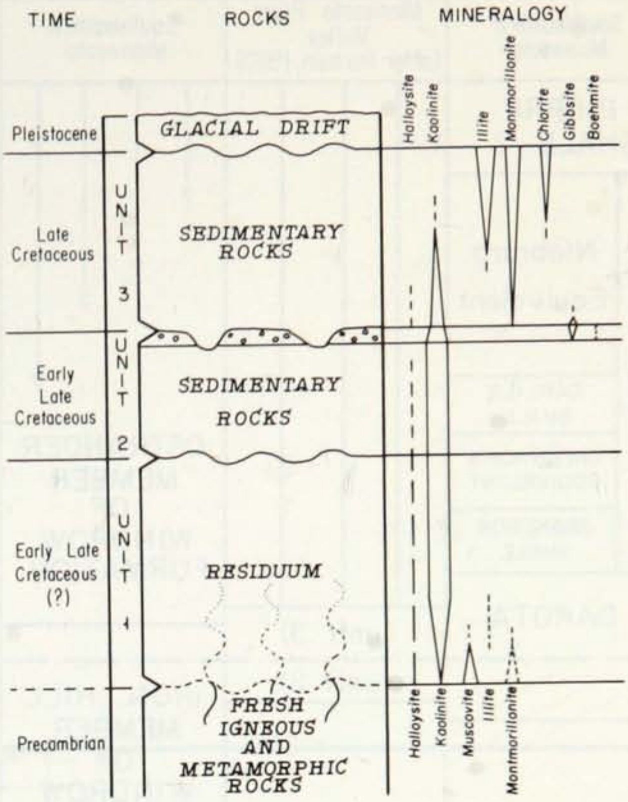


Figure VI-56. Vertical variation in clay mineral assemblages of the weathered residuum and the Upper Cretaceous sedimentary rocks of the Minnesota River Valley (after Parham, 1970).

and Thiel, 1944, 1949; Sloan, 1964), but Bleifuss (this chapter) has shown that the ores more likely are Tertiary in age. Similarly, until recently, the Ostrander Member of the Windrow Formation was a source of kaolinitic ceramic clay in Goodhue County, in southeastern Minnesota (Austin, 1963).

NORTHERN AND WESTERN MINNESOTA

Predominantly marine Cretaceous sedimentary rocks of the Coleraine Formation (Stauffer and Thiel, 1941), in northern Minnesota, and the Colorado Group, in western Minnesota, overlie the residuum. The Coleraine Formation is poorly exposed, and is best known from exposures in the open-pit iron-ore mines in the western part of the Mesabi district. It consists of iron-ore conglomerate, shale, and sandstone that form a mantle over Precambrian bedrock on the Mesabi range, and it grades laterally eastward from dominantly marine to nonmarine. Studies of Cretaceous rocks exposed on the Mesabi range include those by Bergquist (1944), Burgess (1955, unpub. M.A. thesis, Univ. Missouri), McGill (1955, unpub. M.S. thesis, Univ. Minnesota), Everett (1956), and Owens (1956).

Bergquist (1944) described a molluscan fauna from the Mesabi exposures consisting dominantly of species of *Ostrea* and *Exogyra* in the west and of *Cardium* and *Tapes* in the east. Gastropods are rare in the west but become more common to the east. The invertebrate fauna comprises 60 species and varieties, and is considered equivalent to that in the lower Benton shales of the Colorado Group.

Dominantly marine shales of Cretaceous age from other parts of northern and western Minnesota have been reported by many authors (Grout and others, 1932; Thiel, 1947; Bolin, 1956; Rodis, 1963; and Sloan, 1964), and are generally known from small, poorly-exposed outcrops or from drill cuttings. Fossils contained in these rocks include fish teeth, scales, bone fragments, and pelecypods, cephalopods, foraminiferans, radiolarians, and ostracodes.

In southwestern Minnesota, a white or pale yellow sandstone of variable thickness containing thin lignite beds commonly lies above the residuum and below the dominantly marine shales. The sandstone unit, identified as the Dakota Formation, was derived from the weathering products of the Sioux Quartzite, as well as from granitic rocks exposed on the Transcontinental Arch during Cretaceous time. The Dakota Formation is interpreted as a deltaic deposit in South Dakota (Schoon, 1965) and as a continental deposit in Minnesota (Sloan, 1964; Austin, 1970b), and is progressively younger to the east (Schoon, 1965). As the Late Cretaceous epicontinental sea transgressed, the clastic sediments became finer grained and marine shales succeeded the Dakota Formation. Sloan (1964) indicated that nearly continuous deposition of marine shale took place in western Minnesota from Cenomanian to Santonian time; possibly, the Pierre Shale of Campanian time is present beneath the glacial drift in southwestern Minnesota. In western Minnesota, the maximum known thickness of Cretaceous strata, approximately 600 feet, occurs in Lincoln County, along the South Dakota-Minnesota border.

SUMMARY OF AGE OF CRETACEOUS ROCKS

The sedimentary Cretaceous rocks in Minnesota (fig. VI-57) can be correlated with the Dakota Formation, the Colorado Group, and perhaps the Pierre Shale of the western interior United States. Although the age of the underlying residuum developed on pre-Cretaceous rocks in Minnesota is not fixed firmly, most likely it formed during Jurassic or Early Cretaceous time, during the encroachment of warm epicontinental seas from the west, and possibly as late as Early Cenomanian time. Later, during Cenomanian time a climatic change halted development of the kaolinitic residuum over much of Minnesota. As the epicontinental sea advanced across Minnesota under more temperate climatic conditions, thick beds of predominantly shaly strata were deposited on the previously formed residuum and nonmarine sedimentary rocks.

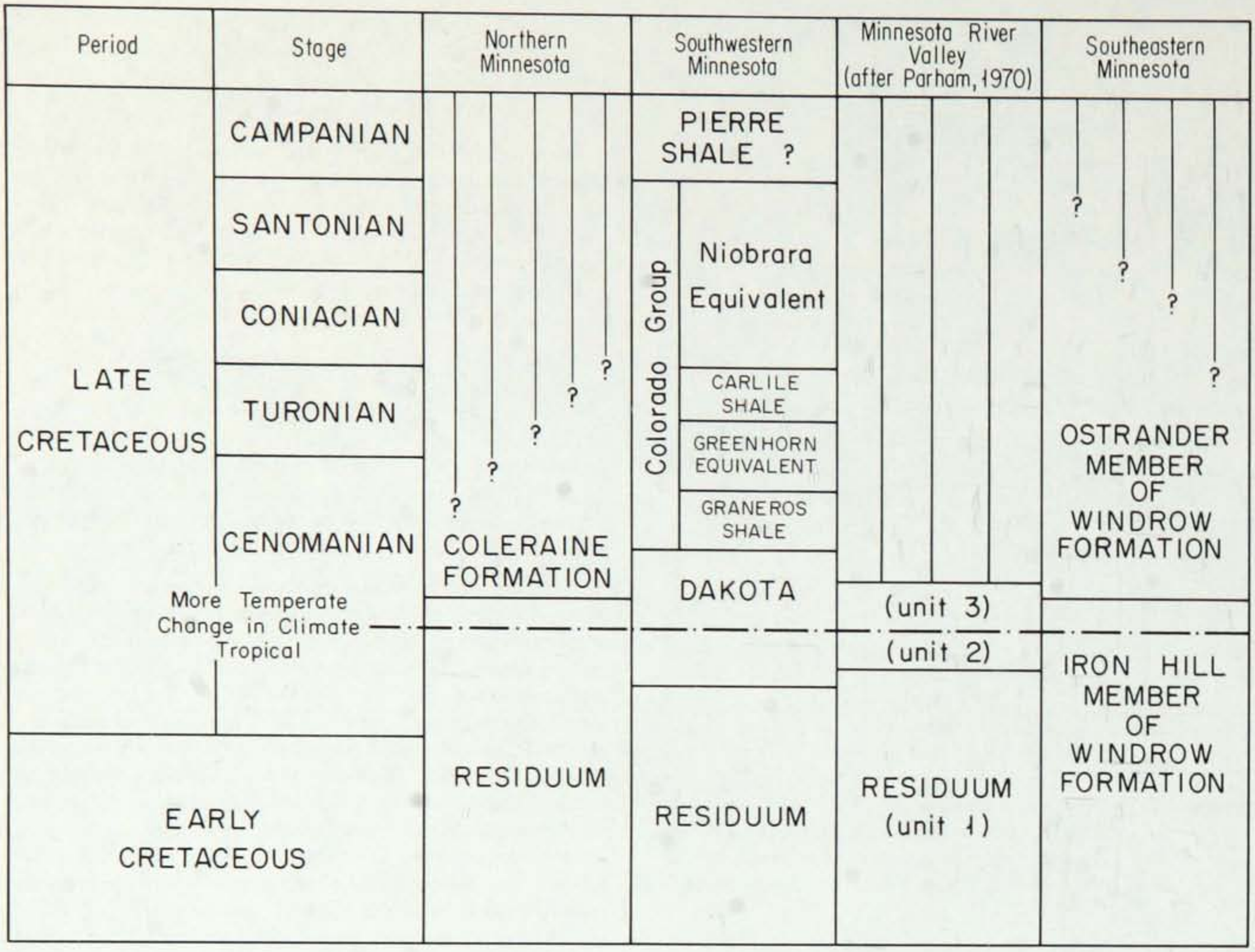


Figure VI-57. Correlation chart of the Cretaceous System in Minnesota (modified from Sloan, 1964).

Chapter VII

CENOZOIC

Quaternary History of Minnesota, H. E. Wright, Jr.
Quaternary Geology of Southwestern Minnesota, Charles L. Matsch
Physiography of Minnesota, H. E. Wright, Jr.

QUATERNARY HISTORY OF MINNESOTA

H. E. Wright, Jr.

The appellation "Land of 10,000 Lakes" epitomizes the Minnesota landscape. To the geologist, the landscape immediately suggests a history of recent glaciation, because, with the possible exception of the limestone terrane of Florida, the youthful glacial landscapes of the northern Great Lakes region—Minnesota, Wisconsin, northern Michigan, and adjacent parts of Canada—provide the greatest concentration and diversity of lakes on the North American continent.

Yet a close look at even a highway map of Minnesota shows that the distribution of lakes is by no means uniform throughout the state (Wright, this chapter, fig. VII-24). Some regions have none, for example the Red River Valley area, whereas others are pock-marked with water-filled depressions of all sizes and shapes. The distribution of lakes is one of the principal clues to the geologic history of Minnesota during the glacial period and subsequent time, and it provides one of the most direct criteria for subdividing the state into discrete geomorphic areas.

Not only the lakes, but much of the river system in Minnesota is related to glaciation. The upper course of the Mississippi River, for example, from its headwaters to St. Paul, depends on the details of the glacial morphology, as do the courses of its tributaries. The Minnesota River and Red River of the North, on the other hand, flow along preglacial lowlands, and the middle course of the Mississippi, from St. Paul to southern Illinois, may be largely localized by the preglacial topography.

Although Minnesota was covered by ice sheets several times during the glacial period, the landforms and surficial deposits for most of the state record only the last (Wisconsin) glaciation. But the glacial record is highly complex, because of the interaction of several distinct ice lobes protruding from the front of the great ice sheet which covered most of Canada for tens of thousands of years. The protrusions were localized primarily by the preglacial bedrock lowlands, which had formed in response to the differential resistance of various rock types to preglacial erosional processes. So it is appropriate that a consideration of the glacial history be preceded in this volume by descriptions of the bedrock geology.

But glaciation is not the last chapter in the geologic history of Minnesota. In the 10,000 or more years since glacier ice left the area, erosion and deposition have modified the landscape. Erosion has generally been relatively minor, however, until the modern epoch of accelerated soil erosion in agricultural regions. Deposition of eroded material has been largely confined to lakes and other depressions, and in many areas this has been combined with deposition of organic detritus originating largely within the lakes themselves. Many lakes have been converted to bogs and marshes in this way, and of course such conversion is the ultimate destiny of most lakes, at least those that have no rapidly eroding stream as an outlet. The stratigraphic succession of microfossils in the sediment of such lakes provides both a direct and an indirect record of landscape evolution that in some cases exceeds in detail the record available from study of landforms or of physical stratigraphy. Knowledge of the nature and chronology of landscape development since the glacial period gives us perspective in viewing the changes that occurred during the glacial period. It also helps us visualize the landscape during the long interglacial and preglacial time intervals, when Minnesota was subject to geomorphic denudation under the influence of probably similar climatic and vegetational environments.

HISTORY OF INVESTIGATIONS

The Winchell Era

N. H. Winchell, the first Minnesota state geologist, with the assistance especially of Warren Upham, systematically studied and mapped the glacial geology of the state during the period 1872-1895, and subsequently they both published papers on various aspects of the glacial geology for another 20 years. Their work involved primarily county-by-county surveys of glacial landforms. The overall history of glaciation of the state is somewhat difficult to work out from their prose, which involved primarily descriptions and explanations of local topographic features. The absence of deep exposures, however, had made it difficult for them to gain an appreciation for stratigraphic relations, which have since provided the key to understanding glacial history in many areas.

Winchell and Upham recognized the existence of two ice lobes, one from the northeast (Lake Superior lobe, now called the Superior lobe) and one from the northwest (Minnesota lobe, now called the Des Moines lobe). The interaction of these two lobes was not worked out fully, but in general it was thought that they were contemporaneous and that they came together to produce the broad belt of high, morainic topography, now called the Alexandria moraine complex, that extends through west-central Minnesota. A total of 12 numbered moraines was traced across the state (Wright, 1962, fig. 1), but it was not always clear which ice lobe was involved in any particular segment.

The relations were seen to be particularly complex north of the Minneapolis area, where a great offshoot of the Minnesota lobe extended eastward to Wisconsin. This offshoot, called by them the Chisago lobe and later by Sardeson the Grantsburg sublobe of the Des Moines lobe, was thought to have followed in part a retreat of the Lake Superior lobe.

It was postulated that meltwater produced by wastage of the Lake Superior lobe brought increased evaporation in the peripheral area, resulting in increased local snowfall and thus in the advance of the nearby Chisago lobe. This was a short-lived advance, however, and as the ice withdrew, meltwater from the eastern and western lobes constructed what is now known as the Anoka sandplain. This feature was later reinterpreted by Sardeson (Leverett, 1932) as an area of sand dunes, but still later the work of Cooper (1935) demonstrated that a fluvial origin, as supposed by Upham, is much closer to the truth.

Subsequent retreat of the western ice resulted in the formation of Glacial Lake Agassiz, but in Upham's view the lake basin was first opened by ice retreat from west to east (Upham, 1896) rather than from south to north, as Leverett later believed.

Speculations on regional climatic relations were not common in the work of Winchell and Upham. These were rather the days of data gathering. The regional glaciation of the continental interior had not been worked out well enough to reveal significant patterns in space or time. One has the impression that the entire effort of geologic description and mapping was focused on the county unit, and that once this effort was completed there was no real occasion for integration and reflection, with the possible exception of Upham's Lake Agassiz monograph.

Winchell, as Director of the Minnesota Geological and Natural History Survey, was equally active in studying the bedrock and in evaluating the economic potential of mineral deposits. Certainly one of his most important geologic studies, however, was the determination of the rate of retreat of St. Anthony Falls on the Mississippi River, and its relation to late- and postglacial chronology. His estimate of the length of time since the retreat of ice from Minnesota, about 8,000 years, has proved remarkably close to the figure accepted today on the basis of radiocarbon dating.

Winchell also sponsored studies of the flora and fauna of the state. Upham, even though a geologist, produced a catalogue listing the flora of Minnesota (in Winchell, 1884b), and, in the same volume, the only known map of the natural vegetation of the state before the time of extensive land clearance—a map very useful in assessing the extent of subsequent agricultural land clearance and other disturbances. Winchell himself had a continuing interest in Indian artifacts and ethnography and always included archaeological notes in his county reports. In fact, after he retired from the state survey, he compiled an ethnographic survey for the Minnesota Historical Society. The productivity and dedication of Winchell in these pioneer years of studying the geology and natural history of Minnesota can be flavored in the quotations reproduced by Schwartz (1964), or more effectively by even a casual reading of the annual reports and the massive six-volume Final Report. The footnotes are particularly revealing. The lists of spot elevations and of localized outcrops and the highly generalized descriptions of the landscape imply a primitive condition for available base maps and knowledge of terrain patterns. These general descriptions are extremely useful in certain respects, however, because they embody the only existing summaries of the undisturbed vegetation of Minnesota—especially valuable in the areas soon destined for forest clearance as settlement expanded (Waddington, 1969).

Although most of Winchell's efforts are represented by the widely-circulated state publications—the Final Report was issued in 5,000 copies—he published separate papers in national journals. For example, in 1901 he completed a paper systematically describing the 21 proglacial lakes then known in Minnesota.

Warren Upham, who had started his glacial studies in New England, left the Minnesota Survey in 1885 but continued to work in Minnesota under the auspices of the U.S. Geological Survey. His major work was a study of Glacial Lake Agassiz, published as a huge monograph (Upham, 1896) rivalling in quality G. K. Gilbert's contemporaneous great work on another Pleistocene lake—Bonneville in the western desert country.

The Leverett Era

With the retirement of Winchell and Upham soon after the turn of the century, and the official end of the state geological survey, glacial studies were essentially dormant in Minnesota until Frank Leverett of the U.S. Geological Survey became active in 1906.

Leverett had worked on the glacial geology of several of the other Great Lakes states, having published monographs on Illinois, Michigan, and Indiana, and (with Taylor) on the Great Lakes themselves; his colleague W. A. Alden had worked contemporaneously in Wisconsin. Only Iowa was studied somewhat independently, largely by George F. Kay and other members of the Iowa Geological Survey. The dominance of Leverett in the picture of Great Lakes geology for the period from 1906 to 1935 is therefore understandable, as is perhaps his controversy with Kay over certain correlation problems (see Matsch, this chapter).

Leverett's era in Minnesota glacial geology was essentially a one-man act, although he had the assistance of Frederick B. Sardeson for some of the studies, particularly in the Minneapolis/St. Paul area and in the outlet area for Glacial Lake Agassiz in westernmost Minnesota. Sardeson published separate folios on both these areas for the U.S. Geological Survey, and in the process he evaluated and revised Winchell's estimates for the rate of retreat of St. Anthony Falls.

Leverett worked steadily in the field in Minnesota, from 1906 to 1912, and when the Minnesota Geological Survey was reactivated in 1911 by W. H. Emmons the work was accelerated by the addition of Sardeson to the program. Field work continued sporadically until 1923. The first general publications were issued by the state survey (Bulletins 12-14, 1915-1920), and emphasized soils and agricultural conditions. They include a large colored map of surficial deposits that is essentially the same as that in Leverett's final monograph for Minnesota and parts of adjacent states, published much later by the U.S. Geological Survey (Leverett, 1932). Leverett had published in 1929 a monograph on the northeastern area under the title "Moraines and Shore Lines of the Lake Superior Region," which has separate colored maps but a text that is largely repeated in his 1932 report, as far as the Minnesota area is concerned.

Meanwhile, the development of his ideas during the 26-year period between the beginning of his field work and the publication of his final monograph can be followed in his numerous short papers in national journals.

Leverett's big colored map of the state stands today as a very serviceable general map of surface material, even though concepts of glacial history have changed considerably. For elevation control in mapping, he placed much reliance on level lines for railways and primary surveys, as had been done in the Winchell era, but the accuracy was great enough for some of the difficult problems about glacial lake levels and outlets to be worked out.

The lack of exposures, plus the contemporary traditions of glacial mapping, resulted in an emphasis in Leverett's work on the morphology of glacial features rather than on the lithology or stratigraphy. Many glacial features were thus interpreted on the basis of form rather than content, and even the forms had to be estimated rather than measured, because of the lack of topographic maps. Consequently some mistakes were made. For example, the Beroun moraine, which is mapped as a long narrow recessional moraine through east-central Minnesota, is actually in large part an esker complex. In some segments it is merely the side of a broad subglacial erosional valley, which has a hummocky appearance when viewed from the valley side. But in general the boundaries of Leverett's map units are remarkably accurate, considering the difficult access to much of the country, the lack of topographic maps, and the lack of exposures.

Leverett's fine geologic map of the state was fully exploited in his monograph. In contrast to the county reports of Winchell and Upham, Leverett's publication was a fully integrated and systematic description of glacial features of the entire state, leading to a detailed consideration of the geologic history. The integration may in part reflect the publication requirements and standards of the U.S. Geological Survey, which during the early decades of this century produced many such monographs covering the geology of broad areas. At the same time, the pace of glacial studies in the Great Lakes region had increased since Winchell's day, with generalists and textbook writers like T. C. Chamberlin tracing patterns of glacial and interglacial features throughout the region and establishing a firm basis for the concept of multiple glaciation. Workers in the states along the margin of glaciation, where evidences for multiple glaciation are stronger than in Minnesota, took the lead in naming and correlating drifts from state to state; among these are men like Kay, Calvin, Shimek, Alden, Leighton, and Trowbridge, all of whom worked in Iowa during this period. Because he had had previous experience in mapping drifts in Illinois and Indiana, Leverett joined in the discussions, in fact one might say became entangled, especially with respect to the placement of Iowan drift in the glacial sequence, and he published many papers discussing matters of correlation, growth and development of ice sheets, and related general problems. He even spent a year in Europe comparing glacial sequences there with those in the Great Lakes region.

The Iowan drift had fairly extensive exposure in Minnesota, so discussion of its status was pertinent in Leverett's 1932 monograph. He held out for recognizing the Iowan as a separate glaciation, between the Illinoian and the Wisconsin, in contrast to the views of the Iowa geologists, who favored placing the Iowan as the first part of the Wisconsin. The controversy became at times acrimonious (Leverett, 1939). He finally yielded in a closing remark in his last published paper (1942). Subsequent events have shown that the Iowan drift is probably simply eroded Kansan drift (Ruhe, 1969), and that the Iowan loess, formerly considered equivalent to the Iowan till, is largely contemporaneous with the maximum of the main Wisconsin glaciation.

The self-containment of Leverett's monograph seems to go too far in some respects, for it ignored much of the work of Winchell and Upham. It introduced to Minnesota entirely new terminology of glacial drifts, ice lobes, moraines, and other features, with little discussion of older terms and concepts. Earlier work was largely dismissed with the following comment (Leverett, 1932, p. 2): "In publications of the Geological and Natural History Survey of Minnesota under the direction of N. H. Winchell, appearing at intervals from 1872 to 1900, a large amount of information as to moraines and other glacial features is presented, and a nearly complete mapping of moraines through most of the counties is shown. It has been found, however, that certain correlations of moraines need revision. This is true of some correlations where but little correction is required in the position or course of the moraine."

The Post-Leverett Era

Following the completion of the investigations of Leverett and Sardeson, glacial studies were inconspicuous for many years. It remained for a botanist, W. S. Cooper, to contribute the most significant publication of these years. He became interested in the plant ecology of the Anoka sandplain. Upon investigating the background geology of the area, he found difficulties with Sardeson's interpretation of the area as a vast region of sand-dune accumulation. Consequently, he launched a complete restudy of the glacial history of the area (Cooper, 1935), and never did reach the point of working on the plant ecology.

After the Second World War, a study of the glacial features of Cook County in northeasternmost Minnesota was undertaken by R. P. Sharp (Grout and others, 1959), and a project on the glacial geology of Dakota County in southeastern Minnesota by Ruhe and Gould (1954). These were followed by my own studies, in conjunction with students, mostly in the eastern and central part of the state. This work is summarized in the present chapter. Although much progress has been made in working out the stratigraphic succession and the complicated relations among ice lobes, a program of systematic mapping was started only recently, and Leverett's 1932 map remains the only compilation for the entire state. A somewhat revised version of this map later was produced for the map entitled "Quaternary Geology East of the Rocky Mountains" (Flint and others, 1959).

BEDROCK CONTROL

The course of glaciation in the Great Lakes region and the elucidation of its history both depend indirectly on the bedrock geology, for the erosional resistance of the bedrock determined the location of the preglacial lowlands that guided the ice lobes protruding from the ice sheet, and the varied lithology of the bedrock provided the raw materials by which the gross direction of ice movement can be deciphered. Each of the Great Lakes occupies a preglacial lowland that is clearly defined by the limits of erosionally non-resistant rocks, as may be seen on a geologic map. Lake Michigan and Lake Huron, for example, follow the belt of Devonian shales around three sides of the Michigan structural basin, and Lake Erie follows the same belt on the northern edge of the Allegheny structural basin. Below these sedimentary formations are resistant Silurian dolomites, underlain in turn by weaker Ordovician rocks. This combination produces the Dorr Peninsula west of Lake Michigan and then the Greenbay Lowland, which localized the Greenbay glacial lobe. Farther east the same combination forms Manitoulin Island and Georgian Bay, and also the escarpment at Niagara Falls and the Lake Ontario basin.

The same type of control prevails in the Minnesota area, although the bedrock structure is different (fig. VII-1). Most of the bedrock is Precambrian, a southward extension of the Canadian Shield, but it also has broad belts of erosionally non-resistant rocks. The most conspicuous area of weak rocks is the Lake Superior basin, which is localized by the relatively soft red sandstone and shale in the center of the Lake Superior syncline (fig. VII-2A). On the south flank of the syncline are the more resistant copper-bearing conglomerates and basalts from the Keweenaw Peninsula, but south of that is a lowland cut in additional sandstones—the lowland that localized the Chippewa lobe. On the north side of the Lake Superior syncline are the resistant lava flows that form the North Shore Highland, which attains altitudes of 2,300 feet, bordered in turn by the great intrusion of the Duluth Complex, which forms a high plateau that escaped glaciation late in Wisconsin time when the surrounding lowland areas were filled with ice.

The Lake Superior syncline closes to the southwest beyond the head of the lake, and a fault near the axis complicates the patterns, but another lowland picks up *en echelon* to the south. This is localized by the poorly cemented Cambrian sandstones that lap onto the Canadian Shield. The lowland continues southwestward to the Minneapolis area (fig. VII-2A); it was followed first by the Superior glacial lobe, which expanded out of the Lake Superior basin and crossed the low divide to the south, but it was later occupied by ice moving in the opposite direction.

The central part of Minnesota is underlain by a complex of igneous and metamorphic rocks having no pronounced differences in erosional resistance. It therefore had no prominent lowlands or highlands to channel the ice lobes; rather, it has been invaded from east and west at different times as ice expanded out of the bordering basins, and here the drift is the thickest in the state (fig. VII-3).

On the west side of the state is a lowland that was as important in the glacial history as was the Lake Superior lowland on the east (fig. VII-2A). This is the Red River Valley, which is underlain by the soft Cretaceous shales that cover the Paleozoic rocks of the Great Plains. In Minnesota, varied Cretaceous sediments lap onto the Precambrian shield as well. The Red River Valley lowland channeled an ice lobe that continued southeastward down the Minnesota River Valley and thence south across a low divide into central Iowa. The Minnesota River Valley, in turn, is bounded on the southwest by a small ridge of resistant Precambrian rock, the Sioux Quartzite, which may have been high enough to have escaped being covered by the Cretaceous seas.

PRE-WISCONSIN GLACIATION

Pre-Wisconsin drift occurs at or close to the surface near the southeastern and southwestern corners of the state, but it has not been investigated much since it was described by Leverett (1932). In the southeast, the drift was mapped as Iowan, generally covered by loess. Ruhe (1969) studied this drift extensively in adjacent Iowa and judged it to be Kansan, from which the subsequently formed weathered zone was eroded before the deposition of the loess. The basis for the reinterpretation is primarily radiocarbon dates on wood buried in the drift at many different localities; all dates are >35,000 years old (BP). The loess on the top yields dates of 29,000-16,500 BP, and thus is of Wisconsin age. A similar relation probably exists in adjacent Minnesota.

A strip of terrain bordering the Mississippi River valley in southeastern Minnesota has often been designated a portion of the Driftless Area, which occupies primarily the southwestern quarter of Wisconsin. This strip has such a thick cover of loess that few exposures of underlying material occur, and the loess everywhere seems to lie directly on Paleozoic bedrock. Glacial erratics can be found in the beds of streams that head in this area, however, so it is possible that glacial drift occurs locally beneath the loess. In fact, Black (Frye and others, 1965) reported that erratics can be found over a large part of the Driftless Area in Wisconsin. Future detailed work may show that, although the Driftless Area as usually delimited may have been ice-free during the main Wisconsin glaciation, it may have had at least some ice cover during one or more earlier glaciations.

In Dakota County, just south of St. Paul, Leverett identified the Hampton moraine of "Old Red Drift." This he assigned to the Illinoian glaciation, because the drift is overlain by Iowan loess, with no weathering profile between, and is underlain by the "Old Gray Drift" leached of its carbonate (Kansan). Ruhe and Gould (1954) essentially adopted this correlation, although they believed that the Hampton moraine owed its height largely to a core of bedrock. Loess rests on top of the main Wisconsin moraines just to the north, however, so the Hampton moraine can only be designated as older than the loess. With recognition of the early Wisconsin Rockian drift (35,000 BP) in the adjacent part of Wisconsin (Frye and others, 1965), presumably equivalent to the Winnebago (Altonian) drifts of northwestern Illinois (Kempton and Hackett, 1968), we have the possibility that the Hampton moraine represents the Rockian instead of the Illinoian.

Deep exposures of drift elsewhere in Minnesota may penetrate pre-Wisconsin drifts, but there are no reliable ways to correlate them. It will be shown that in many cases

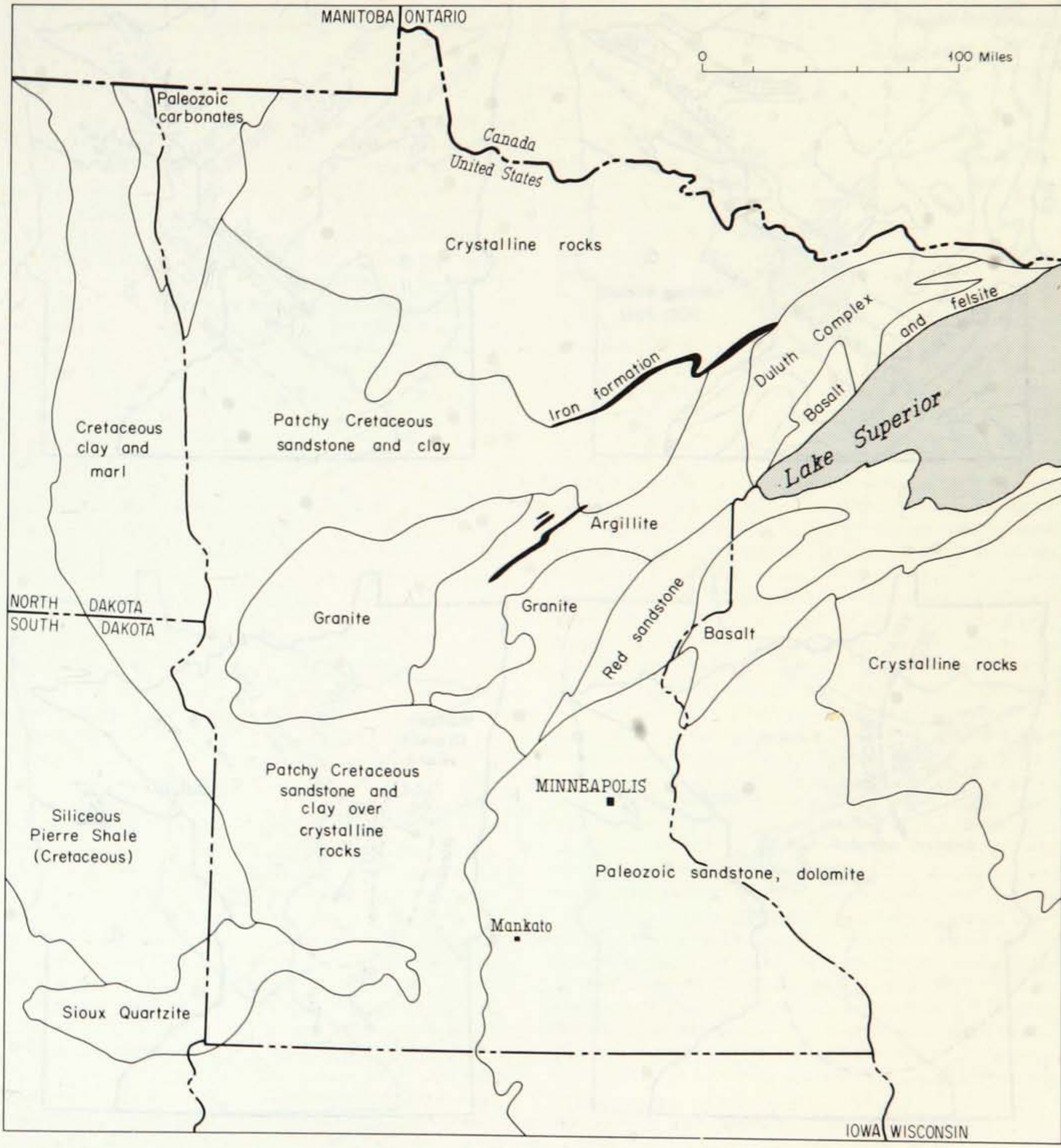


Figure VII-1. Map of bedrock of Minnesota and adjacent areas, to show principal rock types represented in the glacial drifts. The Cretaceous sandstones and clays extending east into northern and southern Minnesota are thin and discontinuous; they differ in lithology from the siliceous marine shales (*e.g.*, Pierre Shale) of the Dakotas. Compiled from state geologic maps and simplified (Wright and others, in press).

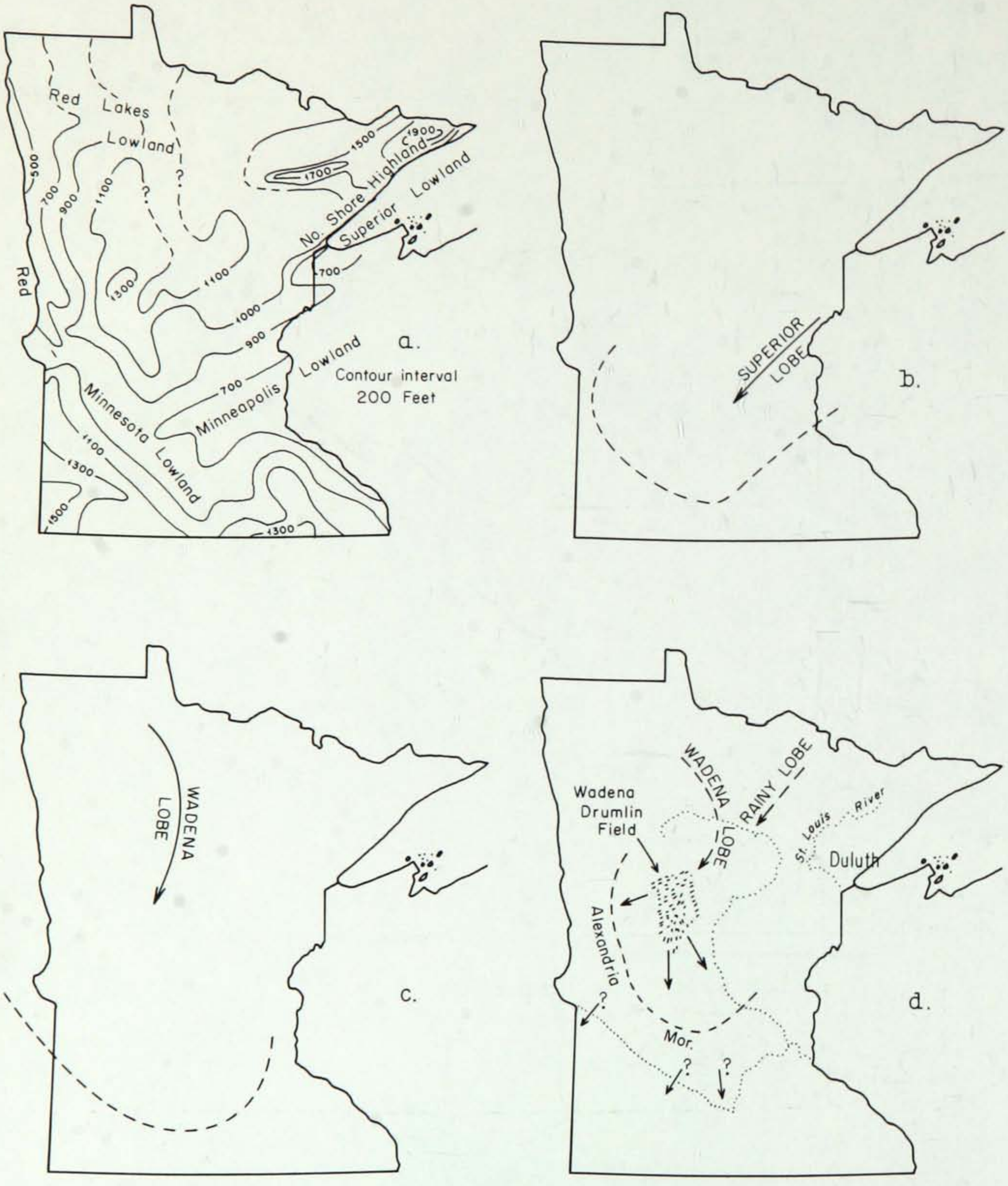


Figure VII-2. Phases of glaciation in Minnesota. Large arrows show direction of ice movement, small arrows show direction of drainage, and groups of dashes show drumlins. See Figures VII-4 and VII-5 for composite maps. A, bedrock topography; B, deposition of Hawk Creek Till by Superior lobe, limits and date uncertain; C, deposition of Granite Falls Till by Wadena lobe, limits and date uncertain; D, Hewitt phase of Wadena lobe. Advance to Alexandria moraine; formation of Wadena drumlins. Eastern limit uncertain. Date early Wisconsin?

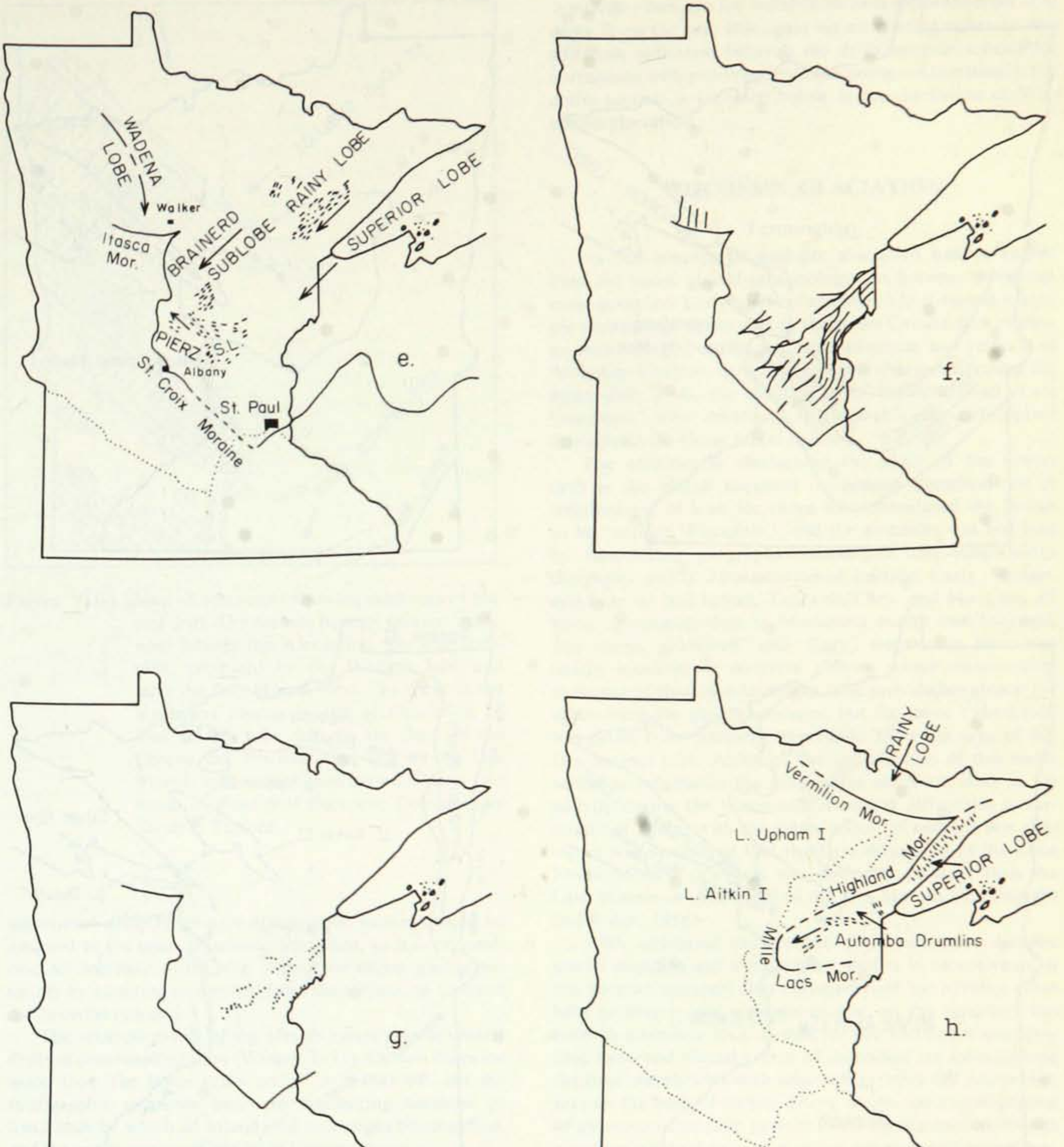


Figure VII-2 (cont'd.). E, St. Croix phase. Advance of Superior and Rainy lobes to St. Croix moraine, and advance of Wadena lobe to Itasca moraine; formation of drumlin fields by sublobes. Date 20,000 BP?; F, erosion of tunnel valleys by subglacial streams beneath Superior and Wadena lobes; G, deposition of eskers in tunnel valleys; H, Automba phase of Superior and Rainy lobes. Advance to Mille Lacs, Highland, and Vermilion moraines; formation of Automba drumlins; formation of proglacial Lakes Aitkin I and Upham I.

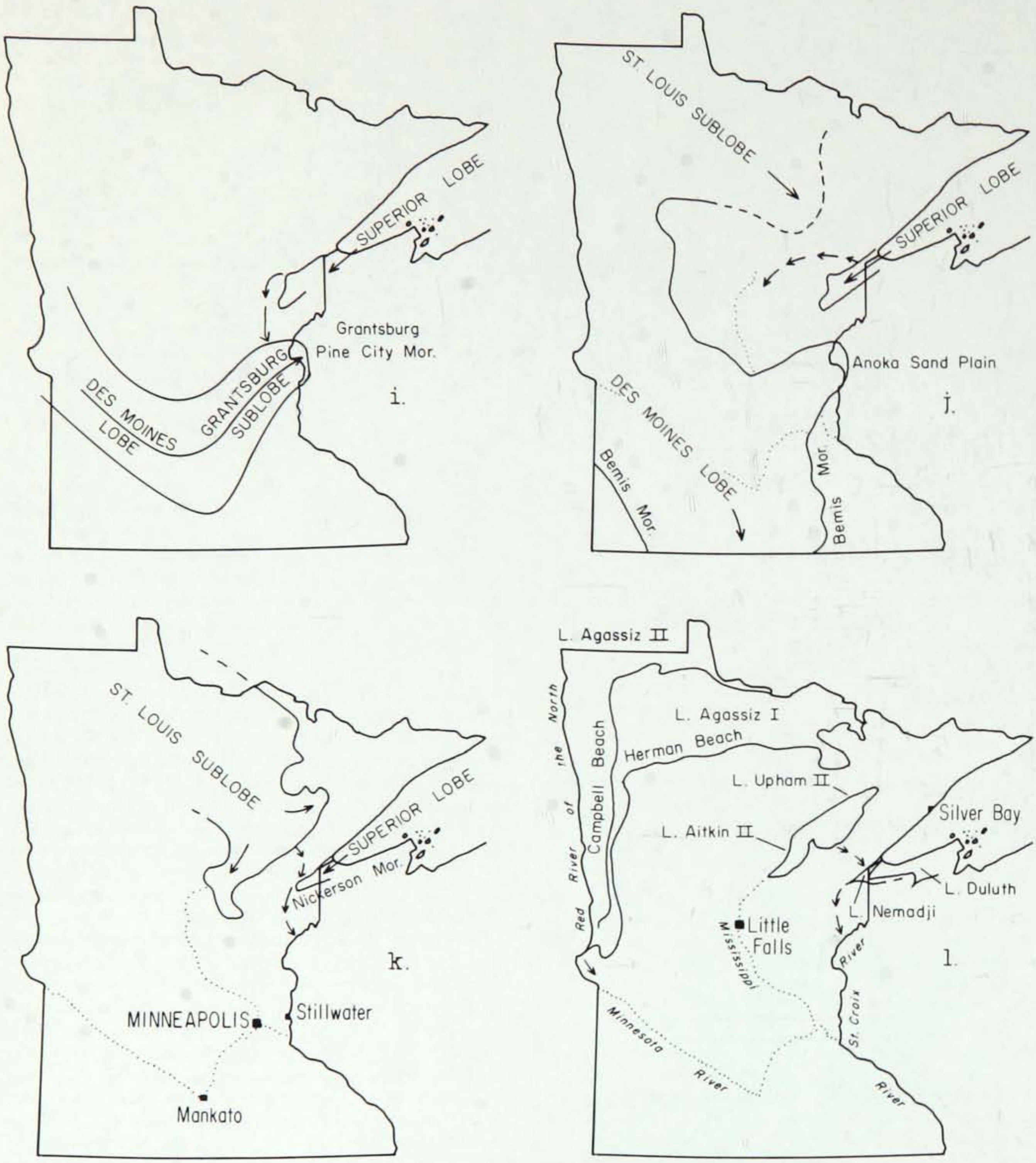


Figure VII-2 (cont'd.). I, Split Rock-Pine City phase of Grantsburg sublobe and Superior lobe. Advance of Grantsburg sublobe to Pine City moraine; formation of proglacial Lake Grantsburg. Date 16,000 BP?; J, Bemis phase of Des Moines lobe. Extension of "Des Moines lobe proper" to Iowa. Formation of Anoka sandplain with wastage of Grantsburg sublobe. Date 14,000 BP?; K, Nickerson-Alborn phase. Advance of Superior lobe to Nickerson and Thomson moraines, and advance of St. Louis sublobe to Alborn moraine. Date 12,000 BP?; L, Agassiz phase. Formation of Glacial Lake Agassiz, with outlet via Glacial River Warren. Formation of Glacial Lakes Aitkin II and Upham II, with outlets via St. Louis River, diverted to St. Croix River.

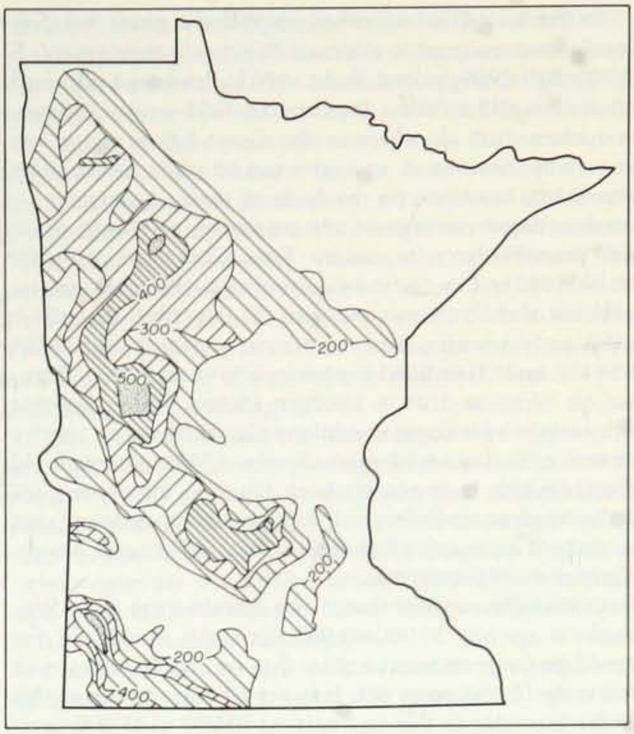


Figure VII-3. Map of Minnesota showing thickness of glacial drift. The arcuate band of greatest thickness follows the Alexandria moraine complex, produced by the Wadena lobe and later the Des Moines lobe. The band in the southwest represents the accumulation of Des Moines lobe drift on the flank of the Coteau des Prairies. The axis of the Des Moines lobe was aligned between these two bands of great drift thickness. Compiled by Sarah P. Tufford.

superposed drifts in a single stratigraphic section can all be assigned to the main Wisconsin glaciation, so it is not realistic to correlate drifts with successive major glaciations simply by counting downward from the surface, as Leverett did in certain cases.

The open-pit mines of the Mesabi range expose several drifts in consistent relation (Winter, 1971). Carbon dates for wood from the lower drifts are all >35,000 BP, but the stratigraphic sequence bears no weathering horizons or fossil beds by which an interglacial unit might be identified, and the entire section might be of Wisconsin age.

In the steep bluffs of the Minnesota River Valley, deep sections of drift are discontinuously exposed for more than 150 miles. Near Mankato, for example, more than 100 feet of drift is exposed, and near Granite Falls and Redwood Falls the drift is seen to rest on Cretaceous sand, clay, and lignite overlying Precambrian crystalline rock deeply weathered to kaolinite (Parham, 1970). Elsewhere in this chapter Matsch describes the sequence of four distinct drifts of three different ice lobes, with carbon dates of 34,000 and >40,000 years ago for wood from beneath the second drift down from the top. But again no weathering zones or fossiliferous sediments between the drifts provide a basis for correlation with pre-Wisconsin glaciations. Accordingly, the entire section is included below in the discussion of Wisconsin glaciation.

WISCONSIN GLACIATION

Terminology

As the concept of multiple glaciation has developed over the years, glacial terminology has become more and more complex. Early workers attempted to maintain a simple terminology applicable to the entire Great Lakes region, to preserve the concept of broad advances and retreats of the entire ice front, driven by climatic changes affecting the entire area. Thus, the terms "Early," "Middle," and "Late Wisconsin" were developed in Leverett's day and applied throughout the Great Lakes region.

The controversy concerning the place of the Iowan drift in the glacial sequence introduced complications in terminology, at least for those who considered the Iowan to be "earliest Wisconsin," and the situation was resolved by introducing geographic names for the subdivisions (Leighton, 1933). Thus, instead of Earliest, Early, Middle, and Late we had Iowan, Tazewell, Cary, and Mankato, all being recognized then in Minnesota except the Tazewell. The names "Tazewell" and "Cary" were taken from two nearby localities in northern Illinois, where cross-cutting moraines of the Lake Michigan lobe provided evidence for subdividing the glacial sequence, but the name "Mankato" was taken from southern Minnesota from the area of the Des Moines lobe. Although the introduction of this name served to emphasize the progressive westward shift in ice activity during the Wisconsin, it caused difficulties in terminology when, with the introduction of radiocarbon dating, it was discovered that the Late Wisconsin of the Lake Michigan lobe (Valders) was distinctly younger than the Late Wisconsin (Mankato) of the Des Moines lobe (Wright and Rubin, 1956).

With additional radiocarbon dates and more detailed glacial mapping and stratigraphic studies in recent years, it has become apparent that correlation of ice advance from lobe to lobe is not a simple matter, so the tendency has been to introduce local names for the successive stratigraphic units and glacial events of individual ice lobes, letting the time correlations with adjacent ice lobes fall where they may on the basis of carbon dating, unless some stratigraphic or geomorphic relation permits closer equivalence. Although a certain regional pattern of major advance and retreat persists for part of the Wisconsin, roughly equivalent to the Tazewell and Cary subdivisions of Leighton (Wright, 1971), the later Wisconsin events continue to cause difficulties in correlation from lobe to lobe.

Accordingly, for Minnesota the sequence of glaciation is recounted as a series of named glacial phases for each of the several ice lobes involved (Wright and Ruhe, 1965), with definite correlation indicated by combining names for adjacent lobes (*e.g.*, Nickerson-Alborn phase of the St.

Louis and Superior lobes). Although the many new names thereby added may give the impression of complexity, the system is intended to permit a more honest representation of what is known, and to reduce the speculation that is inherent in the old system of applying names from other regions on the basis of completely speculative correlations.

Hewitt Phase of the Wadena Lobe

Wadena Drumlin Field

The oldest drift of probable Wisconsin age extensively exposed on the surface in Minnesota is in the Wadena drumlin field in the west-central part of the state, representing the Hewitt phase of the Wadena lobe (fig. VII-2D). About 1,200 drumlins form a fan-shaped pattern in Wadena, Todd, and adjacent parts of Cass, Hubbard, Becker, and Otter Tail Counties (Wright, 1962 and this chapter, fig. VII-37). The pattern indicates that the ice flowed to the southwest, fanning to west and south and terminating at the Alexandria moraine complex. The northeastward plunge of long axes of stones within the till also implies flow from the northeast. The dominance of fragments of Paleozoic carbonate rock in the drift, however, indicates that ice came from the northwest from the Winnipeg lowland in southern Manitoba, the closest area where these rocks crop out. The only other possibility, the Hudson Bay lowland, is unlikely because it is too far away and is separated by many hundred miles of Precambrian crystalline rock that lack a cover of calcareous drift.

This apparent conflict between two different ice-movement criteria is resolved in the following way. The Wadena lobe progressed from the Winnipeg lowland southeastward into the shallow Red Lakes lowland of northern Minnesota. It diverged from the Winnipeg lowland north of the United States border, for the drift contains no fragments of Cretaceous shale, which covers the Paleozoic carbonate rock south of the border. At the same time the Rainy lobe approached the area from the northeast and blocked the Wadena lobe, diverting it from a southeasterly to a southwesterly course. The Wadena lobe then fanned out to form the drumlin field and terminate at the Alexandria moraine complex.

Lithologic evidence for this explanation comes from stone counts of the till (Wright, 1962), which are interpreted as showing how the Wadena lobe, as it was blocked and diverted, incorporated some Rainy lobe ice, and the resulting till contained a mixture of two types of indicator stones. This diluting of the carbonate content of the Wadena lobe till by the addition of eastern components resulted in a weakly calcareous till that was subsequently leached to a greater-than-normal depth. The depth of leaching must be the basis for Leverett's (1932) tentative judgment that the drumlins should be assigned to the Iowan glaciation rather than to the main Wisconsin, even though the usual features diagnostic of Iowan drift—pebble layer, loess cover—are absent. Leverett believed also that the ice moved from southwest to northeast, rather than the reverse, as an expansion of an early Des Moines lobe out of the Red River Valley.

As far as age is concerned, the Hewitt phase has previously been assigned to the main Wisconsin maximum (*ca.* 20,000 BP) (Wright and Ruhe, 1965), because undrained depressions still exist on the drumlin field—early Wisconsin surface drift elsewhere in the Great Lakes region has no such depressions. A case now can be made for an older correlation, however, on the basis of three arguments: (1) a carbon date from organic silts on top of the Hewitt phase drift near Pillsbury, in eastern Todd County, is >40,000 BP (W-1232); (2) a carbon date on basal lake sediments on drumlins of the St. Croix phase of the Superior lobe, which is distinctly younger than the Hewitt phase, is 20,500 BP (I-5443); and (3) unfilled depressions have now been identified on Illinoian drift in southern Illinois, indicating that with certain hydrologic conditions glacial lakes can survive for tens of thousands of years (Jacobs, 1970). Although old lake sediments have not yet been found in the depressions of the Wadena drumlin field, it seems better now to favor an early-Wisconsin rather than a main-Wisconsin assignment for the Hewitt phase.

One might suppose that if the drumlins are early Wisconsin in age (say 30,000-60,000 years old), the weathering should be more extensive than that on nearby calcareous drift only 12,000 years old. It is quite likely, however, that tundra prevailed in this region from 20,000 to 11,000 years ago, according to paleobotanical studies in central Minnesota, so the climate was probably too frigid for much chemical weathering and soil formation during this time, and frost action and solifluction may have disturbed the incipient soils that may have previously formed. The climate for the preceding segments of Wisconsin time is not known for this region but it was probably almost as cold, as suggested by the persistence of stagnant ice in the Alexandria moraine throughout this time (see below). Real forest-soil formation in this area probably did not commence until the postglacial climatic regime developed about 11,000 years ago.

Alexandria Moraine Complex

The Alexandria moraine complex is considered here to represent in its core the terminal deposit of the Wadena lobe when the drumlins were formed, because it so clearly rims the outer margin of the drumlin field. The moraine was entirely overridden at a later date by the Des Moines lobe from the west, however, and much of the core is obscured by this later cover (figs. VII-4, 5). The cover extended eastward over the outer part of the drumlin field from Becker County to southern Todd County, as can be seen from numerous exposures where the two drifts can be distinguished. Within the moraine complex, the surface drift is generally that of the Des Moines lobe, even in deep exposures, so the burial must have been extensive. The total drift thickness here is more than 500 feet (fig. VII-3).

Wadena Lobe Drift in the Minnesota River Valley

Recent mapping southwest of the Alexandria moraine complex by C. L. Matsch, described elsewhere in this chapter, suggests that the Wadena lobe, with its shale-free or shale-poor drift, extended beyond this moraine, across the Minnesota River Valley, and even to South Dakota (figs.

VII-2C, 5). Numerous exposures in the bluffs of the Minnesota River Valley and its tributaries show that the surficial drift (the shale-bearing Des Moines lobe till) is underlain by the shale-poor Wadena lobe drift, called the Granite Falls Till, with a boulder pavement common at the contact but without weathering horizon (except oxidation). The contact, which has a very gentle slope, is considered by Matsch to represent a former land surface subject to slope wash on a relatively arid landscape, causing the removal of fine particles and the concentration of boulders and cobbles on the surface (Wright and others, in press). When the Des Moines lobe later overrode this terrain, the boulders must have been frozen in the ground, so that they were faceted and striated without being turned.

This western extension of the Wadena lobe may represent a phase of Wisconsin glaciation older than the Hewitt phase, or it may be simply an early maximum of the Hewitt phase, before retreat of the ice to the Alexandria moraine complex, where it must have remained for a very long time. Carbon dates on wood from beneath this drift near Redwood Falls in the Minnesota River Valley are 34,000 and >40,000 BP (Matsch, this chapter).

If the Wadena lobe at this time extended south and west beyond the Alexandria moraine, it is not easy to understand the total course of the ice lobe and its relation to the Red River Valley, which today is the deepest lowland of western Minnesota. That is, it is not clear why the Wadena lobe, protruding from the ice sheet in the Winnipeg lowland, did not simply continue south along the Red River lowland, as the Des Moines lobe later did, rather than diverge to the southeast into the Red Lakes lowland, only to be diverted

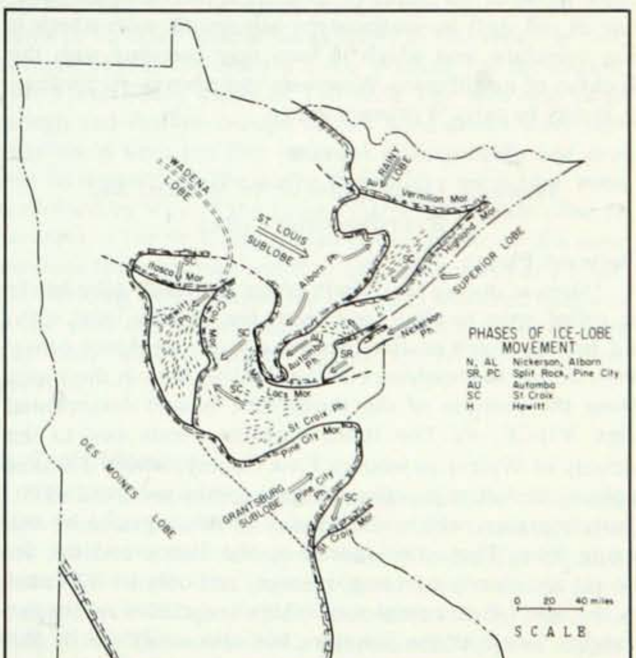


Figure VII-4. Composite map showing main phases of Wisconsin glaciation in Minnesota. Short dashes show drumlin fields.

southwest by the Rainy lobe. Perhaps the Red River lowland as such was not in existence at this time. One could postulate that the Red Lakes lowland was in fact the lowest terrain at that time, and that the Cretaceous rocks of North

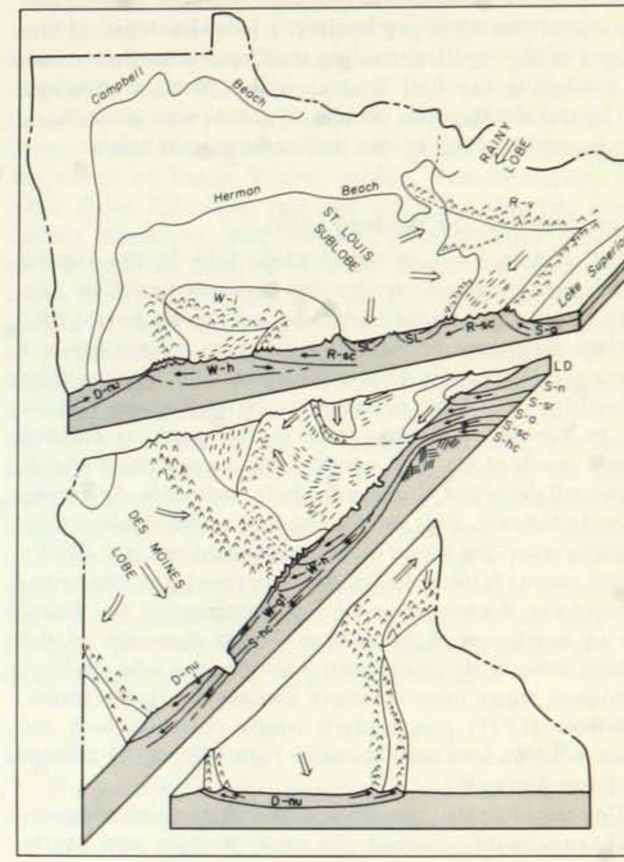


Figure VII-5. Block diagram of Minnesota showing composite of main phases of glaciation, with three cross-sections. Arrows show direction of ice movement. Main moraines and drumlin fields are also shown. Initials stand for the following ice lobes and phases:

Des Moines lobe
- D-nu New Ulm phase

Wadena lobe
- W-i Itasca phase
- W-h Hewitt phase
- W-gf Granite Falls phase

Rainy lobe
- R-v Vermilion phase
- R-sc St. Croix phase

Superior lobe
- LD Lake Duluth
- S-n Nickerson phase
- S-sr Split Rock phase
- S-a Automba phase
- S-sc St. Croix phase
- S-hc Hawk Creek phase

Dakota extended farther east over the site of the Red River Valley. The construction of the massive Alexandria moraine by the diverted Wadena lobe may have produced a topographic barrier large enough (especially if it contained great quantities of stagnant ice) so that with the next major push of ice out of the Winnipeg lowland a lobe developed at the west end of this barrier, eroding the Cretaceous shales and thus producing the Red River lowland. Such a situation helps to explain the rich content of Cretaceous shale fragments in the Des Moines lobe drift, as discussed below.

Eastern Limit of the Wadena Lobe

The eastern limit of the Wadena lobe in the Hewitt phase is poorly defined, for its drift is deeply buried by that of the Rainy and Superior lobes, so extensively buried that the time difference between the advances of western and eastern ice lobes must have been great. The Wadena lobe at this time may have extended as far east as the Milaca area in central Minnesota, and as far south as Dakota County, south of St. Paul, where there are exposures of calcareous till devoid of Cretaceous shale fragments. At a more northerly latitude, drift resembling that of the Wadena lobe is common as the basal till in many iron-ore pits of the Mesabi range (Winter, 1971). On the other hand, the orientation of the Wadena drumlin field implies that the Rainy lobe in north-central Minnesota caused diversion of the Wadena lobe to the southwest. The Wadena lobe drift on the Mesabi range must therefore predate the Hewitt phase, as Winter (1971) proposed. It might correlate with the buried Wadena lobe drift (Granite Falls Till) of the Minnesota River Valley.

Compared to the massive Alexandria moraine complex, which presumably formed the main western and southwestern margin of the Wadena lobe in the Hewitt phase, the eastern margin produced no moraine at all, or at least none that survived later ice advance from the east. This condition is perhaps further evidence that the Wadena lobe was confluent on its east flank with the Rainy lobe, which at this time was advancing from the northeast alongside the Superior lobe. With such lateral confluence, ice flow is directed longitudinally and therefore leaves no moraine; the situation is like that of two confluent tributaries of a valley glacier, except that no medial moraine is formed, because there is no interfluve to provide the necessary rock debris. When the Wadena lobe withdrew, as recounted below, the Rainy-Superior lobe moved forward to its western maximum at the St. Croix moraine.

Retreat of the Wadena Lobe

Few details of ice retreat of the Wadena lobe are recorded. Within the Alexandria moraine complex, the records of ice retreat are thoroughly obscured by later drifts, although the abundance of ice-contact gravels, kettle lakes, and related features in the moraine implies the long persistence of stagnant ice. All younger features within the Wadena drumlin field, such as outwash plains, are clearly related to later glaciations rather than to Hewitt retreatal features. The Wadena lobe must have persisted for a very long time in a steady state to mold the very well developed drumlin field by subglacial ice flow and especially to build the core of the massive Alexandria moraine complex, which has local relief of 200 feet and drift more than 500 feet thick (probably including some pre-Wisconsin drift, of course). When climatic change upset the steady state, it almost seems as if the ice lobe must have disappeared by evaporation or at least that its front retreated so rapidly that no meltwater features such as eskers or proglacial outwash plains were formed to disrupt the well-formed drumlin field.

Regardless of the duration and mechanics of wastage of the Wadena lobe at the end of the Hewitt phase, a clear readvance of the ice in the following glacial phase is recorded by the Itasca moraine, which trends east-west across Hubbard County, truncating clearly the southwestward-trending drumlins previously formed. The drift of the Itasca moraine is essentially identical to that of the Wadena drumlin field —gray to light-brown sandy calcareous till without fragments of Cretaceous shale—and it is easy to visualize the Wadena lobe readvancing to the position marked by the Itasca moraine after a recession and realignment.

Still Older Drifts in the Minnesota River Valley

The Wadena lobe till beneath the boulder pavement of the Minnesota River Valley bluffs is underlain in turn at a few localities by red sandy till with numerous indicator stones from the Lake Superior area (figs. VII-2B, 5). This is the Hawk Creek Till (Matsch, this chapter). No weathering horizon caps this drift, so its assignment to a pre-Wisconsin glaciation is not justified. It is clearly older than the St. Croix phase of the Superior lobe, as is the Hampton moraine of red drift in southeastern Minnesota, with which it may correlate, and which in turn may correlate with the Rockian of neighboring Wisconsin (see above). Accordingly, it may be early Wisconsin in age.

Itasca-St. Croix Phase of the Wadena and Rainy/Superior Lobes

Outwash Plains

Whereas during the Hewitt phase the Rainy lobe has to be called upon to block and divert the Wadena lobe, without leaving much positive evidence for its existence otherwise, during the readvance of the Wadena lobe in the Itasca phase the position of the Rainy lobe is well documented (figs. VII-2E, 4). The Itasca moraine trends east to the vicinity of Walker in western Cass County, where it makes a clear interlobate junction with the southward-trending St. Croix moraine, which was formed in this segment by the Rainy lobe. These two moraines, the Itasca and the St. Croix, are clearly contemporaneous, not only as indicated by the interlobate complex of eskers and related meltwater features found at the junction, but also as shown by the confluent outwash plain that forms an extensive blanket on the Wadena drumlins in front of the two moraines. The Park Rapids outwash plain south of the Itasca moraine is the more extensive of the two, for it forms an almost uninterrupted plain extending with decreasing slope and decreasing particle size south for about 10 miles to the vicinity of Menahga. It does contain a few ice-block depressions, however, such as the numbered Crow Wing lakes, which, significantly, must mark the persistence of buried ice blocks within the Wadena drumlin field—another indication that the Hewitt phase is not separated from the main-Wisconsin glaciation by a climatic interval that was temperate enough to melt the buried deadice.

Outwash from the St. Croix moraine is not so extensive as that from the Itasca moraine, but several definite outwash fans with radial drainage can be identified (Wright, 1962; Schneider, 1961). The sediments of these fans, as well as those of the Park Rapids outwash plain, become thinner outward, so that the drumlins, which are completely buried close to the moraines, emerge to the southwest from beneath the cover. Actually, although the drumlin forms emerge, the till itself is buried by as much as 15 feet of sand, which thus forms a true blanket, in the sense that it covers both drumlins and swales evenly but does not obscure the pattern. The oval hills are still visible on aerial photographs and on soil maps, which show the pattern of sandy soils on the hills and of marsh peats in the swales. This type of non-obscuring blanket extends southwest from the outwash plains about to the Redeye River, beyond which the drumlins are not covered with such sand, and the till is exposed at the surface.

The genesis of the sand blanket is not exactly clear. If outwash streams extended outward from the Park Rapids and other outwash plains and fans, one would expect that the sands would fill in the swales between the drumlins, whereas the sand was deposited on the drumlins themselves as well as in between. The sands contain scattered cobbles and boulders but are not stratified. The best explanation seems to be that a glacial lake existed in this area—in fact, Leverett (1932) mentions a Lake Wadena for the region, with a maximum depth of 130 feet. The lake was large enough and shallow enough so that wind-driven wave currents could keep the fine sediment in suspension and prevent its deposition. The outlet of the lake must have been southward by way of the Long Prairie River, thus close to the front of the St. Croix moraine. The course of this river has since been reversed and the outlet obliterated as a result of overriding by a younger ice lobe. The scattered boulders in the sand can be attributed to berg-rafting, for the lake may have abutted locally against ice of the St. Croix moraine.

St. Croix Moraine

The St. Croix moraine is traceable south for about 100 miles from its interlobate junction with the Itasca moraine near Walker to the vicinity of Albany in Stearns County, west of St. Cloud. This segment is continuous except for a large gap near Pillager, which was eroded later by the Crow Wing River. The moraine is a steep-fronted ridge of rugged topography, bordered on the east throughout its length by the upper Mississippi River valley. Southeast of Albany it is buried by younger drifts in a segment about 100 miles long, beyond which it emerges again as a broad, rugged ridge in the St. Paul area (fig. VII-5). Here, at the point of the Superior lobe, the moraine turns northeast, crosses the St. Croix River south of Stillwater, and continues as a belt of rugged hills that provide the terrain for most of the major ski resorts in northwestern Wisconsin.

The St. Croix moraine, thus traced for 350 miles through Minnesota and Wisconsin, is one of the most sharply defined glacial features in the Great Lakes region, despite its local burial or erosion. Its western segment as far south as the Albany area is composed primarily of brown sandy till and associated ice-contact gravels containing stones of basalt, felsite, gabbro, iron-formation, and other types indicating that the ice came through northeastern Minnesota. Red sandstone, the principal indicator of the Superior lobe, is virtually absent. Its eastern segment, however, from the St. Paul area northeast into Wisconsin, contains red sandy drift having abundant fragments of Precambrian red sandstone and shale diagnostic of the Superior lobe. The St. Croix moraine, although traceable as a single geomorphic feature, was thus formed in part by the Rainy lobe and in part by the Superior lobe. The latter climbed out of the Lake Superior basin across red sandstone bedrock and then moved down the Minneapolis lowland. The Rainy lobe flowed across the upland of crystalline rocks north of Lake Superior. The two lobes, both moving southwest, became laterally confluent in central Minnesota beyond the limits of the Lake Superior lowland and the highland, and they flowed side by side as a single lobe to their common terminus, the St. Croix moraine.

Drumlin Fields

The ice-flow pattern of part of this double lobe can be inferred from the southwesterly trend of a group of about 100 drumlins south of Brainerd. An outlying cluster of this type occurs near Pine River, northwest of Brainerd, consisting of about a dozen drumlins. The drumlins in the Brainerd field are made of brown sandy till identical to that in the St. Croix moraine. It is not clear why the trend of the drumlins is oblique rather than perpendicular to the moraine in this region.

In sharp contact with the Brainerd drumlin field on the south, across the narrow Skunk River valley, is the Pierz drumlin field (fig. VII-6). The Pierz drumlins, which number about 1,600, form a fan-shaped field covering most of Benton County and parts of adjacent counties. An outlier of the field occurs west of the Mississippi River valley close to the St. Croix moraine (Schneider, 1961). The margin of the fan essentially follows the gentle curvature of the St. Croix moraine in the segment from Pillager gap (west of Brainerd) to the Albany area.

The apparently simple relation of the Pierz drumlin field to the St. Croix moraine is complicated by the fact that although the northern part of the field consists of the brown sandy till typical of the Rainy lobe, the more southerly drumlins are made of the red sandy till typical of the Superior lobe. The explanation for this complication is as follows. The Rainy and Superior lobes, which were discrete lines of flow in northeastern Minnesota, being separated in part by the North Shore Highland, became laterally confluent in central Minnesota, where the dividing highland

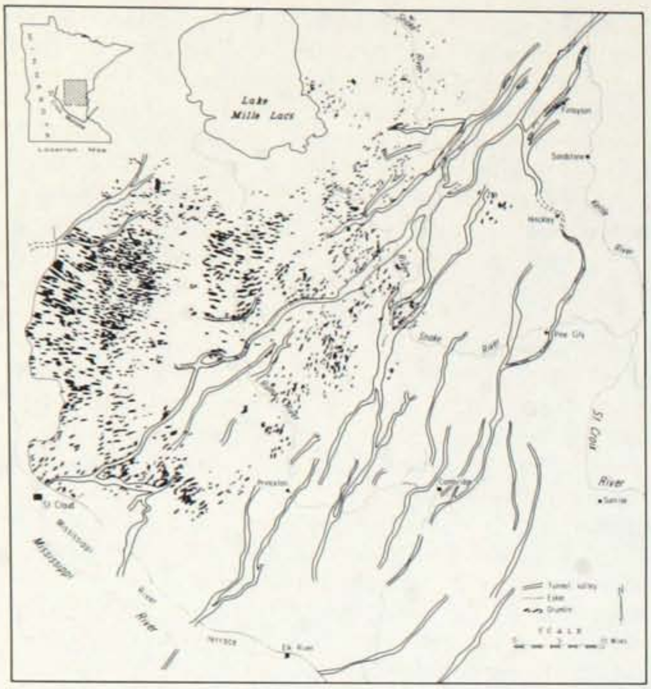


Figure VII-6. Map of part of east-central Minnesota showing drumlins, tunnel valleys, and eskers of the Superior-Rainy lobes, formed during the St. Croix phase of Wisconsin glaciation. Some of the features south of a line from St. Cloud to Princeton to Pine City are obscured by a cover of younger drift. The fan-shaped pattern, interrupted by the Mississippi River, terminates near the St. Croix moraine, the position of which is shown on the inset map.

ceased to exist. Lateral mixing between the two lobes resulted in local stratigraphic superposition of red till upon brown, or vice versa. The Pierz sublobe, which formed the drumlin field, thus contained contributions of drift-laden ice from both of the original lobes. Any drumlins formed by the Superior lobe are largely obscured by a cover of younger drift. The Brainerd sublobe represented a separate line of flow of the Rainy lobe.

In addition to the Brainerd and Pierz drumlin fields in central Minnesota, the Rainy lobe proper produced the Toimi drumlin field in the northeast, between the North Shore Highland and the Mesabi range. Here a series of about 1,400 drumlins trends directly S. 45° W., with no fanning or divergence. The drumlins are just as large or larger than the Wadena drumlins, averaging a mile long and 50 feet high (Wright, this chapter, fig. VII-37). The drumlin field is truncated on the north by the Vermilion moraine, which represents a later advance of the Rainy lobe, on the east and south by the Highland-Mille Lacs moraine of the Superior lobe, and on the west by the Culver moraine of the St. Louis sublobe.

The Toimi drumlins consist of gray, sandy, stony till with so little clay and silt that the material is almost as loose as outwash. The most conspicuous rock type is gabbro, which forms the bedrock in most of the area. The drumlins are buried by younger drift south of the St. Louis River, but where the drift emerges again south of Mille Lacs Lake (in the Pierz drumlin field) and west of Mille Lacs Lake (in the Brainerd drumlin field), it is changed to brown sandy till dominated by fragments of slate, graywacke, and other metamorphic and igneous rocks, which constitute the bedrock over which the ice passed south of the St. Louis River.

An additional patch of Rainy lobe drumlins is found near Hibbing and Eveleth on the Mesabi range. The bouldery drift of which they are composed, as seen in the deep iron-ore pits of the region, occurs between the basal Wadena lobe drift and the surficial red clayey till of the St. Louis sublobe (Winter, 1971).

Proglacial Drainage Channels

As the combined Rainy-Superior lobe retreated from the western segment of the St. Croix moraine in western Morrison County, it formed a series of closely spaced recessional moraines. Proglacial lakes formed at the ice front between successive recessional moraines, and their outlet streams ended four large parallel drainageways leading south into Stearns County (Schneider, 1961). The southern ends of these drainage channels subsequently have been obscured by a younger drift cover, but they probably led ultimately across the entire St. Croix moraine south to the Minnesota River.

Tunnel Valleys

Meanwhile, other features developed under the Superior lobe proper. Meltwater at the bottom of the ice flowed to the terminus, cutting a series of gorges or tunnel valleys that are as much as half a mile wide, 200 feet deep, and 100 miles long (Wright, in press). They all trend generally southwest, parallel to the trend of the broad Minneapolis lowland, which accommodated the ice lobe (figs. VII-2F, 6). This trend reproduces the direction of slope of the ice surfaces and thus of the hydrostatic gradient. The trend is oblique to the regional slope of the terrain, which is southward on the flank of the Minneapolis lowland, as indicated by the course of the modern Rum, Snake, and Knife Rivers. It is also oblique to the trends of the drumlins, which represent patterns of local subglacial ice flow (Wright, this chapter, fig. VII-35).

The tunnel valleys extended to and probably locally through the St. Croix moraine, and their streams emerged at the surface at the front of the ice, to spread great outwash fans consisting of detritus eroded from the gorges. Such a fan, beyond the tip of the lobe, extends from Rosemount almost to Hastings along the front of the St. Croix moraine. The outwash was graded to the Mississippi River, in fact it formed the source of the river at this time. The valley train, as it heads in the moraine, has an altitude of about 900 feet above sea level, and the graded depositional surface can be traced for many miles downstream as the highest terrace in the valley.

Between St. Paul and Albany, the frontal outwash plains and the tunnel valleys that transect the moraine are all deeply buried by younger drift. In fact, the transection of the moraine by tunnel valleys in this segment may have resulted in the creation of enough gaps in the moraine so that the later ice from the west, the Grantsburg sublobe, could break through and flow into the area vacated by the Superior lobe.

Altogether about 12 subparallel tunnel valleys can be traced across east-central Minnesota, despite the fact that the lower courses of several have been buried and partially obscured by younger deposits associated with the Grantsburg sublobe. The longest tunnel valley starts near Moose Lake, just north of the divide between the Minneapolis lowland and the Lake Superior basin. Near Finlayson it is broad and shallow and contains several eskers, and the entire area is partially buried by younger drift. South of the divide it becomes a single valley, half a mile wide, now occupied by the Grindstone River. Grindstone Lake, which fills part of the valley floor, has a surface about 50 feet below the level of the till plain. The lake has a maximum depth of 125 feet. Although the lake sediment has not been cored, most lakes of this type have at least 30 feet of postglacial sediment. Thus, the floor of the tunnel valley was cut by the subglacial stream at least 200 feet into the till plain. Well borings in the region indicate that this valley is cut through the drift and into bedrock.

The Grindstone tunnel valley turns east to Hinckley and then south, where it is partially obscured by a younger outwash valley train and delta leading into Glacial Lake Grantsburg. The valley then includes Cross Lake, whose projecting Norway Point at the northern end is a segment of the esker that generally identifies a tunnel valley. The valley in this area has a blanket of locally varved clay deposited by Glacial Lake Grantsburg.

Southwest of Pine City, the same tunnel valley and its esker pick up a mantle of till as they pass under a moraine of the Grantsburg sublobe, and beyond this they are largely obscured by sands of the Anoka sandplain. However, this same tunnel valley system can be traced by a string of lakes across the sandplain to the St. Croix moraine north of St. Paul. This particular tunnel valley, especially in the segment near Grindstone Lake, may have been eroded by some pre-Wisconsin outlet stream from a proglacial lake in the Lake Superior basin, but its esker indicates that the valley was occupied in late Wisconsin time as well. Most of the other tunnel valleys have no such possible connection to proglacial lakes. Because their courses follow obliquely along the flank of the Minneapolis lowland rather than directly down the slope, the streams must have been localized by some factor other than ground slope, namely the slope of the superjacent ice, *i.e.*, the gradient of the hydrostatic (or cryostatic) pressure.

The mechanics of formation of the tunnel valleys depend on an abundant supply of water (Wright, in press). Tunnel valleys generally are assumed to derive their water from surface melting, but consideration of the thermal regime of the ice lobe makes this source unlikely. Specifically, paleobotanical evidence for Minnesota indicates that atmospheric temperatures were sufficiently low to inhibit the growth of trees in the northeastern part of Minnesota at least until about 11,000 years ago, and the long persistence of stagnant ice in the St. Croix moraine further implies subfreezing mean annual temperatures of both ground and air. The glacier surface must also have been cold, especially at the higher elevations back from the front. For example, if the surface gradient of the Superior lobe resembled that of west-central Greenland in the outer 100 miles of the ice sheet, the altitude of the ice surface over the tunnel valley area must have been at least 6,000 feet. If the air temperature at the ice front was $-2°$ C, then at 6,000 feet it must have been at least $-15°$ C, as in Greenland. Such cold surface temperatures would prevent the deep penetration of surface meltwaters. The tunnel valley water must therefore have come from another source, namely from basal melting.

Basal melting of ice results in part from the escape of crustal heat, which supplies about 0.5 cm per year, and in part from the frictional heat of basal sliding, which can supply several centimeters. But neither source would supply sufficient water each year to erode the tunnel valleys. Accordingly, the hypothesis has been developed that the basal meltwater produced under the thick ice lobe, especially in the Lake Superior basin, was trapped behind the cold toe of the ice lobe, which was frozen to its substratum. The basal meltwater might then build up over thousands of years until the volume became much larger. Furthermore, this process of basal melting must have prevailed throughout the ice sheet, so basal meltwater might have been drawn to the Superior lobe from as far away as Hudson Bay. Eventually the stored basal water worked its way through the frozen toe of the ice lobe, aided by fracture lines resulting from ice flow. Once a series of channels was formed, the stored water escaped rapidly and eroded the tunnel valleys.

Eskers

As the Superior lobe thinned, hydrostatic pressure on the subglacial streams decreased, and the water velocity was no longer sufficient to keep open the ice tunnels against the pressure of ice flow and was too low to permit further erosion of the tunnel valleys (fig. VII-2G). In fact, the velocity was inadequate to transport the sediment supplied to the stream by melting ice, and the tunnels, much reduced in size by pressure of the ice walls, became filled with sediment. These partial fillings of the diminished tunnels became the eskers that mark the bottoms or flanks of tunnel valleys. The ice that constricted the tunnels in these late stages left its record in the form of lakes strung along the valleys today; each lake marks a mass of ice that had crowded into the original tunnel valley and protected it from deposition by the esker stream.

Most of the tunnel valleys and associated eskers occur in east-central Minnesota, in the area affected principally by the Superior lobe. The area of the Rainy component has been largely buried by younger drifts, except north of the St. Louis River, where Rainy lobe drift is exposed on the surface in the form of the Toimi drumlin field. In this area, the drumlin plain is dissected by about four subparallel valleys that may be tunnel valleys cut in the same way as those farther south. These subparallel valleys start gradually and deepen to the southwest over a distance of about 30 miles as they approach the area of the St. Louis River, which was non-existent at the time of their formation. They cannot be traced readily south of the St. Louis River because of burial by younger drift. The Cloquet and Artichoke Rivers follow these valleys in some segments.

Concurrent wastage of the Wadena lobe from the Itasca moraine also resulted in a set of tunnel valleys. These are most conspicuous in Itasca State Park and adjacent areas. Lake Itasca itself, as well as the long strings of lakes and esker-like ridges that continue the southerly trends of the east and west arms of Lake Itasca through the Itasca moraine, are in such tunnel valleys. The many-armed Mantrap Lake follows the linear pattern within the Itasca moraine itself. The debris eroded from the tunnel valleys was spread out in the great Park Rapids outwash plain south of the moraine, as previously described. To the north, the Mississippi River flows northward from Lake Itasca in first one and then another tunnel valley. The system is largely buried by younger drift 15 miles north of the park.

Ice Retreat at the End of the Itasca-St. Croix Phase

So extensive are the indications of ice wastage for the Itasca-St. Croix phase that one must postulate a relatively long interval unfavorable to glaciation. The ice lobes thinned over a broad zone and left behind countless blocks of stagnant ice, many of which survived the retreatal interval and were subsequently buried during younger ice advances, indicating that the climate during the interval was still cold enough to inhibit the thawing of ground ice. The distance of retreat of the ice fronts can be measured within modest limits. The Superior lobe retreated from the St. Croix moraine to a point just north of the divide between the Minneapolis lowland and the Lake Superior basin, a distance of about 120 miles. If it had retreated much farther north, its meltwater would have been ponded between the ice front and the divide, and the subsequent ice advance would have produced a till consisting in part of reworked lake beds, as was the case during later phases of the Superior lobe.

In the case of the Rainy lobe, the front retreated from the St. Croix moraine near Albany for at least 200 miles before its readvance to the Vermilion moraine, which truncates the Toimi drumlin field on the north. In its retreat north of the Giants Range, which is the narrow upland flanked by the Mesabi range, a small proglacial lake was formed whose outlet cut the sharp Embarrass channel across the range (Winter, 1971).

The distance of retreat of the Wadena lobe from the Itasca moraine cannot be easily assessed, but it seems likely that the ice wasted completely back to the Winnipeg area, about 200 miles, because when it readvanced, in the form of the St. Louis sublobe of the Des Moines lobe, it had an entirely different alignment.

Radiocarbon Dating

The radiocarbon age of the Itasca-St. Croix phase cannot be determined accurately. Only minimal dates, from the basal sediments of lakes located on the relevant drifts, are available. On the St. Croix moraine itself, the oldest basal date is 13,270±200 (Y-1326), at Kirchner Marsh, south of St. Paul, and this is demonstrably an ice-block depression (Florin and Wright, 1969). On lake sediments in an interdrumlin depression at Wolf Creek in the Pierz drumlin field, the oldest date is 20,500±400 (I-5443), and basal dates for similar deposits in the Toimi drumlin field are 14,690±390 (W-1763) at Weber Lake and 15,850±240 (I-5048) at Kylen Lake. Basal dates for lakes on drift clearly younger than that of the St. Croix phase extend back to 16,150±550 (W-1973). The Itasca-St. Croix phase is therefore likely to have reached its maximum at least 20,500 years ago, and is correlative with the maximum (Tazewell) extent of the Lake Michigan lobe, rather than with the Cary phase, as was concluded by Leverett and subsequent writers.

Automba Phase of the Superior Lobe

Extent of the Lobe

Following the general deglaciation of the St. Croix phase, the Superior and Rainy lobes readvanced to new positions. The locations of the Wadena and Des Moines lobes are unknown for this time. The Superior lobe advanced out of the head of the Lake Superior basin, but instead of moving southwest across the low divide near Sandstone and thence down into the Minneapolis lowland, as its predecessor had done in the St. Croix phase, it extended west to the region of Mille Lacs Lake in east-central Minnesota (fig. VII-2H). The reason for this course is not clear, for the bedrock in this direction (1,300 feet above sea level) is higher than the Sandstone divide (less than 1,200 feet). Possibly, the Sandstone area held a large amount of stagnant ice, surviving from the St. Croix phase. This ice may have provided a topographic barrier sufficiently high to block the readvancing Superior lobe, and it may have had a component of drift sufficiently large to inhibit remobilization. Although there is evidence for the persistence of stagnant ice in this region at the end of the St. Croix phase, the drift component of the ice was not large enough to leave a moraine, or at least not a moraine that survived overriding by the readvancing ice.

Moraines

A problem related to the course of the Superior lobe in the Automba phase is the contrasting morainic forms on opposite sides of the ice lobe. At the end of the lobe, the very distinct Mille Lacs moraine was formed, at the west end of Mille Lacs Lake (fig. VII-5). The moraine continues around the south side of the lake but fades out eastward, and the left (southern) flank of the lobe is not easily traced from there east to Wisconsin. Leverett (1932), who considered that the Mille Lacs moraine represented a simple retreatal position of the ice as it wasted from the St. Croix moraine, mapped the Kerrick moraine in the general area of the Sandstone divide, but this feature, unlike the Mille Lacs moraine, is not clearly traceable as a belt of rough topography; in fact, its height may simply be an expression of the bedrock divide itself rather than of till deposition.

On the north side of the Superior lobe in the Automba phase, on the other hand, the moraine is fairly continuous. The individual segments are not the same as those mapped by Leverett, whose reconstruction of the Superior lobe for this time was different from that presented here. The Mille Lacs moraine can be traced around to the north side of Mille Lacs Lake, and thence east-northeast as the Wright and Cromwell moraines to the Highland moraine, which follows the crest of the North Shore Highland for 100 miles as a very distinct belt of hummocky topography 5-10 miles broad The Highland moraine truncates the southeastern edge of the Toimi drumlin field of the St. Croix phase, and its outwash follows the troughs between drumlins along what is now the Cloquet and Whiteface River valley. The meltwater must have led into proglacial lakes Upham I and Aitkin I, because the expanded Superior lobe at this time dammed the normal southward drainage of this area into either the Lake Superior basin or the Mississippi River. The outlet of the two lakes presumably was around the west end of the lobe, in a course obliterated by subsequent ice movements.

Drumlins and Flutes

The trend of the Mille Lacs-Wright-Cromwell-Highland moraine along the northwestern flank of the Superior lobe of the Automba phase is matched by a series of drumlins and related ice-movement features, which show the direction of ice flow when the ice margin was stable long enough to construct this prominent moraine. The Automba drumlin field consists of three partially connected areas of drumlins. The largest area is in the Cromwell quadrangle in Carlton County, between Cromwell and Automba. It consists of about 200 drumlins which trend generally northwest, averaging 25 feet in height and less than a mile in length. The field terminates obliquely at the Cromwell and Wright moraines, and is buried on its eastern edge by younger drift. Near Automba in southwestern Carlton County, the drumlins fan from west-northwest to west; in adjacent Aitkin County the drumlin pattern turns southwest to Mille Lacs Lake, where the drumlins form peninsulas projecting into the lake.

Northeast of the Cromwell quadrangle, the drumlins are largely buried by outwash of younger drifts, but they partially emerge from beneath the cover west of Cloquet, and north of Cloquet and the St. Louis River an additional 125 drumlins are visible, trending north-northeast toward the Highland moraine.

Northwest of Duluth, where the Highland moraine turns more to the northeast to follow the crest of the North Shore Highland, the Automba drumlins lose their identity in the numerous hills of bedrock, but farther up the shore a new pattern dominates on the bedrock slopes leading to the Highland moraine. This pattern is conspicuous on both topographic maps and aerial photographs (Wright, this chapter, fig. VII-32). It consists of linear ridges and scarps in bedrock, as well as linear accumulations of drift. The pattern thus does not consist entirely of drumlins; it is referred to as the Highland flutes (Wright and Watts, 1969). It may be traced along the north shore slope into Cook County, where it was recognized by Sharp (Grout and others, 1959) as "drumloid topography" but attributed to the Rainy lobe moving into the Lake Superior basin rather than to the Superior lobe moving out to the northwest.

The combining of the Automba drumlin field and the Highland flutes with the Mille Lacs-Wright-Cromwell-Highland moraine provides a good record of flow of the right-hand (northwest) half of the Superior lobe during the Automba phase. Because of the depth and sharpness of the Lake Superior basin, the Superior lobe in this phase was like a relatively narrow outlet glacier rather than a broad bulge on the ice sheet. Nonetheless, in the area of the Highland moraine, which marked the side of the ice lobe and thus resembled a lateral moraine, the ice flow was directly normal to the ice margin. Farther toward the terminus of the lobe, however, in the area of the Cromwell and Wright segments of the moraine, the iceflow direction was oblique to the lateral ice margin. At the end of the lobe, near Mille Lacs Lake, the drumlins once again show the perpendicular direction of ice flow.

Ice Thickness

It is assumed that all drumlins, flutes, and moraines were formed at the same time, when the position of the ice margin was stable for a time sufficiently long to permit these major morphologic features to form, presumably as a result of stable climatic conditions. The gradient of the ice surface at this time was at least as steep as the gradient of the Highland moraine, which represents the lateral ice margin. This moraine has an altitude of about 1,900 feet above sea level near its northeastern end near Isabella, north of Silver Bay in western Cook County. From there it descends uniformly at a gradient of about 75 feet per mile to an altitude of about 1,500 feet west of Duluth, at the point where the ice tongue began to expand west out of the basin.

The thickness of the ice can be determined for this segment, although the figure must be a minimum because the cross profile of the surface is not known. Presumably the center of the lobe had a higher altitude than the lateral margin against the Highland moraine, so that ice flow was directed toward the lateral margin to form the Highland flutes. The present floor of Lake Superior off Silver Bay has an altitude of about 375 feet below sea level and has the form of a long, narrow trough. Sub-bottom geophysical profiles imply that sediment in this trough is at least 1,000 feet thick (Farrand, 1969). A 682-foot core through this sediment shows that lake sediments, probably of late-glacial age, extend to a depth of 200 feet, below which are clayey tills and additional layers of lake sediments, the ages of which, though unknown, possibly postdate the Automba phase of ice advance under consideration. The ice was therefore at least 1,900+375+200 (2,475) feet thick and perhaps more than 3,000 feet thick, at the point about 170 miles from its terminus at the Mille Lacs moraine.

From the Silver Bay area the ice thinned southwestward, as the surface descended and the floor of the basin rose. In the Duluth area the ice was probably about 1,000 feet thick, and as it crossed the divide out of the Lake Superior basin and fanned westward to the Mille Lacs area, a distance of about 50 miles, it was probably only a few hundred feet thick.

Composition of the Drift

The drift of the Superior lobe in the Automba phase, red sandy till characterized by stones of the Lake Superior basin, especially red sandstone and shale, is generally similar to that of the St. Croix phase. This description applies principally to drift south and southwest of Duluth. The Highland moraine and associated glacial features in the North Shore Highland lack the red material, so the color of the Superior lobe drift in this region is brown rather than red, except where it is enriched by fragments of the red syenite ("redrock") that is a local facies of the Duluth Complex.

Development of the Ice Lobe

The lack of distinction between the Superior lobe drifts of the St. Croix and Automba phases means that the features of these two ice advances must be distinguished primarily on geomorphic grounds. The strongest reason for making this distinction is the Automba drumlin field, which indicates that the Superior lobe had an alignment definitely at variance with that during the St. Croix phase. Westward movement of the ice lobe out of the Lake Superior basin to the Mille Lacs Lake area was not possible during the St. Croix phase, because the Rainy lobe occupied that region, as indicated by the alignment of the Brainerd drumlin field. But during the Automba phase the Rainy lobe had withdrawn far northeast to the Vermilion moraine, as recounted below, so that the Superior lobe had access to central Minnesota. The reason why the lobe at this time headed for the Mille Lacs region rather than southwest to the Minneapolis lowland has already been discussed.

A somewhat different way of looking at the Automba phase, a view first developed during my discussions in the field with E. J. Cushing in 1970, may be mentioned as an alternative explanation for the poor development of a moraine on the south side. During extensive wastage of the combined Superior-Rainy lobe at the close of the St. Croix phase, the Superior lobe portion retreated less rapidly, because the ice was much thicker over the Lake Superior basin than over the North Shore Highland and supplied more ice. As a consequence, the Rainy lobe sector retreated far north to the Vermilion moraine, creating space for the still-vigorous Superior lobe to expand as a sublobe westward to part of that area vacated by the Rainy lobe, *i.e.*, as the advance referred to as the Automba phase. In this interpretation, the Superior lobe proper continued to flow into the Minneapolis lowland and thus formed no moraine.

One stratigraphic relation may involve both the St. Croix and Automba phases of the Superior lobe. This is shown in a series of exposures near Finlayson in Pine County. Here is found the headward part of the long Grindstone tunnel valley, whose course was previously traced in the discussion of wastage features associated with the St. Croix phase. Near Finlayson this tunnel valley consists of a broad but very shallow strip cut below the general till surface and containing several discontinuous eskers. The entire area, including the eskers, is marked by caps of red clayey till that are assigned to a readvance of the Superior lobe at a still later date, after the Automba phase. But locally under the clay-till cap is a layer of red sandy till that may be assigned to the Automba phase. The best exposure is on the Indian Lake esker, in a roadcut about half a mile southwest of Finlayson, where 2 feet of red clayey till overlies 8 feet of red sandy till, which in turn overlies red sandy gravel that constitutes the core of the esker. Two alternative explanations can be given for this stratigraphic relation. The first is that one or both tills represent simply the ablation till left behind when the roof of the ice tunnel in which the esker was formed melted, leaving behind its included debris. The difficulty with this explanation is that in Minnesota no eskers have ever been found in which the till cap cannot more easily be explained by overriding by a readvancing ice lobe. This very same Grindstone tunnel valley, for example, has eskers with no caps at all from the Finlayson area south to Hinckley, where the eskers are overlain by younger outwash sands of the Hinckley fan. Farther onward in the Pine City area the eskers are overlain by the clays of Glacial Lake Grantsburg and then, south of Pine City, by till of the Grantsburg sublobe. Furthermore, the till on the Finlayson esker has a definite preferential orientation to the long axes of its linear stones, a feature indicating deposition of the till during active southwestward flow of ice rather than superposition of particles during the downwastage of stagnant ice.

The second explanation for the double till cap on the Finlayson eskers is that the later ice movement was associated not with the Automba phase of the Superior lobe but with the Split Rock phase that followed, and that the till deposited by this ice advance was locally sandy or clayey, depending on whether or not the local lake sediments that were being overridden by the readvancing ice were sandy or clayey. In this explanation, the overriding ice might thus pick up some clayey sediments and some sandy sediments and redeposit them in turn. Elsewhere there are certain indications, from composition of the drift of the Split Rock phase and especially of the still later Nickerson phase, that the texture of the drift varied locally, as explained below, but for the case in point the stone orientations in the two till caps are distinctly different: the stones of the lower, sandy till show a southwestern preference, whereas those of the clayey till show a western preference. Thus, two different ice movements are implied, consistent with the hypothesis that the esker, trending southwest as a result ultimately of the southwest trend of the Superior lobe during the St. Croix phase, was overridden first by ice moving west in the Automba phase after its realignment, and overridden later by ice moving southwest in the Split Rock phase, as described below. This double overriding of eskers, without appreciable destruction of the esker form, is further manifestation of the protection to eskers provided by the deadice persisting in the adjacent parts of the tunnel valley, or perhaps is a manifestation of the weakness of glacial ice in its erosive power under certain circumstances. In any case, the Finlayson eskers are as typical in form and basic core composition as any other eskers, despite their caps of till.

Relation to the Rainy Lobe

The Highland moraine of the Automba phase of the Superior lobe can be traced northeast along the crest of the

North Shore Highland about as far as the Isabella area, and throughout this stretch it truncates, slightly obliquely, the Toimi drumlin field. Near Isabella it is joined from the west by the Vermilion moraine, which truncates the Toimi drumlin field on the north and extends west as a sharp morainic ridge across the eastern end of the Mesabi range, near Babbitt, and thence straight west to bound Lake Vermilion and Nett Lake on the south. The Vermilion moraine is one of the most distinct moraines in Minnesota, as far as topographic representation is concerned. In the Isabella area the Vermilion and Highland moraines meet at about a 30° angle. In the interlobate junction is a series of eskers, implying the effluence of important interlobate subglacial streams. Northeast from the interlobate confluence, the two moraines lose their identity. Apparently, where two adjacent ice lobes come together obliquely, the accumulation of debris is negligible, in comparison with the situation of two tributaries of a valley glacier, in which the rock ridge between provides the debris for a medial moraine. This means essentially that when two ice lobes are flowing side by side, there is a negligible lateral ice flow, but as soon as the lobes have a free margin the ice flows directly to that margin. The junction of the Highland and Vermilion moraines is exactly comparable to the junction of the St. Croix and Itasca moraines, described previously.

Split Rock-Pine City Phase of the Superior and Grantsburg Lobes

Superior Lobe

Little record remains of the wastage of the Superior lobe in the Automba phase. The next event in the history of the area is recorded by the red clayey till that forms a discontinuous blanket over the southwestern end of the Lake Superior basin. This drift reaches an altitude of about 1,250 feet above sea level, extending southwest almost to the drainage divide between the Lake Superior basin and the Minneapolis lowland (fig. VII-21). It is expressed as a cover generally only a few feet thick over red sandy till or outwash of previous phases of Superior lobe glaciation. It covers eskers, tunnel valleys, and till uplands, and has local variations in thickness from zero to about 20 feet. It is largely confined to the relatively narrow lowland at the head of the Lake Superior basin, especially where this lowland, underlain by eastward-dipping red sandstone and shale, is bounded on the northwest by an escarpment leading up to the older crystalline rocks at Denham. But near the north end of this escarpment near Denham, the ice apparently advanced westward as a small protuberance off the main lobe, leading to the Split Rock River valley, and in this area the local westward expansion of the ice was sufficiently well defined to produce a field of about 50 small drumlins, composed of the typical red clayey drift of the Split Rock advance. These drumlins, which trend westward just west of the village of Moose Lake, are much smaller than the northwestward-trending Automba drumlins to the north. They represent a minor western lateral bulge in the main trend of the Superior lobe at this time, which otherwise headed southwest down the Lake Superior basin to the rock divide west of Sandstone. No terminal moraine is apparent, but small proglacial lakes may have formed at the front of the ice tongue at this time. These lakes were drained by spillways that led from the Finlayson area east to tributaries of the St. Croix River.

The till of the Split Rock phase is recognized generally by its clayey texture. Presumably the Superior lobe at the end of the Automba phase had retreated far enough into the Lake Superior basin for sizeable proglacial lakes to form at its margin, and in these were deposited clayey and silty sediments, with sand and stones added from wasting icebergs. This means that the ice must have withdrawn sufficiently far into the Lake Superior basin to create one or more lakes that were deep enough to permit the accumulation of clayey sediments.

Meltwater from the protuberance up the Split Rock River valley was directed westward into the Snake River. Farther northeast, along the lateral margin of the ice lobe, the meltwater produced definite outwash fans, graded westward ultimately to the headwaters of the Snake River by way of extensive flatlands among the hills of the Automba phase drift to the west. Farther northeast the lateral margin of the ice at the Split Rock phase is marked not by the limit of red clayey drift but rather by a distinctive ice-contact slope fronted by coalescing outwash fans and backed by five distinct eskers. The relations are diagrammatic: subglacial tunnels emerging at the ice front deposited eskers in their downstream ends and spewed gravel and sand as aprons in front of the ice. The eskers, clearly displayed on the Cloquet and Barnum topographic maps, distinctly lead northwest to outwash fans, which coalesce to a plain at an altitude of about 1,280 feet—a plain that grades westward and contains linear features that record the westward radiating pattern of individual fans. The eskers are short, only 1-2 miles, partly because their headward ends may have been truncated by the erosional channels of the diverted St. Louis River of a later date, but mostly because the eskers may represent the downstream ends of major subglacial tunnel valley streams emerging from under the thick Superior lobe. Because of the abrupt shallowing of the Lake Superior basin in this region — the floor rose at a gradient of about 30 feet per mile from the axis of the lobe in the Duluth area to its lateral margin — the subglacial streams lost hydrostatic pressure as they approached the thinner ice margin, and therefore deposited their loads of gravel and sand in their lower reaches, and in the proglacial fans, as eskers.

The plain to which the major eskers are graded — the Cloquet outwash plain — was actually preceded by a slightly higher and thus earlier outwash plain farther northwest. This feature, the Sawyer outwash plain, has an altitude of about 1,300 feet above sea level, and is separated from the Cloquet outwash plain by an ice-contact slope about 20 feet high. One of the eskers of the Cloquet series, the Bob Lake esker, fed the Sawyer outwash plain first. After forming the plain, the meltwater drained in part northeastward via the low areas within the drift plain and moraine of the Automba and St. Croix phases that preceded, eventually reaching the area of the St. Louis River. This river had not yet come into existence as a master stream, however, because the Lake Superior basin was still plugged with ice. The meltwater must therefore have headed westward to the basin of

Glacial Lake Upham, which must have had its outlet via Glacial Lake Aitkin to the Mississippi River.

The more southerly part of the Sawyer outwash plain drained directly west to the headwaters of the Snake River via what are now broad, flat, swampy areas at the head of the Snake River. Most of the sediment settled out in these flat areas, and the relatively clean water that drained southward cut the sharp Snake River gorge.

Meltwater from the Superior lobe in the Split Rock phase not only headed west from the northwestern side of the lobe but also moved southwest, directly down the lowland into the Grindstone River drainage, following one of the old St. Croix phase tunnel valley systems for about 15 miles. The fact that Grindstone Lake, more than 100 feet deep, is located athwart this course indicates that St. Croix phase stagnant ice still existed at the time of Split Rock phase drainage. The meltwater followed the tunnel valley's winding course to Hinckley, where it deposited a broad outwash fan that practically buried the tunnel valley eskers and continued east as a delta into Glacial Lake Grantsburg. Actually, the easterly course of the meltwater in the vicinity of Hinckley was controlled in part by the persistence of St. Croix phase stagnant ice blocks to the south — now represented by till-covered areas lower than the Hinckley outwash plain and lower than the upper limit of the Glacial Lake Grantsburg deposits.

Grantsburg Sublobe

Glacial Lake Grantsburg was formed north of the Grantsburg sublobe of the Des Moines lobe as a result of damming of the Mississippi River and other drainage from the north. Its limits as mapped by Cooper (1935) extend east from St. Cloud across the St. Croix River and well into Wisconsin. Most of the drainage into the Minnesota portion of the lake was from the Grantsburg sublobe itself, rather than from meltwater of the contemporaneous (Split Rock phase) Superior lobe to the north, because the sediments of this part of Lake Grantsburg are primarily gray rather than red. The lake sediments rarely exceed a few feet in thickness; locally they are rhythmically laminated (varved). Apparently the lake here was short-lived, perhaps existing only for a century or so. But in the Wisconsin portion of Glacial Lake Grantsburg, especially in the St. Croix River valley, the lake clays are much thicker. Extrapolation of the varve counts through the non-varved section implies that the lake lasted for almost 2,000 years (Wright and others, in press).

The meltwater relations recounted above demonstrate the approximate contemporaneity of the Split Rock phase of the Superior lobe and the maximum extent (Pine City phase) of the Grantsburg sublobe. This sublobe extended northeast along the axis of the Minneapolis lowland, where the transecting crests of both the Alexandria moraine complex and the St. Croix moraine may have been relatively low. The breakthrough may have been localized further by the presence of several tunnel valleys cut through the St. Croix moraine in this region, as previously postulated, to a depth sufficient to provide an entry to ice from the west. As the sublobe moved toward its terminus near Grantsburg, Wisconsin, it was delimited on its southern (right) flank largely by the St. Croix moraine, which at this time probably contained a large amount of stagnant ice and may have been an even more prominent ridge than it is today. At no point from western Minneapolis to Grantsburg, a distance of 75 miles, did the Grantsburg sublobe top the St. Croix moraine, although its outwash locally leaked through the moraine to the Mississippi and St. Croix River drainages.

The drift of the Grantsburg sublobe is the typical calcareous gray silty till of the Des Moines lobe, light brown where oxidized. It is generally rich in small fragments of Cretaceous shale from the Red River Valley, as well as in Paleozoic carbonate from Manitoba. But one of the most distinctive features of the drift is its lower part, which rests on the red drift of the Superior lobe. This portion commonly consists not of the typical gray silty till but of interlaminated red sandy till and gray silty till, in layers as thin as an eighth of an inch, extending through a zone several feet in thickness. Each lamination is remarkably "pure" in color, texture, and lithology. The individual laminations can be traced laterally for several hundred feet in some exposures. Ordinarily they consist of till, but, in some exposures, they consist of sand. Commonly the unequidimensional pebbles in the layers show an obviously preferred orientation of the linear pebbles in the laminations, in the gray till above, and in the red till below.

The gross interlaminations implied to Upham (1900), who first described them in the Minneapolis area, that eastern and western ice lobes were contemporaneous in the area, with alternating deposition by each. Leverett (1932) made no mention of the relations, although his glacial history implied a discrete separation of the Superior lobe glaciation and the Grantsburg sublobe glaciation. I (1953) attempted to elaborate the Upham explanation, but subsequent work, originated by E. J. Cushing (unpub. manuscript), has demonstrated that the more likely explanation attributes the interlamination to a structural rather than a sedimentary process. As the Grantsburg sublobe advanced from the west across a low area of the St. Croix moraine, which still included large masses of drift-filled stagnant ice, it picked up blocks of red drift (or drift-filled ice), stretched them out into layers, and deposited them concordantly with layers of gray till derived from up-glacier. The exact mechanism of erosion of the older drift and of deposition of the interlaminated complex is not easily postulated, but the parallel stone fabric is critical to the interpretation. The sparse sand laminae show no primary depositional structures; rather they are massive, with sand-grain orientation parallel to pebble orientation in adjacent till laminae. Apparently, masses of sand, as well as masses of till, were picked up by the overriding ice and were both strung out in the same manner, with the internal structure being produced by flow processes. Where till rests directly on outwash, stringers of the latter may be traced into the former, indicating in part how the overriding ice picked up sand and incorporated it into itself.

The parallel stone orientation in the underlying red till implies that a certain amount of this till also was picked up and redeposited by the overriding Grantsburg sublobe. Insufficient fabric analyses have been made of thick till sections, however, to indicate precisely how much of the underlying red till was eroded and then redeposited with a different fabric. The persistence of stagnant ice in the moraine may account for the common occurrence of rebedded structures, for a deadice moraine might not only provide a substratum with many irregularities readily subject to erosion by the overriding ice, but would also provide till masses that already had an easily deformable matrix of ice.

Actually, such foliate structure probably is common wherever one drift is overridden by a later glacier. The contrasts in color and lithology make the case at hand conspicuous. Similar structures have been found at the contact of other dissimilar drifts in Minnesota, *e.g.*, the Wadena lobe drift under the Rainy lobe drift of the St. Croix moraine (Schneider, 1961).

Glacial Lake Grantsburg

The Grantsburg sublobe constructed the Pine City moraine along part of its terminus in Pine County, and it locally built other short moraines. In general, however, the deposited till was not uniformly thick and in some cases the underlying red drift is scarcely covered. The ice blocked drainage from the north, thereby building Glacial Lake Grantsburg, whose importance in correlation has already been mentioned. Lake Grantsburg's limits were mapped by Cooper (1935) from patches of lake clay, some of which are so thin that they are mostly incorporated into the soil. The clay is not continuous, primarily because masses of stagnant Superior lobe ice, surviving from the St. Croix phase, formed islands in the lake — islands now represented by low areas in which red till is exposed at the surface. The lake probably did not survive very long, because the clay is generally so thin. It is locally marked by cyclical laminations that are presumably varves. The lake did not last long enough at a constant level to permit the development of shorelines anywhere. Nor is there a clear spillway channel eroded around the east end of the sublobe in Wisconsin. Cooper postulated that the lake had its outlet over the terminus of the ice itself.

Most of the Lake Grantsburg clay is gray, indicating a source in the Grantsburg sublobe itself. If an important amount of meltwater had been coursing down from the north from the Superior lobe at this time, much more of the sediment would have been red. This relation is a factor critical to the conclusion that the Superior lobe of the St. Croix phase and also of the Automba phase had withdrawn north across the Sandstone drainage divide. During the Split Rock phase of the Superior lobe, meltwater drainage apparently came down the Mississippi, Snake and Grindstone Rivers to Lake Grantsburg, but most of the sediment had been screened out by lakes or flat depositional plains.

Where Lake Grantsburg crosses the St. Croix River valley, however, the clays are dominantly red. Here the sediment was supplied primarily by the Knife River and the upper St. Croix River, both of which headed directly in the Superior lobe rather than in proglacial lakes that trapped the red sediment.

Anoka Sandplain

As the Grantsburg sublobe withdrew from its short-lived maximum, Lake Grantsburg drained, and the meltwater formed a series of coalescing outwash plains wherever the wasting ice exposed low ground. In this way the vast Anoka sandplain was constructed. Its principal outlet was northeast to the St. Croix River by way of the Snake River near Braham and Pine City. The meltwater was derived in large part from wasting ice, for close to the islands of till the outwash sand contains lenses of gravel (Farnham, 1956), but the plain also received water from the diverted Mississippi River. Small lake plains were formed in addition to outwash plains, especially in the southern part of the sandplain, where water flow was still restricted by the persisting St. Croix moraine and its glacial reinforcements supplied by the onlapping Grantsburg sublobe. Such small plains just north of Minneapolis have been described (Stone, 1966); the outlets of the lakes were either north to the sandplain proper, and thus east to the St. Croix River, or south through the moraine to the Mississippi River.

Upham (Winchell and Upham, 1888, p. 415) recognized the Anoka sandplain as a glaciofluvial feature formed during wastage of the Minnesota lobe (Grantsburg sublobe). Sardeson, however, whose work was incorporated in the monograph of Leverett (1932), considered that the sandplain was formed by wind. Cooper (1935), in an extensive study of the entire region, demonstrated that sand dunes were confined to very local areas, and that the plain was basically a glaciofluvial feature. Subsequent work (Farnham, 1956) has confirmed this interpretation, although it is now known that much of the southern part of the plain is actually lacustrine (Stone, 1966).

The Anoka sandplain, although flat over broad areas, locally has relief that can be attributed to several different features. Positive features are either areas of sand dunes or incompletely buried tills of glacial drift left by either the Grantsburg sublobe or the Superior lobe, or by both. Some of the areas of drift were named moraines by Leverett (1932) and Stone (1966). One of the most interesting of these is just north of Elk River, where an esker complex related to the St. Croix phase of the Superior lobe is overlain by till of the Grantsburg sublobe, and this by outwash that forms a pitted plain, bordered by ice-contact slopes on three sides, which slopes northward to grade into the Anoka sandplain (Wright, 1956).

Negative relief features in the Anoka sandplain are mostly ice-block depressions, resulting from buried ice of either the Grantsburg sublobe or the Superior lobe. The depressions, occurring along distinct lines, mark tunnel valleys that were formed during wastage of the Superior lobe of the St. Croix phase.

As the Grantsburg sublobe wasted, the course of the outwash streams — both the Mississippi River and local streams — continued to shift generally southward, as lower ground was gradually opened by the wasting ice (Cooper, 1935). Some of these courses are revealed by the pattern of channel bars left by braided streams (Cushing, 1963, unpub. Ph.D. thesis, Univ. Minn.). The gradual southwestward slope of the plain, determined by Cooper (1935) before the availability of topographic maps, reflects the regional slope of this part of Minnesota, on the flank of the Minneapolis lowland. Eventually, the master stream broke through the St. Croix moraine at Minneapolis, and the straightening of the course of this stream, from St. Cloud to Minneapolis, resulted in a distinct entrenchment of the Anoka sandplain. The wide valley thereby produced was called by Cooper (1935) "the Mississippi valley train." Its sandy sediment is slightly coarser than the sand of the Anoka sandplain proper, presumably because of the more distinctly channeled course of the river and thus a higher velocity. As a result, the valley train area was marked by prairie vegetation and soil, which extended into the deciduous forest of central Minnesota as a long finger. Cooper believed that the sand dunes on the Anoka sandplain started to form when the water table lowered as a result of this dissection. Present-day topographic maps, however, indicate that the draw-down effect on the water table on the sandplain close to areas of relatively shallow stream dissection probably is not enough to control wind deflation. The features might better be attributed simply to strong winds during a time of dry climate, combined with an abundant source of sand. The dune-forming winds were southwesterly, according to the orientation of the dunes. Today the strong summer winds, although from the southwest, are rarely strong enough to move sand even when the vegetation is sparse enough to permit deflation. The dune fields do not directly border the valley train, so it can be assumed that the source of the sand was the sandplain itself rather than the valley train.

When the Mississippi River broke through the St. Croix moraine at Minneapolis, it established its modern course, which ultimately became a gorge. Earlier gorges of a similar type crisscross the Minneapolis-St. Paul area, as seen on maps showing the contours of the bedrock surface (figs. VII-7 and 8). One former course of the river leads south from Anoka through western Minneapolis, where it is marked by a prominent string of lakes (Calhoun, Harriet, and others). Another course extends southwest from Anoka to St. Paul, and is also marked by a string of lakes (Johanna, McCarron, and others). These courses were filled with as much as 350 feet of till and outwash during the Wisconsin or perhaps earlier glaciations.

Mississippi River

The course of the Mississippi River from Anoka to St. Paul at the end of the Pine City phase was controlled only by low spots in the St. Croix moraine, which may still have been filled with stagnant ice. At St. Paul, it intercepted a former valley that apparently was still distinct enough to receive the new river. The river thus extended its valley train, which is preserved as terrace remnants sloping downstream from 810 feet above sea level at Fort Snelling.

These terrace remnants are preserved best at the mouths of tributaries, where they were out of range of subsequent deep dissection along the Mississippi River. The tributary deposits generally are fine sands and silts, locally with the thin laminations that characterize lake deposits. Apparently, outwash sedimentation was so rapid along the glacial Mississippi River at this stage that backwaters extended up the tributaries, many of which originated in the periglacial terrain east of the Des Moines lobe rather than on the ice itself. These tributaries apparently could not deposit enough sediment to keep pace with Mississippi River outwash deposition, so small lakes formed in their lower reaches.

Meanwhile, the Grantsburg sublobe and its parent Des Moines lobe continued to retreat, until it uncovered the Red River lowland, in which Lake Agassiz was formed as a proglacial lake. The history of this lake is recounted below, but here something may be mentioned about the effect of its outlet stream (Glacial River Warren) in the Minneapolis area.

The Mississippi River and its tributaries deposited valley trains in the area during the time of ice retreat, as previously mentioned, and, as the source of the meltwater and the sediment retreated to the west, the early valley train was dissected in stages, leaving terraces. These are best seen in the valley below St. Paul, where at Grey Cloud Island, for example, terraces at altitudes of 750 feet and lower represent intervals of downcutting from the main surface at 810 feet. In this segment, the river simply re-excavated an older valley that previously had been cut into the bedrock, and the dissection and terrace formation was probably rapid. But upstream from St. Paul, the new course of the river was superimposed through the till of the St. Croix moraine onto the flatlying Platteville Limestone (Ordovician), and dissection was inhibited. In fact much of the terrace between Fort Snelling and St. Paul is actually a stripped surface on the Platteville Limestone, with a veneer of gravel.

Retreat of St. Anthony Falls

As the new river course entered the west side of the old valley at St. Paul, a waterfall was formed. By this time Lake Agassiz had come into existence, and the Minnesota River, which joins the Mississippi at Fort Snelling, was vastly enlarged by the outlet waters (fig. VII-9). This Glacial River Warren, freed of its sediment load by the settling basin of Lake Agassiz, could flow on a gradient lower than that of the outwash stream that preceded it. With so much more volume, the river had much more erosive power. Below the waterfall, the river rapidly cleaned out much of the filling of drift and terrace sediments, and it cut into bedrock to an altitude as much as 175 feet below the modern floodplain (Zumberge, 1952). The waterfall was thus heightened, and it retreated upstream as the soft St. Peter Sandstone under the Platteville Limestone was eroded at the plunge pool at the base of the falls. Huge slabs of limestone, broken from the falls but not subsequently removed, can still be seen on the sides of the gorge formed by the retreat.

The River Warren falls retreated 8 miles upstream from St. Paul to Fort Snelling, where the River Warren was joined by the Mississippi, then a much smaller stream with no major source in a glacial lake. The waterfall split into two parts. The River Warren falls continued to retreat upstream an additional 2 miles, at which point another buried gorge of the Mississippi was intersected. This falls thus became extinguished, because on the west side of the buried valley the caprock of Platteville Limestone was no longer present. The other branch began its retreat up the Mississippi, finally reaching its present position today as St. Anthony Falls, about 8 miles above Fort Snelling.

The retreat of St. Anthony Falls was plotted by Winchell (in Winchell and Upham, 1888, p. 313-341) in a resource-

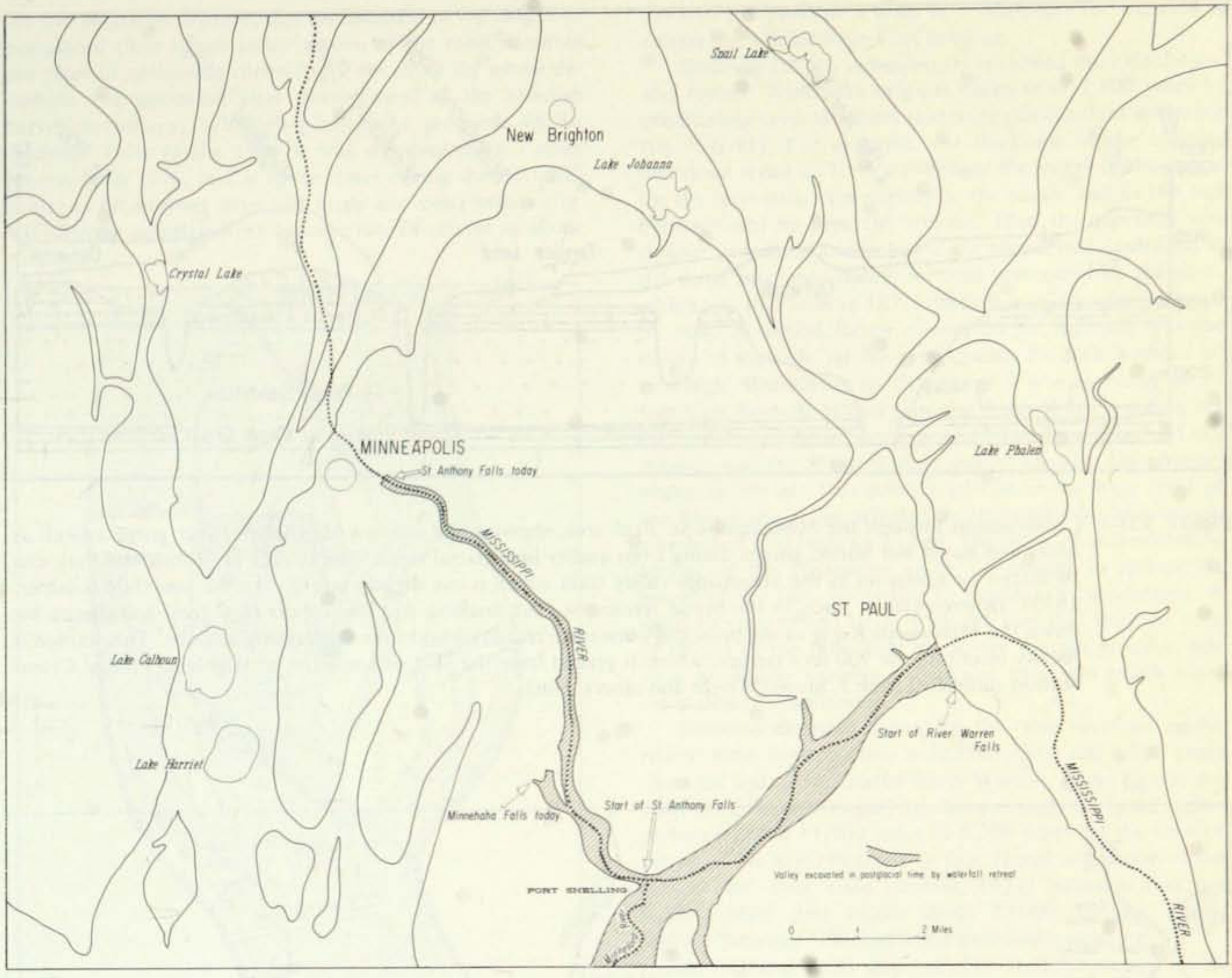


Figure VII-7. Map of buried valleys in the Minneapolis-St. Paul area, showing the gorges cut in postglacial time by the Glacial River Warren, the Mississippi River (St. Anthony Falls), and Minnehaha Creek (Minnehaha Falls). The various buried valleys date from earlier interglacial intervals. The deepest one, beneath Lake Harriet, has more than 500 feet of drift within it. The valleys transect nearly flat-lying early Paleozoic sedimentary rocks. Map modified from Payne (1965) and Mossler (this volume).

The following sequence of events is recorded: (1) Several gorges were cut into bedrock during earlier interglacial intervals. (2) All gorges were buried by Superior lobe and Grantsburg sublobe during Wisconsin glaciation. (3) During retreat of Grantsburg sublobe, a new course of the glacial Mississippi and Minnesota Rivers was established, which resulted in the formation of an outwash terrace at 810 feet (120 feet above present floodplain). (4) Subsequently, the ice retreated far enough north to form Glacial Lake Agassiz, which had an outlet via the eroding Glacial River Warren down the Minnesota River Valley. Erosion resulted in the: a) rapid removal of drift along the new segment, superposition of the valley bottom onto Platteville Limestone from Fort Snelling to St. Paul, and removal of old valley fill downstream from St. Paul; b) birth of River Warren Falls at entrance to the re-excavated segment at St. Paul; c) retreat of River Warren Falls to Fort Snelling and beyond for 2 miles to a point where the Lake Harriet buried channel was reached, and here the falls was extinguished; d) birth of St. Anthony Falls where the postglacial Mississippi River entered the River Warren at Fort Snelling; e) retreat of St. Anthony Falls 8 miles to its present position; f) birth and retreat of Minnehaha Falls in similar manner when St. Anthony Falls passed the entrance of Minnehaha Creek. (5) Subsequent diversion of the Lake Agassiz outlet to the north resulted in a change from the River Warren (high discharge, low sediment load) to the modern Minnesota River (low discharge, low sediment load), with consequent alluviation amounting to 80 feet at Fort Snelling and 175 feet at South St. Paul.

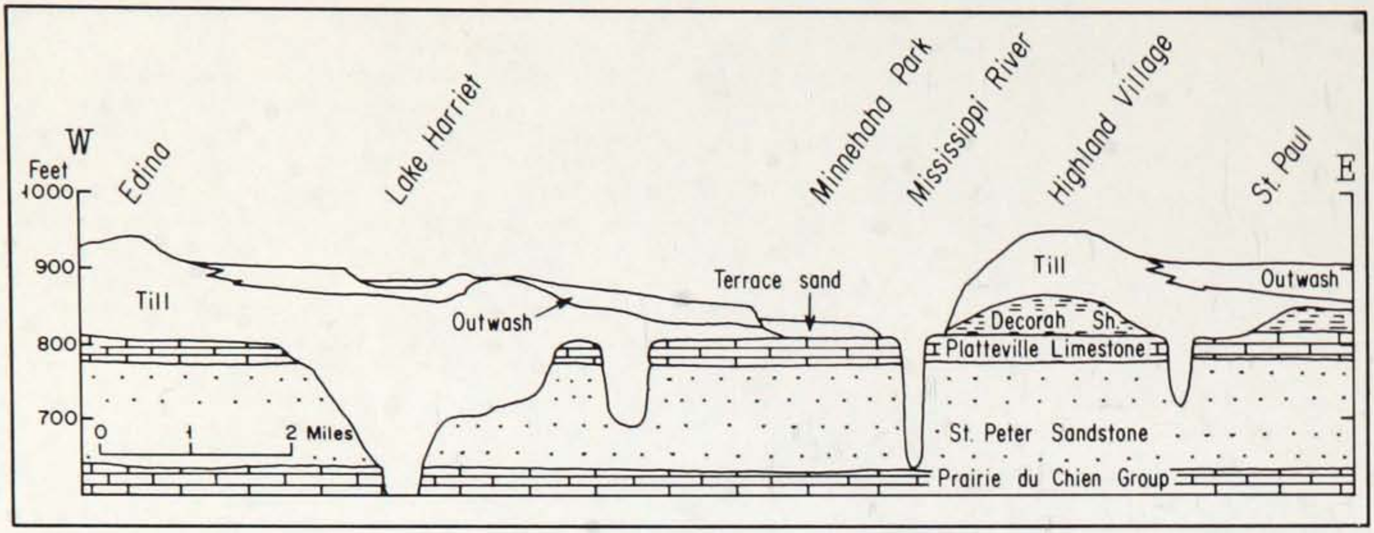


Figure VII-8. Cross-section through the Minneapolis-St. Paul area, showing the modern Mississippi River gorge as well as older and larger but buried gorges dating from earlier interglacial times. The terrace at Minnehaha Park can be traced far upstream as the Mississippi valley train which is cut slightly into the Anoka sandplain (Cooper, 1935). Downstream it leads to the broad terrace at Fort Snelling and the airport (810 feet) and thence far down the Mississippi River as the principal outwash terrace related to the Grantsburg sublobe. This terrace is locally inset into the 900-foot terrace, which is graded from the St. Croix moraine at Highland Village. Cross-section simplified from J. Stone (Wright and others, 1965).

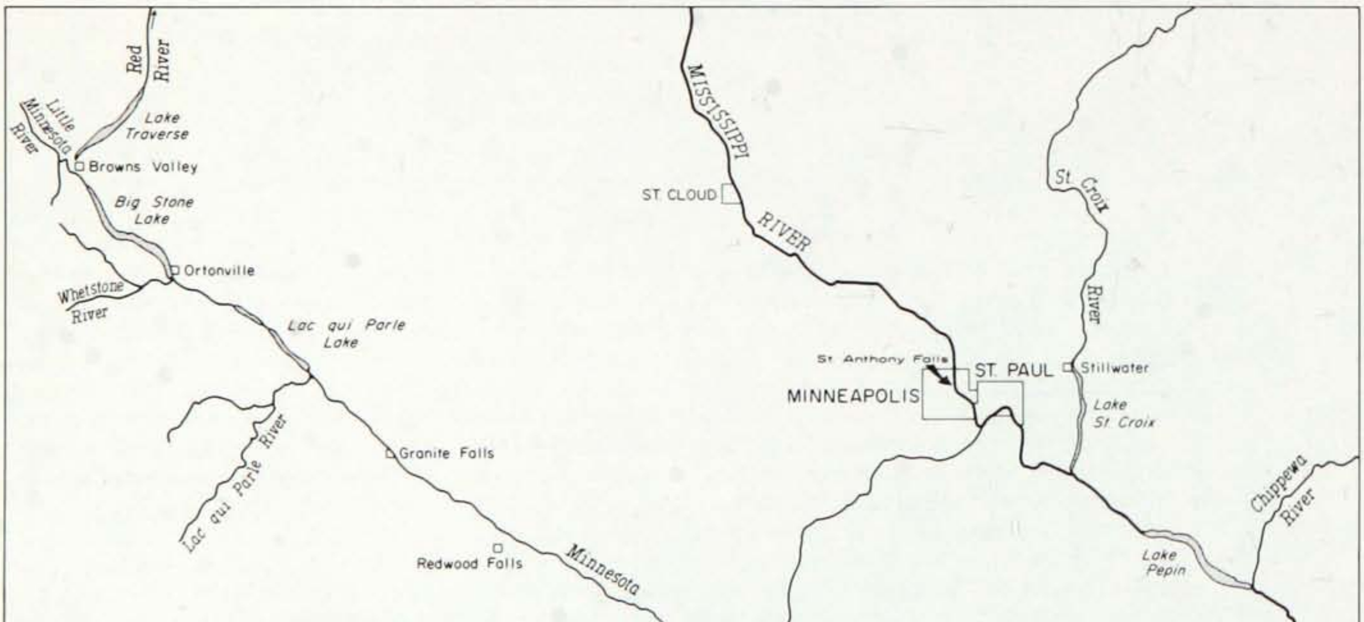


Figure VII-9. Map of Minnesota, St. Croix, and Mississippi Rivers showing locations of river lakes. Lake Traverse, Big Stone Lake, Lac qui Parle, and Lake Pepin in the valley of the Glacial River Warren were dammed by the tributaries shown, after the diversion of Glacial Lake Agassiz beheaded Glacial River Warren. Lake St. Croix was dammed in turn by the Mississippi River.

ful and thorough manner, and his estimate of the length of postglacial time stands today as one of the most accurate exercises in geological chronology, matching for astute deduction the somewhat later development of the Swedish varve chronology. Winchell plotted the positions of St. Anthony Falls at the time it was discovered by Father Hennepin in 1680, and at seven times during the next 200 years, as determined from old maps and descriptions (fig. VII-10). He calculated the average rate of retreat as about 5.5 feet per year, or a total of 7,800 years for the falls to retreat the 8 miles from Fort Snelling.

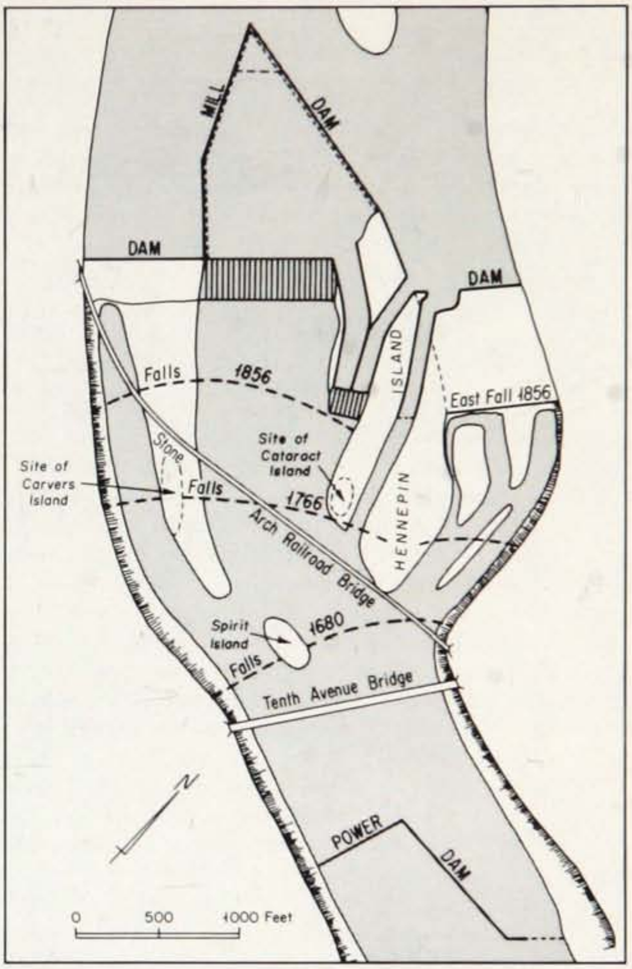


Figure VII-10. Successive positions of St. Anthony Falls from the time of its discovery in 1680 by Father Hennepin and the time of its stabilization by dam works in 1871. The rate of retreat accelerated because of the islands and the thinning of the limestone caprock (fig. VII-11). If it had not been protected, by now it would be at Nicollet Island, which marks the end of the caprock and soon after it would be extinguished. The gorge would then extend rapidly north through the unprotected bedrock of St. Peter Sandstone, and then even more rapidly up the drift-filled old gorge for several miles at least to Coon Rapids. From Winchell (1888), redrawn by Sardeson (1916).

Sardeson (1916) subsequently reviewed the calculations and revised Winchell's original estimate of 7,800 years by considering several factors that were not constant with time (fig. VII-11). For example, the thickness of the capping limestone is not uniform throughout the length of the gorge, for the formation dips slightly to the south, and its top had been beveled by previous erosion. Thus the cap rock was thinner upstream, and in fact if the retreat had continued at the same rate until today, without protection by the dam, which was first built in 1871, the falls would be extinguished by now. A second factor considered by Sardeson was the height of the falls. At the present time the falls is about 40 feet high. But earlier in its history it was as much as 75 feet high, because at that time the River Warren was in the full flood of its gorge cutting, and later, after Glacial Lake Agassiz and the River Warren ceased to exist, the channel began to fill up. This postglacial rise in the base level of the Minnesota River affected its tributary, the Mississippi, which thus also built up its bed. The wave of sedimentation spread up the Mississippi gorge far enough to reduce the height of the falls, according to Sardeson's calculations. A third correction had to do with the fact that the first mile of retreat of the falls from the River Warren junction was in a previously existing valley and thus was much more rapid than the remainder.

Sardeson's revised figure for the time involved in the retreat from Fort Snelling is 12,000 years, and 8,000 years since the end of the Glacial River Warren. These figures are even closer than Winchell's to the presently accepted radiocarbon dates of 11,700 years to 9,200 years for the time of action of the River Warren. In fact, recent evaluation of the radiocarbon time scale (Stuiver, 1971) indicates that the 9,200 carbon date equals about 10,000 calendar years, midway between Winchell's and Sardeson's estimates.

St. Anthony Falls split into two sections around an island after it had retreated about 1.5 miles. The eastern section was larger, so it retreated more rapidly, and after passing the head of the island it beheaded the western branch of the river. At the time of beheading, the western falls had retreated less than half a mile, and it was then abandoned. It had moved past a small tributary from the west, however, namely Minnehaha Creek, and it thereby dispatched a tributary falls up that stream. This event may be calculated at about 10,000 years ago. Minnehaha Falls has subsequently retreated only 1,000 feet upstream from the junction, for it is a very small stream.

Des Moines Lobe Proper

The Grantsburg sublobe, in the usual account of the glacial sequence (Wright and Ruhe, 1965), is viewed as an offshoot of the Des Moines lobe proper, which extended south as far as Des Moines in central Iowa. The principal outer feature of this lobe is the Bemis moraine, which marks not only the outer limit of undrained depressions — the usual indicator of the main Wisconsin glaciation — but also the inner limit of loess (Ruhe, 1969; Matsch, this chapter). The Bemis moraine can be traced with ease around the entire lobe, but its relation to the Grantsburg sublobe was not

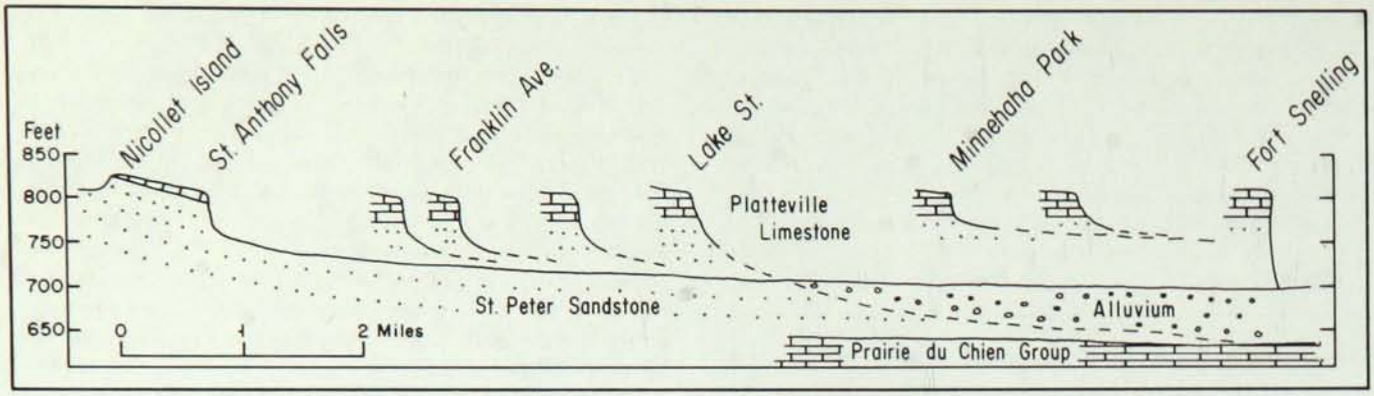


Figure VII-11. Longitudinal profile of Mississippi River gorge, as cut by the retreat of St. Anthony Falls in postglacial time from its point of origin near Fort Snelling. Intermediate profiles show changes in height of the falls. Note thinning of the limestone caprock as Nicollet Island is approached, at the edge of the bedrock basin. When the falls reached Lake Street, the Glacial River Warren (which the Mississippi River joined at Fort Snelling) was beheaded, and its successor, the Minnesota River, began to alluviate, and the fill extended up the Mississippi gorge. Redrawn from Sardeson (1916).

clear until study of the new topographic maps and the availability of new carbon dates made an alternative interpretation of the sequence more reasonable (Wright and others, in press). This story says that the "Des Moines lobe proper" is an offshoot of the Grantsburg sublobe, rather than vice versa.

The Des Moines lobe advanced down the Minnesota River Valley, after crossing the divide from the Red River Valley, apparently following the lowest course initially, as might be expected. This course took it to the big bend of the Minnesota River Valley at Mankato, and thence northeast to Minneapolis and beyond to Wisconsin, in the form of the Grantsburg sublobe, climbing to an altitude of little more than 900 feet above sea level near its terminus about 16,000 years ago (fig. VII-2I). As the ice grew thicker, it spilled over the low divide at an altitude of 1,100 feet south of Mankato and flowed down the Des Moines River valley into Iowa as the Des Moines lobe (fig. VII-2J), reaching its terminus about 14,000 years ago, according to carbon dates on wood beneath the drift (Ruhe, 1969). This sequence of events makes the Bemis moraine correlative with the Cary phase of the Lake Michigan lobe, represented by the moraines around the head of Lake Michigan in northeastern Illinois (Wright, 1971).

The Bemis moraine on the west side of the Des Moines lobe tops the crest of the Coteau des Prairies, which was originally probably a wedge-shaped bedrock plateau in northeastern South Dakota and adjacent areas, separating the Des Moines lobe from the James lobe. The gradient of the Bemis moraine along the edge of the Coteau is about 3 feet per mile. Where it joins the correlative moraine of the James lobe, the Bemis moraine has an altitude of 2,000 feet above sea level. Directly across the Minnesota River Valley, the old Alexandria moraine complex, which partly delimited the Des Moines lobe on the east, has a maximum altitude of 1,700 feet. The center of the valley is 1,000 feet above sea level, so the ice was at least 700 feet thick. The ice spread farther east, however, so it must have been even thicker in the middle of the lobe to provide the transverse surface gradient necessary for flow. It crossed over the Alexandria moraine complex and covered the outer edge of the Wadena drumlin field and even part of the St. Croix moraine near Albany. The entire Des Moines lobe, being at least 300 miles long from the north end of the Coteau des Prairies to central Iowa, and only about 130 miles broad, resembled a wide valley glacier as much as it did a bulge on an ice sheet. Unlike a valley glacier, however, it apparently had a major flow component to the sides (Wright and others, in press).

Subsequent thinning of the ice left a series of "lateral moraines" on the west side, *i.e.*, on the east scarp of the Coteau. They can be traced around the end of the lobe in central Iowa, where they are much broader and less pronounced. In back of the Bemis moraine is the Altamont, and in back of this the Algoma. Younger features mapped by Winchell and Leverett as moraines adjacent to the Minnesota River Valley may be partly gravel ridges of uncertain origin, or they may be till areas eroded by lake-outlet streams. The town of Mankato, located in the center of the Des Moines lobe area, shows deep exposures of the drift along the sides of the Minnesota River Valley. This locality gave the name originally to the entire late-Wisconsin ice advance of the Des Moines lobe (Leighton, 1933), but more recently the term "Mankato phase" has been applied only to a post-Bemis interval of Des Moines lobe activity, culminating in the Algoma moraine, which has radiocarbon dates near 13,000 BP. The Grantsburg sublobe by this time was probably entirely wasted, except, of course, for some buried ice blocks. The Des Moines lobe itself had uncovered the Mankato area by 12,700 years ago, according to a basal radiocarbon date from a pond deposit at nearby Madelia (Jelgersma, 1962).

Nickerson-Alborn Phase of St. Louis and Superior Lobes

St. Louis Sublobe

The Des Moines lobe had a second eastward protrusion in addition to the Grantsburg sublobe. This broad tongue, the St. Louis sublobe, spread east from the Red River lowland into the Red Lakes lowland of northwestern Minnesota, the area once occupied by the Wadena lobe (fig. VII-2K). The latter situation differed from the former in two respects, however. First, the ice must have left the main lobe south of the Canadian border, because its drift is characterized by fragments of Cretaceous shale as well as a silty texture, and the distribution of Cretaceous shale demands such a course. Second, this time there was no obstruction to the eastward extension of the ice, as had been the case when the Rainy lobe blocked and diverted the Wadena lobe to the southwest.

The St. Louis sublobe was restricted in its southward expansion by the north flank of the Itasca moraine, but the ice buried several of the tunnel valleys that had formed during wastage of the Wadena lobe from the Itasca moraine. The north flank of the sublobe was restricted by higher ground in Canada. The ice extended east beyond the interlobate junction of the Itasca and St. Croix moraines near Walker, and then spread south in a sub-sublobe as far as the Mille Lacs Lake area, where it crossed and then covered the inner flank of the Mille Lacs moraine (fig. VII-5).

Another sub-sublobe was diverted around the southwest end of the Giants Range and an old moraine east of Grand Rapids, and it then flowed north to the range and east along its south flank as far east as Aurora (Winter, 1971). It reached an altitude of 1,550 feet above sea level, and at one point, near Buhl, it even extended through the range in a narrow finger. This sub-sublobe came to a limit as it buried the western edge of the Toimi drumlin field and parts of the Highland, Cromwell, and Wright moraines at an altitude of about 1,400 feet. Here it formed its own moraine—the Culver moraine, with outwash plains grading off east to the St. Louis River (Wright, this chapter, fig. VII-32).

A third sub-sublobe overrode the west end of the Giants Range and extended northeast as far as Lake Vermilion, where it was blocked in part by the Vermilion moraine. But it overrode the western part of this moraine too and extended to the Canadian border near Crane Lake.

The drift of the St. Louis sublobe is variable but distinctive. In the western area—north of the Itasca moraine—the till is siltier than the Wadena lobe drift it overlies, and it contains fragments of Cretaceous shale. Between the Mille Lacs area and the Giants Range, however, the till is typically a pebbly clay, with local occurrences of the western-type shale-bearing till. The color of the pebbly clay or clayey till ranges from light to dark to reddish brown. It resembles the clayey till of the Split Rock phase of the Superior lobe in texture and structure, but is less reddish—in some cases only slightly. It also consists of reworked lake deposits.

Leverett (1932) correctly mapped the distribution of much of the St. Louis sublobe drift, but he offered no explanation for the distinctive color and clayey character of much of it. I (1955) proposed that the clayey facies of the drift represented actually a northerly and westerly expansion of the Superior lobe—thus an extension during the Split Rock phase. Subsequent mapping of drifts in northeastern Minnesota, with availability of new topographic maps, reveals a consistent 15-mile gap between the limits of the red clayey till of the Superior lobe and the Culver moraine of the St. Louis sublobe, and it also shows that the former was restricted to altitudes below 1,300 feet at the head of the Lake Superior basin, whereas the latter reached 1,550 feet on the Giants Range.

It is now considered that the clayey till owes its origin to reworking of clay and silt deposited in the Lake Upham and Lake Aitkin basins during the Automba-Vermilion phase of the Superior and Rainy lobes. The reddish-brown color is attributed to clay originating in the Superior lobe, the dark-brown color to clay in the Rainy lobe, and the light-brown color to clay in the St. Louis sublobe, which may already have been advancing from the west at this time, far enough to feed calcareous outwash into Lake Aitkin I. When the St. Louis sublobe advanced in the Alborn phase, the northern sub-sublobe, which remained north of the Giants Range, carried normal brown silty till from the west (Winter, 1971). The other two sub-sublobes filled the Lakes Aitkin and Upham basins, eroded the lake sediments, and redeposited them as till around the southern, eastern, and northern margins. In some exposures all three color types are present, even in discrete stratigraphic arrangement with pebble bands between. In other exposures lenses of normal western-type shale-bearing St. Louis sublobe till occur. The entire till complex may be attributed to deposition by various threads of ice that had access to the substratum at different points, some local, some far to the west.

The St. Louis sublobe in the Alborn phase filled the low ground of north-central Minnesota, and it was bounded on the southeast and south by moraines. Although some outwash must have escaped south through the bounding Mille Lacs moraine into the Mississippi River, the remainder must have been directed east down the St. Louis River, for between these two areas the bounding moraines—above an altitude of 1,300 feet—prevented the southward spread of meltwaters. If outwash was directed down the St. Louis River at this time, the Superior lobe must have been withdrawn sufficiently far into the basin to allow the river to be diverted southwest down the Moose River to the St. Croix River. This relation provides the opportunity to correlate the Alborn phase of the St. Louis sublobe with the Nickerson phase of the Superior lobe.

As the St. Louis sublobe withdrew from its Alborn maximum, Glacial Lakes Upham II and Aitkin II formed at its front. The latter probably drained into the former, because it has no large spillway to the Mississippi. Lake Upham in turn drained down the St. Louis River, and as the lakes enlarged so did the outlet stream, making possible eventually the formation of wide erosional valleys.

Superior Lobe

The Superior lobe at the end of the Split Rock phase had retreated once again into the Lake Superior basin, and it readvanced in an even narrower lobe to the Nickerson-Thomson moraine (fig. VII-2K). The Nickerson moraine, along the Carlton/Pine County line east of Moose Lake, consists of a very hummocky belt of morainic topography about 5 miles broad, with a local relief of 100 feet, extending eastward for about 20 miles, until it becomes lost on the south flank of the Lake Superior basin in Wisconsin (Wright, this chapter, fig. VII-34). Deep exposures of massive red clayey till are abundant in this area. The moraine is bordered and offlapped on the south by the rather extensive Willow River outwash plain, which leads into the Kettle River and thus to the St. Croix. The point of the Superior lobe at this time was 20 miles in back of the point of the lobe in the Split Rock phase, when the outwash had been directed down the old tunnel valleys. This time the outwash escaped from the tunnel valley system and fed more directly into the St. Croix River valley, which was more deeply cut than before because Lake Grantsburg had been drained, the Grantsburg sublobe withdrawn, and the Anoka sandplain dissected at its eastern end. In fact by this time the River Warren was active in the Mississippi valley, and the wave of deep dissection by the River Warren may have extended far up the St. Croix River, bringing about the dissection of the Anoka sandplain and even some of the Kettle River gorge.

The Nickerson moraine extends west barely across the Moose River valley, and then it turns back to the northeast as the Thomson moraine. This feature maintains a clayey character and hummocky topography until it recrosses the Moose River; from there on to the northeast it consists largely of red sandy till and outwash. It is fronted by distinct outwash fans and plains at an altitude of about 1,200 feet. These have since been partly dissected, but they can be traced west far enough to show that the outwash went around the nose of the ice lobe by way of the Glaisby Brook channel into the Kettle River. The Glaisby Brook channel is occupied by an underfit stream, and it seems likely that the stream that formed the valley was not only large but was also a dissecting rather than a depositing stream. It will be recalled that the St. Louis sublobe was standing to the west at this time, and that, as it withdrew, proglacial Lake Upham II formed at its front and discharged water down the St. Louis River. The addition of clear lake water, from which much of the glaciofluvial sediment had been removed, made the dissection of stream valleys easier.

As the St. Louis sublobe withdrew farther, Glacial Lake Upham II became larger, and was joined by Lake Aitkin II. By this time the outlet stream down the St. Louis River valley was a major feature. It continued to be diverted at the Thomson moraine, but its erosion of the moraine front produced successively lower diversion channels of successively larger size (Wright and others, 1970). In the course of this erosion, broad segments of the Thomson moraine or outwash plain became isolated. They undoubtedly contained buried stagnant ice. Ultimately, the St. Louis River broke through the last of the moraine, which by this time had been abandoned by the ice front in its retreat into the Lake Superior basin. The river flowed into proglacial Lake Nemadji, which also had its outlet southwestward into the Moose River—the seventh of the recognizable diversion channels (Wright, this chapter, fig. VII-34). This outlet was the largest of all, at an altitude of 1,060 feet above sea level, with boulders scattered along the floor where it cut through the Thomson moraine. Sand was deposited in the nearshore portions of Lake Nemadji, and a sharp strand line was formed at its edge. The strand line, and thus the lake, can be traced about 15 miles east from the outlet along the Wisconsin side of the basin.

As the ice withdrew farther into the basin, a lower outlet was uncovered in Wisconsin at a point about 50 miles east of the Moose River outlet. It led to the Brule River, a headwater of the St. Croix River. The lake level lowered about 50 to 1,010 feet and stabilized as Glacial Lake Duluth. Because Lake Nemadji was a much larger lake, its drainage cut a deeper gorge in the St. Croix, and the Kettle River gorge was left hanging.

Further retreat of the Superior lobe uncovered still lower outlets to the east, to the Lake Michigan and Lake Huron basins. Various lower lake levels are recorded along the north shore of Lake Superior (Farrand, 1969; Sharp, 1953). During this time the land was being tilted southward as a delayed response to removal of the load of glacier ice from the earth's crust, and this factor had some influence in the location of outlets to the lakes. The strand lines—at least those from Duluth eastward—are thus all inclined to the southwest.

Meanwhile, as the Superior lobe was retreating into the Lake Superior basin for the final time, and large proglacial lakes were forming in front of it, as well as in front of the retreating St. Louis sublobe, the Des Moines lobe withdrew into the Red River lowland, and Glacial Lake Agassiz formed at its front. Upham (1896) believed that the western (Minnesota) ice lobe retreated east into the area now called the Alexandria moraine complex, with Lake Agassiz forming to the west. Nikiforoff (1947), who mapped the soils and beaches of much of this area, considered that a large mass of separated ice remained at the southern end of the basin and that Lake Agassiz actually started in the north and spread south. But Leverett's (1932) concept of the Des Moines lobe retreating regularly to the north seems to be most reasonable, even though some of the retreatal moraines he mapped across the valley are not recognizable.

As the Des Moines lobe retreated up the Minnesota River Valley, and before it reached the Red River Valley, small proglacial lakes were formed. The most conspicuous of these is Lake Benson, which was held by a narrow moraine near Granite Falls. It is marked by thin deposits of silt in its southern part and by broad outwash fans of the Pomme de Terre, Chippewa, and Minnesota Rivers in its northern part, which fed from the ice front to the north. It has no strand lines, so it was probably short-lived. Its maximum depth was about 50 feet. The most striking feature of the area is a series of sand-floored linear channels cut to a depth of 10-20 feet below the general area of silt-covered till. Some of the low ridges between these channels were mapped by Leverett as lateral moraines of the still-shrinking Des Moines lobe. They are now interpreted as channels formed when Lake Benson catastrophically drained, presumably caused by breakage of the ice-cored moraine that dammed it (Matsch and Wright, 1967).

Glacial Lake Agassiz

From the moraine at Granite Falls, the Des Moines lobe withdrew to the Big Stone moraine, which essentially forms the divide between the Minnesota and Red River Valleys. This moraine has low relief and is inconspicuous, but it served the purpose of damming southward drainage from the ice front and thus forming Glacial Lake Agassiz.

At least two small proglacial lakes preceded Lake Agassiz behind the Big Stone moraine. One of them, Lake Milnor, was largely on the South Dakota side; it drained through a prominent channel at an altitude of 1,100 feet. The other, on the Minnesota side, drained through the Fish Creek channel at an altitude of 1,070 feet (Matsch and Wright, 1967).

As the ice withdrew and Lake Agassiz enlarged, its outlet became established at Browns Valley, where it remained for a long time. The Big Stone moraine, which contained large locally derived boulders of granite, was dissected by the outlet stream, the incipient Glacial River Warren, which cut down as well through outwash valley trains formed in front of the Big Stone moraine. In time the granite boulders paved the outlet channel through the moraine, and downcutting temporarily ceased (fig. VII-12). This allowed stabilization of the lake level, and a conspicuous strand line—the Herman beach—was formed (Matsch and Wright, 1967). Its altitude is 1,060 feet in the outlet area. On the east side of the lowland it can be traced northward almost continuously on soils maps, topographic maps, and aerial photographs for 150 miles to the Maple Lake area, where it turns abruptly east into the Red Lakes lowland (fig. VII-2L). In this northward-trending segment, its altitude increases to about 1,160 feet as a result of postglacial crustal tilting. The eastern area of Lake Agassiz at this time extended for about 150 miles to the region north of the Giants Range, and it buried the western end of the Vermilion moraine of the Rainy lobe. Beyond this beach to the east were older, higher proglacial lakes, which must have emptied eastward into the Lake Superior basin.

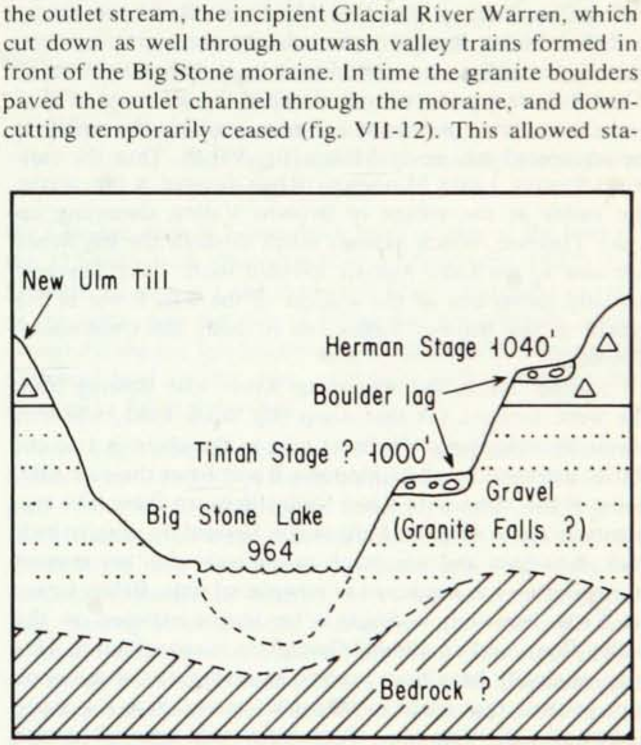


Figure VII-12. Cross-section of Minnesota River Valley at Big Stone Lake, showing two boulder-paved terraces of the Glacial River Warren that grade upstream to the strand lines of Glacial Lake Agassiz. Big Stone Lake was formed by the fan of the Whetstone River, when the flow of River Warren ceased as a result of the shift of the Lake Agassiz outlet to the north. From Wright and others (1965).

The Herman beach on the west side of the Red River lowland is traceable for 350 miles from the southern outlet into Manitoba (Elson, 1967). The lake at this time may have had a central lobe of ice, or it may have been completely open. In the latter case, it must have been at least 400 feet deep in the center, for the present sediment surface in the center of the lowland opposite the Maple Lake bend has an altitude of about 860 feet, the beach has an altitude of 1,160 feet, and the thickness of clay is about 100 feet. At least the lake was open enough so that waves could form prominent strand lines. It was about 65 miles wide in its southern arm.

For most of its length the Herman beach consists of a single or a few beach ridges of sand or pebble gravel, with characteristic cross-bedding, but at the Maple Lake bend as many as 10 closely spaced ridges occur. Typically, spits, hooks, interridge lagoons, and low cliffs are found. In the eastern arm, which was once as much as 100 miles broad, no lake sediment was deposited, except possibly for a thin smear of sand or pebbles. Apparently the wave action generated by westerly winds over a fetch of 100-250 miles was enough to prevent sedimentation in water shallower than about 250 feet. This area contains none of the undrained depressions characteristic of the ground moraine shoreward from the beach, so erosion was deep enough to eliminate the knolls between the depressions. In much of this area of the eastern arm, the till still reveals a conspicuous linear pattern that probably represents some type of minor moraine, so wave erosion must not have been strong enough to obliterate the pattern or to put a thick gravel on the till surface.

The best radiocarbon date for the withdrawal of Glacial Lake Agassiz from the Herman beach is 11,740±200 (Y-1327), taken from the basal organic sediment of a beach pond on the east side of the lake. Such a pond is not subject to problems in many lakes caused by the persistence of buried ice, because the lake water had caused the melting of all buried ice along the shores. The recorded radiocarbon date is approximately equivalent to the end of the so-called Two Creeks interstadial interval of the Lake Michigan lobe. But at this time the St. Louis sublobe was completely wasted, and Glacial Lake Aitkin was dry, at least temporarily, according to the occurrence of buried soil and peat which formed at this time near its southern end (Farnham and others, 1964). The Superior lobe, however, still stood at the Thomson moraine, and the Rainy lobe was somewhere in northwestern Ontario, far enough south so that all eastern outlets of Lake Agassiz were still closed by ice (Zoltai, 1961).

The lowering of Lake Agassiz from the Herman level can be attributed to the breaching of the boulder pavement at the outlet channel (Matsch and Wright, 1967). As the lake grew larger with ice retreat, the volume of water going through the outlet increased proportionately, and ultimately the velocity increased to a point at which the boulder pavement could be removed. Erosion of the outlet proceeded until another boulder pavement was formed to match the Norcross strand line, at an altitude of 1,040 feet in the outlet area. At this time the breadth of the lake was smaller, and the eastern arm became much reduced. Repetition of the process produced the Tintah strand line at 1,020 feet, and then the Campbell strand line at 980 feet (fig. VII-12). At this level the outlet stream reached to granite bedrock, and further downcutting ceased. The Campbell strand line is well developed on both sides of the lowland, and the eastern arm was practically eliminated, at least within Minnesota. The strand line rises northward to about 1,080 feet near Roseau, with an overall gradient thus of 1 foot/mile. The lake at this time was perhaps 200 feet deep opposite Maple Lake. The strand line is locally complex, and in some places distinct wave-cut cliffs can be identified. The shallow water sediments for the Campbell phase are mostly sand in the southern part of the basin; to the north, the broader areas of sand, as well as the more numerous sand beaches, can be attributed to the presence of a previously formed vertical wedge of sand whose top was simply eroded and spread laterally by the wave currents (Winter, 1967). Still farther north, near Roseau, no sediment at all exists for at least 30 miles lakeward from the Campbell beach, and linear ground moraine forms the dominant pattern.

The Campbell strand line was abandoned 9,200 ± 600 years ago (W – 1057). This event records the retreat of the ice far enough into Ontario to uncover outlets east to Lake Superior (Zoltai, 1961). There is evidence that the lake level had previously lowered enough to allow the deposition of alluvial deposits, including wood fragments, on the flanks of the lowland below the level of the Campbell beach. Radiocarbon dates on such wood show a broad range ending about 10,000 BP, and Elson (1967) postulated not one but two low-water intervals for the lake. In any case, the return of the lake to the Campbell level could have been caused by a readvance of the ice lobe from the northeast over the outlet to Lake Superior, or alternatively by the southward tilting of the land. Correlation of ice margins in this region with lake-outlet channels and with lake phases is not simple, partly because the topographic control is inadequate, and partly because material for radiocarbon dating is lacking. One possible direct stratigraphic tie is afforded by a band of red clay in the upper part of the Lake Agassiz deposits north of Rainy Lake. The red clay clearly came from a stream whose source was to the east, in a lobe from the northern part of the Lake Superior basin (Marks moraine) where the drift is red. The red band is probably in deposits dating from the last Campbell phase of Lake Agassiz (Elson, 1967); if so, it dates the Marks moraine as about 9,500 BP.

As Lake Agassiz retreated from the Campbell beach for the last time the southern outlet was abandoned for good, and eastern outlets to Lake Superior were utilized as several lower strand lines were formed. Even an outlet northwest to the Mackenzie River occurred at one time (Elson, 1967). Although the northern part of the lake expanded, as the ice withdrew toward Hudson Bay, the southern part became restricted, despite continual southward tilting of the land. By 8,300 years ago the lake was restricted to Manitoba, although still with an outlet eastward to Lake Superior. By 7,300 years ago, sea water worked its way through Hudson Straits and around the west side of Hudson Bay, and Lake Agassiz drained to this point, as do modern remnants like Lake Manitoba and Lake Winnipegosis. Within Minnesota, the only large remnants of Lake Agassiz are the huge Red Lakes in Beltrami County, Thief and Mud Lakes in Marshall County, and Rainy Lake and Lake of the Woods on the Canadian border.

The final beheading of Glacial River Warren 9,200 years ago caused some interesting developments along its entire length, within what are now the Minnesota and Mississippi River valleys, and this allows us once again to tie together the history of two widely separated drainage basins. The abrupt termination of the River Warren meant that its successor stream, the Minnesota River, no longer had the volume to supply the velocity necessary for transporting the sediment supplied to it by its tributary streams. These tributaries accordingly dropped the sediment in alluvial fans at their points of entrance, causing the valley to be segmented into several lakes (fig. VII-9). Thus the eastward-flowing Little Minnesota River formed a fan across the valley at the village of Browns Valley, damming up Lake Traverse, which extends north through the Big Stone moraine to the Lake Agassiz lowland itself. Lake Traverse actually forms one of the sources of the Red River of the North so the Browns Valley fan is really the Continental Divide.

Farther south, the Whetstone River, also flowing from the west, forms a fan that dams Big Stone Lake, which is about 30 miles long. The next lake in the chain is Lac qui Parle, dammed by the Chippewa River from the east side. Most of the Minnesota River Valley between these lake segments as far as Granite Falls is still floored by granite bedrock, however, and not much in the way of either erosion or deposition has happened in postglacial time. Below Granite Falls, however, bedrock is no longer exposed on the valley floor, and an alluvial floodplain is conspicuous. This may originally have been marked by river lakes of the same type as those that occur further upstream, such as Big Stone Lake and Lake Traverse, which have since been filled with sediment. The only remaining lake of this type is Lake Pepin, far downstream, 50 miles south of St. Paul. Here the dam was caused by another Chippewa River, from the Wisconsin side (fig. VII-13). The sandy sediment carried by this steep tributary was easily transported by the River Warren, but in postglacial time the relatively feeble Mississippi River has not only not been able to remove the fan dam but has not yet filled the lake with sediment. There is evidence that Lake Pepin, which has a maximum depth of 50 feet, originally extended upstream as far as St. Paul, where borings

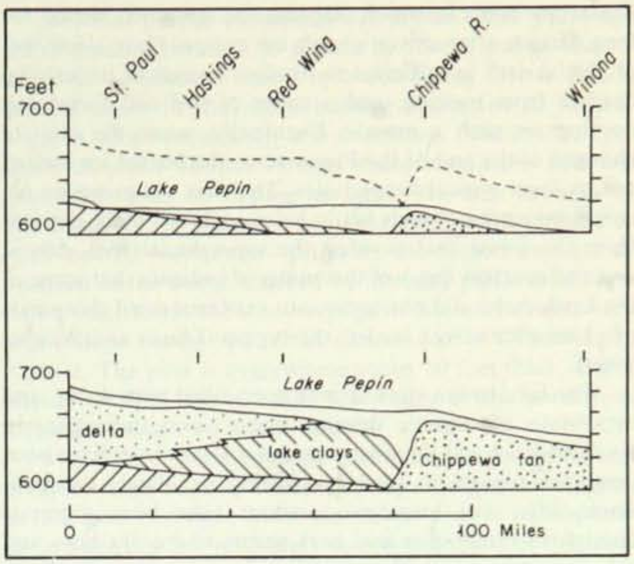


Figure VII-13. Longitudinal section through Mississippi River floodplain from St. Paul to Winona, showing formation of Lake Pepin by the deposition of a fan from the Chippewa River, followed by its gradual filling by lake clays and by progradation of the Mississippi River delta at its head. Alluvial deposition on the Mississippi near Hastings in turn dammed the St. Croix River, forming Lake St. Croix. Redrawn from Zumberge (1952).

for a bridge show a 7-foot-thick layer of clay at the base of the 175-foot-thick alluvial sediment (Zumberge, 1952). Ultimately, of course, Lake Pepin will fill by continued growth of the delta, and the rate of filling has undoubtedly accelerated during the last hundred years as a result of increased sediment load in the river, related to soil erosion in the Minnesota River watershed.

An interesting consequence of the postglacial sedimentation of the Mississippi River is the formation in the lower part of the St. Croix River of Lake St. Croix, which joins the Mississippi 25 miles downstream from St. Paul. The St. Croix River, like the Minnesota-Mississippi, was deeply excavated in late-glacial time by the outlet waters of a glacial lake, namely Lake Duluth, and at this time the great potholes in the basalt near Taylors Falls were excavated, as was the gorge itself (Alexander, 1932). After this outlet stream was beheaded when the Superior lobe retreated enough to uncover another channel farther east, the St. Croix River lost its erosive power. But unlike the Mississippi River, which was supplied with a great load of sediment from tributaries draining silty till in sparsely forested country, the St. Croix River drained regions of stony, sandy till and bedrock in heavily forested country. Sedimentation in the St. Croix therefore could not keep pace with that in the Mississippi, which thereby essentially built a dam across the mouth of the St. Croix, producing a lake 20 miles long and up to 30 feet deep, which of course has a slowly prograding delta at its head (at Stillwater).

Summary of Mississippi River History

The Minnesota-Mississippi-St. Croix river system altogether has a complex history of cutting and filling that illustrates in one of the finest ways the principle of the graded stream, as enunciated by Mackin (1948). It is worth summarizing the events so that all the shifts can be seen together.

During Wisconsin glaciation, a myriad of old river gorges in the Minneapolis-St. Paul area became plugged with drift, and outwash sediment was fed almost directly into the Mississippi River. The two phases of Wisconsin glaciation that affected this area resulted first in the formation of the St. Croix moraine and then in the overlap of this moraine by the Grantsburg sublobe. These events are represented along the Mississippi River by terraces starting respectively at altitudes of about 900 feet and 810 feet, in a fill-cut-fill relation. The rapid filling of the valley at these times, after its deep dissection during the preceding interglacial interval, resulted from overcharge of the system by glacial sediment such that, in Mackin's terms, the stream was forced to deposit in order to build up its gradient and thereby provide the greater velocity required to transport the increased sediment load. Far downstream, the deposition caused backflooding of non-glacial tributary valleys and even the complete isolating of sections of the valley wall, as at Frontenac, which is essentially an island of upland bounded by two broad channels of the Glacial Mississippi River.

Retreat of the Grantsburg sublobe and the Des Moines lobe led to stepwise erosion of the 810-foot-high terrace. The rationale here is that with ice retreat the bulk of the coarse outwash is deposited immediately in front of the ice, so that downstream the balance of the load can be transported on a lower gradient—thus dissection occurred to attain this gradient.

The dissection was accelerated as the Des Moines lobe withdrew into the Red River lowland and formed Glacial Lake Agassiz and the River Warren about 12,000 years ago. At about the same time proglacial Lakes Upham and Aitkin in northeastern Minnesota drained into the St. Louis River and thence down the St. Croix River, to be joined soon after (in time) by the outlet waters of Glacial Lake Duluth. These lakes screened out practically all glacial sediment, so that even lower gradients sufficed to carry what load was supplied by tributaries. The Mississippi River below St. Paul rapidly excavated its buried channel down to bedrock. The St. Croix River and the Mississippi above St. Paul had established new courses, however, and rapids and waterfalls were formed as dissection proceeded. The main waterfall on the Mississippi, starting at St. Paul, retreated upstream to Fort Snelling and thence up the Minnesota River until it ran out of Paleozoic caprock. At Fort Snelling it passed the junction of the upper Mississippi River, which at this time was a smaller segment because it had no glacial lake at its head. (Glacial Lake Aitkin drained east to the St. Louis River.) A smaller waterfall proceeded up the Mississippi, as St. Anthony Falls, and it in turn dispatched a third-order falls up its tributary, the Minnetonka River.

Meanwhile, further ice retreat caused diversion of the outlets of these several proglacial lakes into other watersheds, and the Minnesota and St. Croix Rivers were essentially beheaded. The loss of these great water volumes upset the balance once again. The sediment loads of the streams, which by this time consisted only of material added by the various non-glacial tributaries downstream from the divides, nonetheless were too great to be carried by the much-reduced volume of river water. Some of the tributaries were particularly well loaded with sediment, and these built fans across the Mississippi River floodplain, thereby segmenting the valley into a series of lakes. Many of these lakes have been filled by delta deposition at their heads, but several still exist even after 9,200 years of postglacial time. Eventually, the river floodplain will be built up once more to a gradient that will supply the velocity necessary to transport the sediment load supplied by the tributaries. Actually, the river regime has been so upset during the last hundred years by the works of man that it is difficult to evaluate the natural tendencies of the river: dredging for navigation prevents channel build-up; levees constructed along stream banks inhibit floodplain deposition; dam building reduces flood flows; groundwater depletion and municipal water-use change the water volume; and soil erosion on the watershed increases sediment loads.

POSTGLACIAL HISTORY

Compared to the drastic changes on the Minnesota landscape brought by glaciation, the postglacial geologic history has been mild indeed. Slopes soon became stabilized by vegetation, and most subsequent geologic action has been confined to the major river valleys and the lake shores, where water movements have been sufficiently vigorous to erode and transport rock and mineral particles. Meanwhile, steady deposition of finer mineral particles and organic detritus into lakes has produced a stratigraphic record of landscape and lake development that is amenable to detailed study.

The postglacial history of the Mississippi-Minnesota-St. Croix River system has already been summarized. In this system the great change brought about by glaciation in the headwaters resulted in an equally great adjustment after removal of the headwater factor, and the postglacial regime has produced an uncommonly conspicuous record of geologic processes—formation of river lakes and their gradual filling by delta deposition. In addition, typical features of river floodplains, such as meanders, oxbow lakes, point bars, natural levees, and floodplain lakes, can be seen along these rivers. These features indicate the complexities of the depositional environment during the postglacial epoch of gradual valley filling that has prevailed since the deep dissection of late-glacial time. Whether a quasi-equilibrium has been reached cannot be determined without some means of dating the floodplain deposits, to see if the rate of filling has decreased to very low values.

The gradual modification of lakes by geologic processes in postglacial time has been conspicuous throughout Minnesota. Almost all the lake basins result from the irregular down-melting of stagnant glacial ice during the waning phases of glaciation. Moraines and outwash plains in particular are areas in which stagnant ice remained buried for long periods after active ice left the region. Only a few feet of glacial drift is sufficient to insulate a mass of underlying deadice from melting, and a cover of soil and forest can develop on such a mantle. Eventually, when the climate changed at the end of the Pleistocene, the buried ice melted out to form typical kettle holes. The first sediments to accumulate in many such kettle holes included plant detritus from the forest that covered the superglacial rock debris, and radiocarbon dates of this material indicate that some of the kettle holes did not come into existence until thousands of years after active ice left the region (Florin and Wright, 1969).

The lake basins thus formed soon filled with water, and deposition of organic detritus began soon thereafter. In medium-sized and large lakes, wind-driven waves have been powerful enough to erode projecting headlands and produce cliffs, and longshore currents have been vigorous enough to build spits and bars across re-entrant bays and otherwise to smooth the coastline. As a result, many lakes have been segmented into two or more separate basins, and the present-day shapes are far different from those of the original ice-block depressions.

In the course of infilling of lakes by sedimentation, the remains of many organisms have been included and preserved as fossils. Some of the fossil types provide a continuous record of environmental history since the inception of the lakes. Perhaps the most informative fossil types are pollen grains, which are blown from many types of flowering plants, especially trees, in the area surrounding a lake. Study of the pollen stratigraphy of a large number of Minnesota lakes and bogs has provided the basis for the following summary of the late-glacial and postglacial vegetational and climatic history.

While the Superior and Rainy lobes were still advancing and retreating in northeastern Minnesota, an area in the central and east-central part of the state featured tundra vegetation, and the sediments deposited in the lakes of this area have a relatively high component of silt carried from the unstable hill slopes. At this time most of southern Minnesota was marked by boreal spruce forest, which spread north about 11,500 years ago as the ice finally left the state (Wright, 1968). The spruce forest itself could not survive the climatic change, and it was replaced by other forest types in rapid succession, principally by pine about 10,000 years ago. The trend toward a warmer, drier climate continued, and by 8,000 years ago prairie vegetation, which had succeeded the spruce forest almost directly in the Dakotas, had spread to central Minnesota. By 7,000 years ago the trend reached its maximum, and at that time the prairie-forest border was about 75 miles northeast of its present position.

At the time of maximum warmth and dryness, many of the smaller lakes and marshes in western and southern Minnesota periodically dried up, or at least experienced low water levels. The changing lake conditions are recorded not only by the pollen sequence but also by larger plant fossils, such as seeds and fruits (Watts and Winter, 1966), and by algae (*e.g.*, diatoms) and various invertebrate organisms (mollusks, cladocerans, ostracodes).

Reversal of the climatic trend about 7,000 years ago led to gradual invasion of prairie by forest, and to the advance of coniferous forest into deciduous forest. Poorly drained areas in the north became converted to blanket bogs, in which perennially wet conditions inhibit decomposition of plant detritus. Under these conditions peat accumulates, even on sloping surfaces. The best example of this condition is the great peatland north of Red Lake in north-central Minnesota—probably the largest continuous peatland in the world, marked by intricate patterns of vegetation that reflect the slow seepage of water down the very gentle slope on the bed of the eastern arm of Glacial Lake Agassiz. The peat is everywhere about 10 feet thick, and it started to form about 3,000 years ago, as the coniferous forest advanced west on the upland.

Meanwhile, sedimentation continued in the lakes, and many of them became shallow enough around the margins so that mats of sedges, heath plants, and conifer trees spread toward the centers, ultimately converting many of the lakes to bogs, and by now there are more bogs than lakes in northern Minnesota. In the deciduous forest and prairie regions, fringing sedge mats also developed around lakes, but their rate of infilling seems to be less than that of the northern bogs. Nonetheless, the natural fate of all such lakes is extinction, and in another 5,000 years only the largest and deepest lakes will be left, at the present rate of filling and fringing. Accelerated organic productivity in polluted lakes and increased silt inflow from soil erosion in agricultural regions will certainly hasten the process of filling, and the depth and quality of many lakes have already been grossly altered as a result of such disturbances. A lake as an ecosystem is affected by many environmental factors, of which geomorphic and hydrologic factors are most important. The quasi-equilibrium established under natural conditions over thousands of years, with the gradual shifts impelled by climatic changes, has been dramatically upset in many lakes by human disturbances of various kinds, and many of the changes are faithfully recorded in the sediments, which are subject to normal kinds of stratigraphic investigation.

QUATERNARY GEOLOGY OF SOUTHWESTERN MINNESOTA

Charles L. Matsch

The land surface of southwestern Minnesota is underlain mainly by sediments of Quaternary age, and most of the landforms themselves are the result of erosional and depositional events that were closely controlled by climatic fluctuations during the Pleistocene Epoch. The most dramatic result of these climatic changes was the periodic advance and retreat of glacial ice across the region. Geologists have studied the great variety of Quaternary sediments—especially the glacial drifts—in the tri-state area of northwestern Iowa, eastern South Dakota, and southwestern Minnesota for almost 100 years, but as yet there is no unanimity of opinion concerning their interpretation. The controversy has focused on the following questions: (1) how many drift sheets are present?; (2) where do the drifts fit into the midcontinent Quaternary time scale?; and (3) what is the distribution of each of the major drift units? In this paper, I will review the previous work, present a summary of new stratigraphic studies, and propose a geologic history of the region based on my interpretation of these recent studies.

PREVIOUS WORK

Warren Upham was the first to map the glacial deposits of western and southwestern Minnesota. His initial reports (Winchell, 1880, 1881, 1884a) traced the deployment of ice lobes in western Minnesota (fig. VII-14) and outlined the history of development of Glacial Lake Agassiz. His work culminated in a series of reports and geologic maps of individual counties (Winchell, 1884a, 1888), and in a detailed monograph on Lake Agassiz (1896) that established a remarkably sound framework for future Quaternary studies in Minnesota and adjacent states. Upham recognized that the region had been glaciated more than once, and he worked out a sequence of retreatal moraines for the last glaciation in Minnesota that involved a general ice recession from southwestern Minnesota toward the northeast. Chamberlin (1883) incorporated much of Upham's work in glaciation in Minnesota that involved a general ice recession from southwestern Minnesota toward the northeast. Chamberlin (1883) incorporated much of Upham's work in Minnesota and Iowa into his own grand summary of the last major glaciation of the United States. Chamberlin (1894) assigned the surface deposits of northwestern Iowa and southwestern Minnesota to three different ice sheets, which he mapped as "Kansan," "East Iowan" (later called "Iowan"), and "East Wisconsin" (subsequently shortened to "Wisconsin"). The western boundary of the East Wisconsin drift on his map apparently is the outer edge of what is now called the "Bemis moraine." During the succeeding twenty years, the Iowa Geological Survey sponsored field work in northwestern Iowa that resulted in several revised interpretations of the Quaternary sequence. It is worthwhile to review them here because they reflect changing ideas on the midcontinent Quaternary sequence and exemplify the difficulties of mapping Quaternary deposits in certain areas.

Bain (1897) agreed with Chamberlin that Kansan, Iowan, and Wisconsin drift sheets were exposed in the tri-state area; however, he markedly revised the boundary between the Kansan and Iowan. In a later report, Bain (1898) focused attention on the drift that Chamberlin (1894) had assigned to the Iowan. After considering assigning it to the (1) Kansan, (2) Illinoian, (3) Iowan, and (4) extra-morainic Wisconsin, he reaffirmed its correlation with the Iowan drift of eastern Iowa. In a report on the geology of Carroll County, Iowa, Bain (1899) interpreted all the county outside the limits of the "Wisconsin Moraine" to consist of Kansan drift. He recognized two types: (1) "normal" Kansan, having a strongly weathered surface horizon, and (2) "abnormal" Kansan, lacking a weathered zone at the surface. He attributed the absence of a weathered zone to erosion before burial by loess. With respect to his previous work in northwestern Iowa, he wrote (Bain, 1899, p. 88): "No attempt can be made here to fix the age of the extra-morainic and fresh-looking drift in the counties to the north. The work of the present field season has shown that the reference of this drift to the Iowan is probably wrong."

Based on work by Wilder (1900), and especially by MacBride (1900, 1901), the Iowa Geological Survey published a "Preliminary Outline Map of the Drift Sheets of Iowa" (Calvin, 1901) that designated a considerable area in northwestern Iowa lying outside of the "Wisconsin Moraine" as "Wisconsin."

A map entitled "Map of the Surface Formations of Minnesota," by Frank Leverett and F. W. Sardeson, dated 1916, was included in a report (Leverett and Sardeson, 1919) published by the Minnesota Geological Survey. On that map (fig. VII-15) the drift of southwestern Minnesota is divided into two units: (1) "a drift older than the Wisconsin " ("Old Gray Drift"), and (2) moraines and till plains of Wisconsin age ("Young Gray Drift").

In 1909, J. E. Carman retraced the Wisconsin drift boundary on the western side of the Des Moines lobe north of Carroll County, in northwestern Iowa. On the basis of his field work during the next five years, Carman concluded that all of northwestern Iowa west of the "Wisconsin Moraine" was Kansan. The judgment is reflected in the "Map of Iowa Showing Drift Sheets," published as Plate LXV of the Iowa Geological Survey's Annual Report for 1913. Four years later, Carman (1917) reproduced this map to reaffirm his belief that no drift sheet younger than Kansan existed outside the limits of the Wisconsin moraine.

During the time of Carman's work in northwestern Iowa, many geologists were questioning the existence of an Iowan drift sheet in eastern Iowa, the type region for this drift. Undoubtedly, this fact influenced Carman's judgments con-

PLATE VI. Geological and Natural History Survey of Minnesota, 9th Report

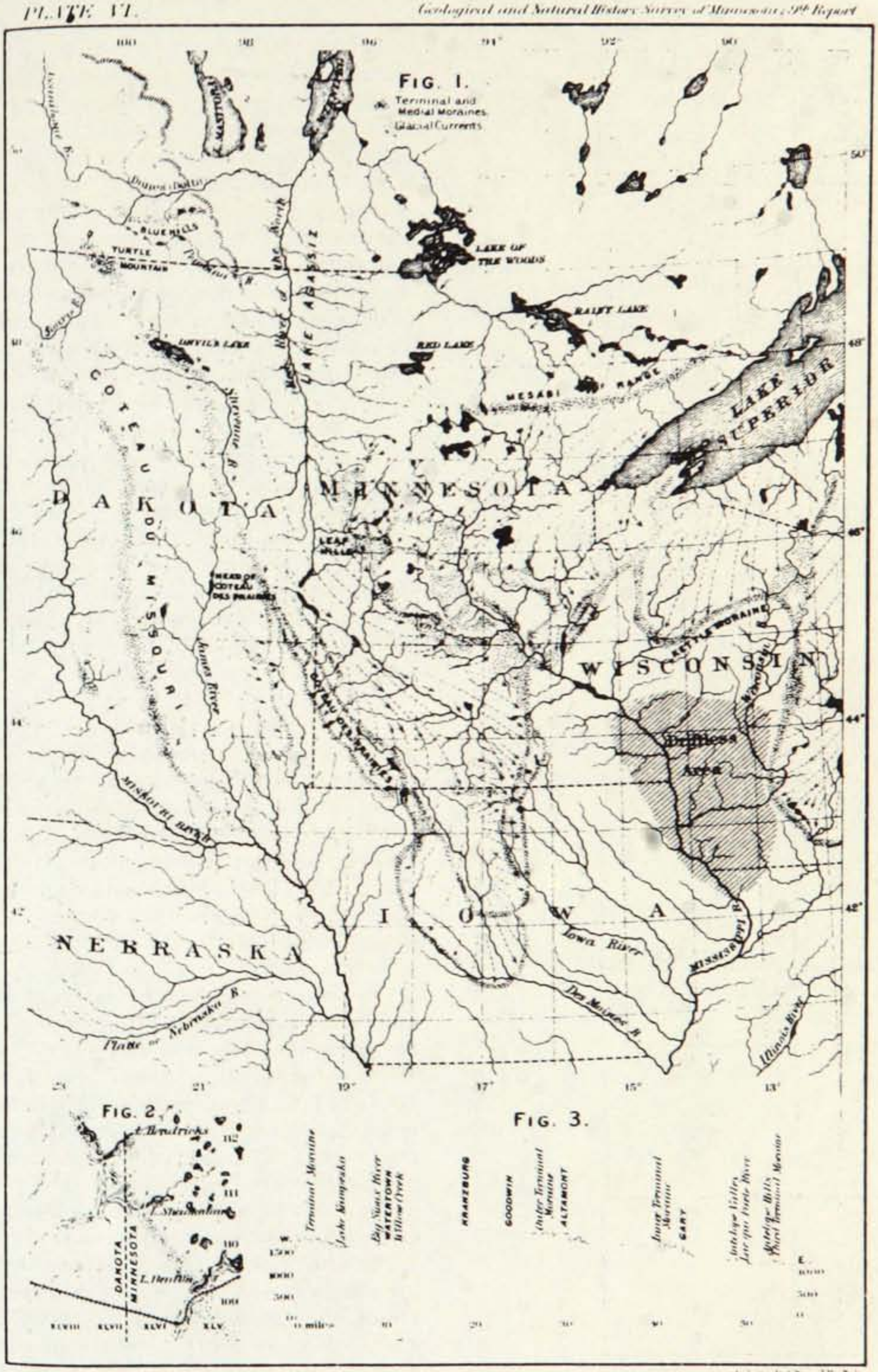


Figure VII-14. The course of the terminal moraines by Warren Upham (in Winchell, 1881).

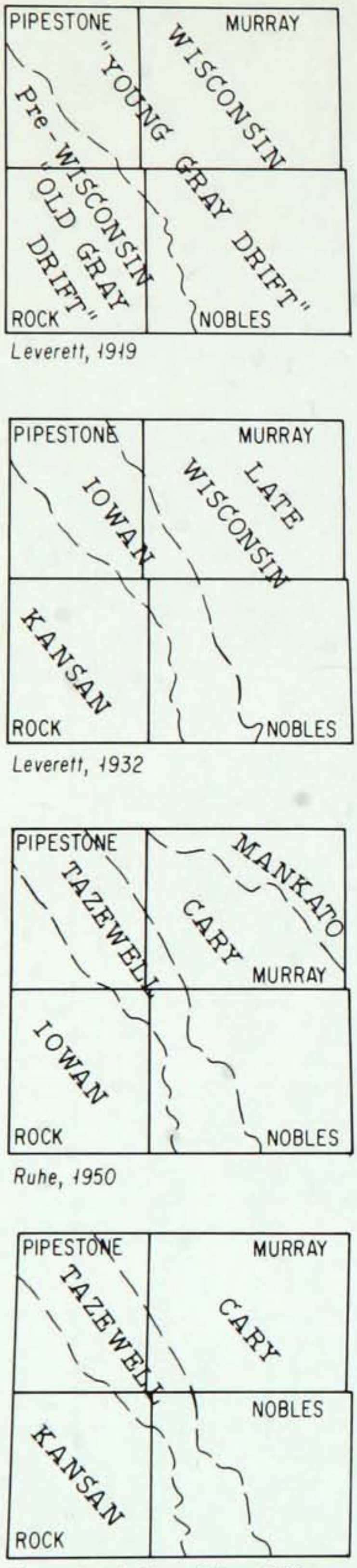


Figure VII-15. Various subdivisions of the surface drifts of southwestern Minnesota.

cerning the pattern of glaciation in the northwestern part of the state. In 1917, Alden and Leighton published a report that reaffirmed the existence of a post-Kansan-pre-Wisconsin drift sheet in northeastern Iowa; they stated that this drift sheet succeeded the Illinoian and preceded the Wisconsin. The report was accepted by most geologists and reestablished the Iowan as one of the stages of Pleistocene glaciation. This general acceptance inspired Leverett (1922, p. 101) to designate a strip of drift in southwestern Minnesota outside the Bemis moraine as "apparently somewhat older than the Wisconsin drift, and referred provisionally to the Iowan stage of glaciation." The extension of this drift into Iowa is the extra-morainic drift called "Kansan" by Carman (1917). In light of the reaffirmed Iowan in northeastern Iowa, Carman (1931) revised his earlier interpretation of the distribution of drift in northwestern Iowa to include recognition of an area of Iowan drift that he had previously called Kansan. Leverett's (1932, p. 29) final map of southwestern Minnesota delineated the distribution of three drifts: (Late) Wisconsin, Iowan, and Kansan (fig. VII-15).

By 1929, no one seriously doubted the existence of Iowan drift; however, geologists familiar with the area continued to debate its relationship to the other glacial stages. Leighton (1931) considered it the earliest substage of the Wisconsin, whereas Leverett (1939) favored its representing a late substage of the Illinoian. Leverett (1942) later conceded that the Iowan was an early Wisconsin drift. Kay and Graham (1943) concurred, and labeled the deposits of northwestern Iowa Wisconsin (Iowan) and Wisconsin (Mankato). The vision of the Iowan as a separate glacial stage faded away.

The boundaries established by Carman (1931) in northwestern Iowa were redefined by Smith and Riecken (1947) on the basis of topography and loess texture and thickness. Their interpretation expanded the area of surface exposure of Iowan drift at the expense of the Kansan. Mainly on the basis of drainage patterns, Ruhe (1950, unpub. Ph.D. dissert., Iowa Univ.) further divided the area into four substages of the Wisconsin: the Iowan, Tazewell, Cary, and Mankato (fig. VII-15). Flint (1955) extended this interpretation into South Dakota.

In succeeding years, the existence of an Iowan drift again came into question. After extensive study, Ruhe and his colleagues (Ruhe and others, 1957; Ruhe and others, 1968; Ruhe, 1969) concluded that the drift mapped as Iowan in Iowa was Kansan drift from which the Yarmouth and Sangamon Soils had been eroded. Ruhe (1969) dropped the Iowan as a substage of the Wisconsin, and now recognizes three till sheets—Kansan, and two Wisconsin-age drifts, Tazewell, and Cary—in northwestern Iowa and, by extension, in southwestern Minnesota (fig. VII-15).

In eastern South Dakota, recent work has resulted in the recognition of an Illinoian drift sheet (Tipton, 1959; Steece, 1959), and division of all later glacial deposits into Early Wisconsin and Late Wisconsin (Lemke and others, 1965).

REGIONAL GEOMORPHOLOGY

Southwestern Minnesota is dominated by two striking regional geomorphic features, the Minnesota River Valley, a wide and deep trench that served as the southern outlet for Glacial Lake Agassiz, and the east flank of the Coteau des Prairies, a broad regional topographic highland whose crest in Minnesota is an important drainage divide. Water drains from the southwest side into the Big Sioux River, and from the northeast side into the Minnesota and Des Moines Rivers.

In this part of the state, the Minnesota River Valley follows the southeastward-trending axis of a topographic trough that is more than 150 miles long and 100 miles wide. This drift-mantled topographic sag, known as the Minnesota River lowland, reflects a similar configuration of the underlying bedrock surface. It is the southern continuation of a more extensive bedrock low that was the dominant control on ice flow during the last glaciation of the area. Other regional low topographic trends in Minnesota served to channel the continental ice sheet into discrete lobes (fig. VII-16). The lobe that last flowed along the Minnesota River lowland has been called for many years the Des Moines lobe because the ice moved southward along what is now the Des Moines River valley to a terminus not far from the city of Des Moines, Iowa. An earlier tongue with a more easterly axis of flow is known as the Wadena lobe (Wright, 1962). Two distinct ice masses have been distinguished in northeastern Minnesota, the Rainy lobe, which advanced across the Rainy River and the Mesabi range, and the Superior lobe, which flowed from a large ice reservoir in the Lake Superior basin.

The valley of the Minnesota River is 1 to 3 miles wide and as much as 200 feet deep. It holds a remarkably straight southeasterly course for 150 miles between Ortonville and Mankato, where it turns sharply to the northeast. In many places the valley is floored by Precambrian, Paleozoic, and Mesozoic bedrock. Terrace segments, all of which are the result of late- and early postglacial erosional and depositional events closely related to the melting of the Des Moines lobe, are preserved at various heights above the modern floodplain.

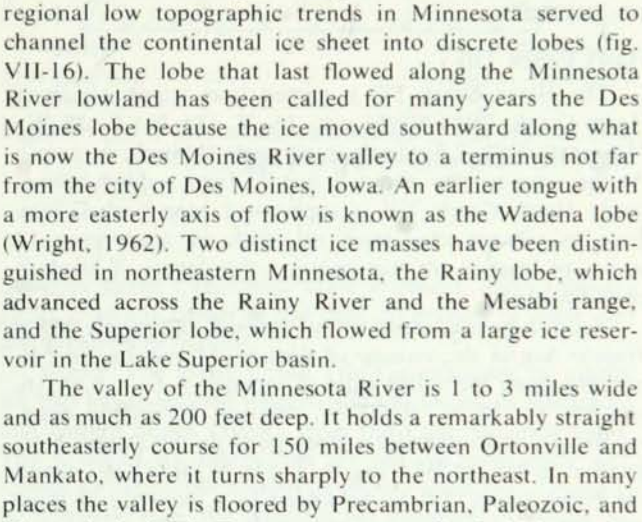


Figure VII-16. Glacial map of southern Minnesota and adjacent South Dakota (modified from Leverett, 1932).

The southwest flank of the Minnesota River lowland rises to an altitude of more than 2,000 feet, to crest as a "height of land" called the Bemis moraine on the eastern flank of the Coteau des Prairies. Glacial drift is as much as 500 feet thick on this part of the Coteau. The crest of the Coteau serves as a boundary between a regionally well-drained landscape to the southwest and a poorly-drained one to the northeast. The headward extension of streams from the Big Sioux River valley has resulted in the drainage of any depressions that might have been formed by glacial activity.

The northeast flank of the Coteau is marked by a series of regional steps whose treads are poorly-drained belts of hummocky terrain, with steeper and fairly well-drained risers. This regional terracing resulted mainly from differential glacial erosion and ice-marginal deposition during the advance and retreat of the last glacier. This ensemble of riser and tread, along with the general pitch of the entire landscape to the southeast, results in a drainage pattern that is composed of two major elements: low-gradient mainstreams that flow southeastward down the regional slope, and short, steep-gradient tributaries that flow northeastward down the flanks of the Coteau. Examples of stream piracy are abundant.

BEDROCK

Three major rock types underlie the glacial drift of southwestern Minnesota: high-grade metamorphic and igneous rocks of Early Precambrian age; (2) the Upper Precambrian Sioux Quartzite; and (3) poorly consolidated marine and continental Cretaceous shales and sandstones. These are described in other papers in this volume. Coarse-grained pink or white granitic gneiss probably is the major constituent of the Precambrian crystalline complex, with minor rock bodies of more mafic composition. At many places along the valley a soft kaolinitic regolith as much as 100 feet thick overlies the Precambrian bedrock. This thick clay-rich zone is part of a weathering profile developed during Cretaceous time (Parham, 1970). The Sioux Quartzite crops out extensively in Cottonwood, Pipestone, and Rock Counties, and in the vicinity of New Ulm in Nicollet County. Typically, it is a dark pink to dark red, coarse- to fine-grained clastic rock that breaks across the interlocking quartz grains. It stands as a regional topographic high because it is extremely resistant to erosion.

Detailed knowledge of the Cretaceous strata is lacking because exposures are poor and subsurface information is scarce. Generally, these beds consist of poorly consolidated quartz sand, lignitic clay, and soft dark-gray shale. Within the Cretaceous sediments a distinctive interval that overlies the regolith developed on the Precambrian crystalline complex is a grayish-white to brown, hard pisolitic clay.

GENERAL DRIFT TYPES

An important early observation on the glacial deposits of Minnesota was the recognition of two basic types of drift: a yellow to gray, silty, calcareous till and a red, sandy, carbonate-free till (Winchell, 1877, 1880). The basic difference in these two types results from their derivation as glacially eroded clastics from geographically separate and petrologically diverse areas in Minnesota and adjacent localities. Since those early observations, more subtle lithologic subdivisions of the drift have been made as the areal geology of the region has become better known (Arneman and Wright, 1959; Schneider, 1961).

The red sandy till contains a distinctive suite of rock fragments that is characteristic of the Precambrian bedrock of a large area around Lake Superior. Although complicated in detail, this rock province contains the following diagnostic rock types: black, fine-grained basalt; purple or red felsite; gray gabbro; red granophyre; dark gray or black diabase; and red to pink shale and pink arkosic sandstone. Drift of this type is associated with activity of the Superior lobe.

The yellow or gray calcareous till had its source in the vicinity of the Winnipeg lowland, where a belt of Paleozoic sedimentary rocks, mostly fine-grained limestone and dolomite, occurs in a zone as much as 140 miles wide and more than 400 miles long that trends northwestward across southern Manitoba and extends southeastward into Minnesota for a short distance. The presence or absence of fragments of a hard, brittle, noncalcareous, light to dark gray or greenish-gray siliceous shale makes it possible to separate this drift into two types. The most likely source for this shale is the Upper Cretaceous Pierre Shale, which covers an extensive area in eastern North Dakota, eastern South Dakota, and southern Manitoba. Calcareous gray drift that does not contain this shale originated in the lowland east of the Pembina escarpment. The ice lobe that deposited it is called the Wadena lobe (Wright, 1962). Shale-rich calcareous till represents deposition from ice that flowed from a more westerly source. It comprises the surface deposits over much of southwestern Minnesota, and its deposition is ascribed to the advance and retreat of the Des Moines lobe.

PLEISTOCENE STRATIGRAPHY

The lithology of the drift deposits in southwestern Minnesota was discussed in very general terms by both Upham (in Winchell, 1880, 1881) and Leverett (1932). Both recognized that the calcareous gray surface drifts contained rocks from the Winnipeg lowland, and they therefore ascribed the drift sheets to glaciers that moved into the area from the northwest. Upham (in Winchell, 1880, p. 115) recognized wood and peat buried within the drift and was the first to report this evidence of multiple glaciation.

The glacial drift in southwestern Minnesota is stratigraphically complex. Superposed tills of different lithology that are separated by paleosols, accretion gleys, striated boulder pavements, and forest beds indicate at least four distinct glacial episodes. So far as is known, however, only two of the drifts have significant areal distribution. The others are exposed only in deep cuts, especially along the sides of the Minnesota River Valley and its tributaries.

Major Till Units

Three different glacial tills comprise the bulk of Quaternary sediments exposed along Big Stone Lake and the Minnesota River Valley between Ortonville and Granite Falls. From oldest to youngest, these are: (1) a lower, pink to reddish-brown, sandy clay loam till, containing stones of Lake Superior provenance and a small percentage of limestone, herein called the "Hawk Creek Till;" (2) a middle, yellow to yellowish-brown, calcareous loamy till, containing mostly limestone, dolomite, and granitic pebbles, herein called the "Granite Falls Till;" and (3) an upper light olive-brown, calcareous clay loam till, containing pebbles predominantly of siliceous shale, limestone, dolomite and granitic rocks, herein called the "New Ulm Till." In many places, the upper two tills are separated by a planed and striated boulder pavement that generally is one-stone thick. In other exposures farther down the valley, the reddish-brown sandy till is superposed on a still older calcareous gray drift.

Hawk Creek Till

The Hawk Creek Till is named from exposures along Hawk Creek, SW¼NW¼ sec. 16, T. 116 N., R. 38 W., Minnesota Falls 7.5-minute quadrangle (Matsch, in prep.). At the type section about 5 feet of light reddish-brown, sandy Hawk Creek Till is superposed atop a shale-free, gray, calcareous, clay loam till, with an intervening 12 inches of leached silt that contains thin layers of blackish plant fragments.

Reddish-brown sandy till is exposed extensively along Watson Sag in the vicinity of Watson, Minnesota, and especially along the valley sides in sections 5, 6, 7, 8, 9, 16, and 17, T. 118 N., R. 41 W., Watson 7.5-minute quadrangle. An excavation for a stock pond on the Dennis Norby farm in NE¼NW¼ sec. 8, T. 118 N., R. 41 W., shows the Hawk Creek Till as the lowest in a sequence of three tills. Other occurrences of reddish sandy till are reported along the South Dakota side of Big Stone Lake (Robert Rutford, 1968, oral comm.). In his report on Big Stone County, Upham (in Winchell, 1884a, p. 628) noted the occurrence of this red till in deep wells all the way to the foot of the Coteau des Prairies in Grant County, South Dakota. Southeast of Watson, the Hawk Creek Till is exposed only rarely. Thus, its occurrence at the type section east of Granite Falls may be an outlier of an extensively eroded drift sheet.

Half a mile south of Franklin, a reddish-brown sandy till is exposed along the east side of the road in the NE¼ SE¼ sec. 11, T. 112 N., R. 34 W., Morton 15-minute quadrangle. In addition, sandy till is exposed at the base of a shallow borrow pit near the intersection of two county roads, SW¼NE¼ sec. 34, T. 112 N., R. 26 W., Le Sueur 7.5-minute quadrangle.

Characteristics. The Hawk Creek Till is distinguished from other Quaternary deposits in southwestern Minnesota by its distinctive color, texture, and lithology. The color ranges from pink (5 YR 7/3) on dry, oxidized exposures, to reddish-brown (5 YR 4/3) on wet exposures of unoxidized till. The textural designation of the Hawk Creek Till is sandy clay loam (fig. VII-17). The till is composed of a distinctive suite of rock fragments that includes a large percentage of rock types from the Lake Superior region, such as red felsite, pink sandstone, gabbro, and even sparse Lake

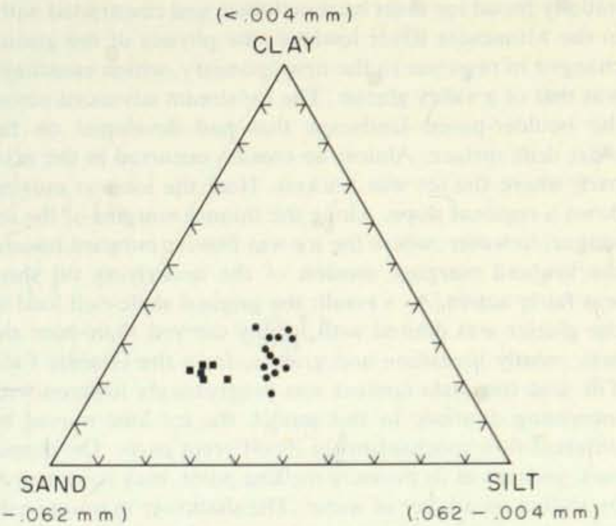


Figure VII-17. Grain size distribution in the Hawk Creek Till (■) and the Granite Falls Till (●).

Superior agate. Cretaceous shale is absent, and carbonate rocks generally comprise less than 20 percent of the 1 mm and greater size fraction.

Origin. The Hawk Creek Till was deposited by glacial ice that advanced into western and southwestern Minnesota and eastern South Dakota from the Lake Superior basin. The limits of its distribution are not yet sufficiently well known to define the margin of the ice sheet that deposited it; however, even the conservative geographical limits inferred from present knowledge lead to the conclusion that this drift represents a glaciation of major importance in terms of climatic change in Minnesota and adjacent states.

Granite Falls Till

The most common Quaternary stratigraphic succession in the deep cuts along the Minnesota River Valley and its tributaries consists of two calcareous tills separated by a variable thickness of outwash or a boulder pavement, or marked by a contact between unoxidized till over oxidized till. The lower unit is named the "Granite Falls Till," for numerous exposures in the vicinity of Granite Falls (Matsch, in prep.). At its type section, a roadcut, cen. sec. 28, T. 116 N., R. 39 W., Granite Falls 7.5-minute quadrangle, a variable thickness of as much as 20 feet of Granite Falls Till is separated from the overlying New Ulm Till by a striated and faceted boulder pavement that is one-stone thick. There it contains sparse shale in the coarse-sand fraction (less than 5 percent), and thus contrasts strongly with the overlying shale-rich (over 50 percent in the coarse-sand fraction) New Ulm Till. In all exposures investigated so far the till is calcareous to its contact with the overlying drift. At some localities, limestone and dolomite are constituents of the boulder pavement. Individual carbonate fragments appear fresh and unweathered throughout the entire body of drift. The only evidence of weathering is oxidized till and the presence of iron and manganese oxides along joints, which commonly are closely spaced, giving the till a crumbly aspect when it is spaded. Commonly, sand and gravel lenses are enclosed in the till.

The Granite Falls Till is exposed extensively along both shores of Big Stone Lake, and is fairly continuous along the valley sides of the Minnesota River to Mankato and beyond. On the east bank of Hawk Creek, half a mile southwest of the type locality for the Hawk Creek Till, in the SE¼SW¼ sec. 16, T. 116 N., R. 38 W., Minnesota Falls 7.5-minute quadrangle, the Granite Falls Till overlies a dark brown, blocky, clay-rich sediment that is interpreted to be an accretion gley. Beneath the gley lies the Hawk Creek Till. At this locality, the till is darker and more clay-rich than at the type locality, presumably the result of contamination from the underlying clay zone. The clay has a blocky structure and contains mollusk fragments. The sand fraction (6 percent) consists mainly of rounded quartz grains; fragments of red felsite and pink sandstone, however, are moderately abundant.

Characteristics. The most distinctive characteristic of the Granite Falls Till is its stone content; shale is absent or present in small amounts (1 to 5 percent). The two major types of rock fragments greater than 1 mm in diameter are carbonate and granitic rocks. Generally, these comprise 80 percent or more of this size grade and are present in nearly equal amounts. It is not uncommon to find a few rock fragments that were derived from the Lake Superior region. Presumably, these were incorporated from the underlying red Hawk Creek Till.

Texturally, the till ranges from sandy loam to loam to clay loam (fig. VII-17). In many exposures, masses of silt, sand, and gravel are part of the main body of till. Typically, the dry oxidized till is pale yellow (2.5 Y 8/4) or yellow (2.5 Y 8/6) when dry, and light yellowish brown (2.5 Y 6/4) or light olive brown (2.5 Y 5/4) when wet. Unoxidized till is rarely exposed, but typically is gray (10 YR 6/1) when dry and dark gray (10 YR 4/1) when wet.

Origin. Because it is relatively free of Cretaceous shale fragments, the Granite Falls Till must have been deposited by an ice sheet that bypassed the broad region underlain by that rock type in the eastern Dakotas. Lithologically, it is similar to the highly calcareous till in the Wadena region, 130 miles north of Granite Falls, and is presently correlated with that till sheet. The great areal extent and thickness of this till indicates that its deposition was a major glacial event in mid-America.

New Ulm Till

Surface deposits over most of the region mainly consist of till, outwash, and lake sediments associated with the last glaciation of southwestern Minnesota. A large proportion of these sediments consists of distinctive yellow to olive brown (oxidized) or dark gray (unoxidized) calcareous till that contains abundant shale fragments, called the "New Ulm Till" for excellent exposures in and near that city on the Minnesota River (Matsch, in prep.). The type section is a roadcut along Minnesota Hwy. 68 extending from a point 1.75 miles southeast of New Ulm, at the SE. cor. NE¼ sec. 4, T. 109 N., R. 30 W., New Ulm 7.5-minute quadrangle, for a distance of approximately 2,000 feet to the south bluff of the Cottonwood River. The base of the till, exposed in an isolated hill on the west side of Hwy. 68 about 500 feet south of the bridge over the Cottonwood River, lies unconformably on several older Quaternary deposits. Wood collected from the top 6 inches of the underlying black loamy till has a radiocarbon age greater than 39,900 years BP (I-4931). Inasmuch as the contact between the black loamy till and the overlying New Ulm Till is an erosion surface, this date does not mark the time of deposition of the New Ulm Till. The basal contact over most of the area along the Minnesota River Valley is a boulder pavement developed on the underlying Granite Falls Till.

Exposures of the New Ulm Till are numerous throughout the area and only a few need be pointed out here to serve as alternate reference sections. About 100 feet of the till is exposed just north of the intersection of county Hwys. 14 and 15, in the NE¼ sec. 21, T. 110 N., R. 30 W., about one mile east of New Ulm. Here, the upper 20 feet is oxidized to a pale yellow, and the unoxidized base rests on a bouldery gravel. Excellent exposures also can be seen along the entire length of the Minnesota River Valley, in drainage ditches and tributary streams, as well as in roadcuts. A roadcut in NE¼SW¼ sec. 19, T. 19 N., R. 36 W. clearly shows the relationship of this till to the Granite Falls Till.

Lithology. The New Ulm Till is characterized by three major rock types in the coarse-sand fraction, siliceous shale, carbonates (mainly fine-grained dolomite and limestone), and granitic rocks, including igneous quartz. The percentage distribution of shale in the index grade size for the surface till is shown in Figure VII-18. The data show that there is a regular decrease in shale content on either side of the topographic axis of the Minnesota River lowland.

The most reasonable explanation for the systematic decrease in shale content of the New Ulm Till relates to the regimen of the glacier that deposited it. This explanation assumes that the ice-eroded sediment had a fairly constant and high content of shale after the glacier had crossed the broad area covered by siliceous Cretaceous shale. As the initially broad ice sheet became lobate and constricted within the Minnesota River lowland, the physics of the glacier changed in response to the new geometry, which essentially was that of a valley glacier. The ice stream advanced across the boulder-paved landscape that had developed on the older drift surface. Almost no erosion occurred in the axial part, where the ice was thickest. Here the ice was moving down a regional slope. Along the thinner margins of the ice tongue, however, where the ice was flowing outward toward the lowland margins, erosion of the underlying till sheet was fairly active. As a result, the original shale-rich load of the glacier was diluted with locally derived shale-poor debris, mostly limestone and granite, from the Granite Falls Till, and the shale content was progressively lowered with increasing dilution. In this model, the ice lobe moved by different flow mechanisms in its different parts. The deeper part, perhaps at its pressure melting point, may have moved by sliding on a layer of water. The shallower margins, constricted and also colder, may have moved more by shear.

Size analyses for a large number of samples of the New Ulm Till are presented diagrammatically in Figure VII-19.

Origin. The New Ulm Till constitutes the largest volume of the surface deposit called "Young Gray Drift" by Leverett and Sardeson (1919) and Leverett (1932). Long ago, Upham (1896) ascribed this drift to the activity of an ice lobe, now called the Des Moines lobe, that flowed south and southeast along the axis of the Red River-Minnesota River lowland. Exposures of this till in the vicinity of Mankato, Minnesota, were the basis for the recognition of the Mankato Substage, a controversial subdivision of the Wisconsin Glacial Stage (Leighton, 1933, 1960; Zumberge and Wright, 1956; Wright and Rubin, 1956; Wright, 1964; Frye and others, 1968; Ruhe, 1969).

"Extra-morainic" Shale-bearing Till

A broad belt of loess-covered till lying just outside the Bemis moraine in Lincoln, Pipestone, Murray, and Nobles Counties is similar lithologically to the till that comprises

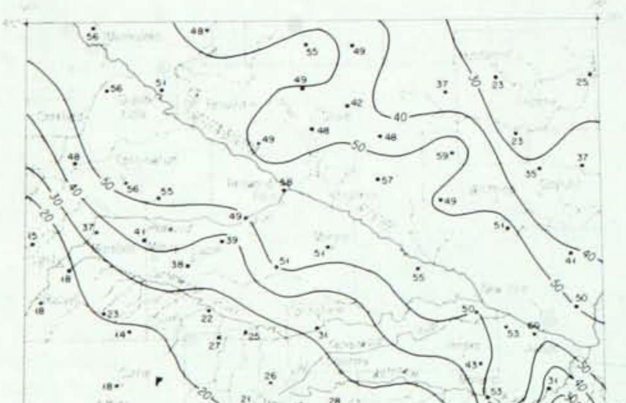


Figure VII-18. Distribution of siliceous Cretaceous shale, in percent, in the sand size grade 2 mm to 1 mm, New Ulm Till, southwestern Minnesota.

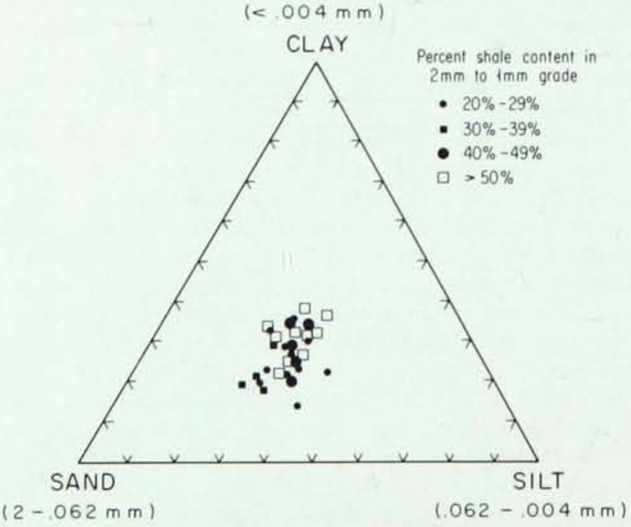


Figure VII-19. Grain size distribution, New Ulm Till, southwestern Minnesota.

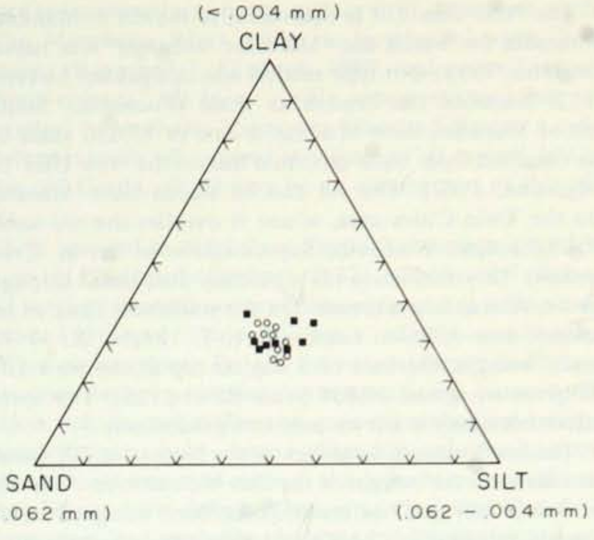


Figure VII-20. Grain size distribution in the New Ulm Till comprising the Bemis moraine (■) and "extra-morainic shale-bearing till" (●).

the moraine. The till contains significant amounts of siliceous Cretaceous shale (15-20 percent), but tends to be slightly more silty than till in the moraine (fig. VII-20). It differs markedly from the clayey, shale-poor "older drift" that lies farther southwest. Leverett and Sardeson (1919) at first included the extra-morainic shale-bearing till within the boundary of "Young Gray Drift," but later Leverett (1932) decided that it represented a stage of glacial activity "older than the Wisconsin drift," but younger than the Illinoian, called the Iowan. Ruhe (1950, *op. cit.*, 1969) interpreted this drift as representing an early substage of the Wisconsin, the Tazewell. Similar drift in eastern South Dakota is mapped as "Early Wisconsin."

The lithology of this drift belt indicates that it was derived from the same bedrock terrane as the till in the adjacent Bemis moraine, a feature that has long been interpreted as marking the southwest edge of the late-Wisconsin Des Moines lobe. In addition to its extra-morainic position, the presence on its surface of a pavement of wind-cut stones and a loess cover, neither of which extends across the moraine, proves that it is older than the Bemis moraine. Whether or not this drift represents a regionally significant advance and retreat of the Des Moines lobe cannot be ascertained until more stratigraphic information is available. Currently, it is interpreted as representing an extra-morainic position of the Des Moines lobe during its general main-Wisconsin activity in the region.

Other Quaternary Deposits

Several deep cuts along the valley sides of the Minnesota River in the vicinity of Redwood Falls and Morton expose a complex succession of Quaternary deposits, including tills, outwash, and clays and silts that are not laterally persistent. The stratigraphy of two exposures is presented in Figure VII-21. The North Redwood section is located in

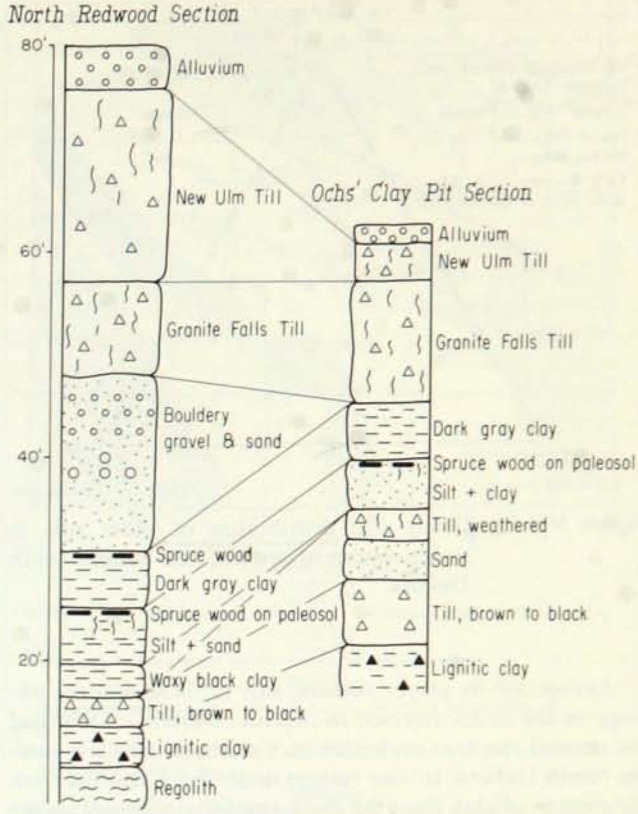


Figure VII-21. Stratigraphic sections at North Redwood and Morton, Minnesota and their inferred correlation.

the SW¼ sec. 29, T. 113 N., R. 35 W., and the Ochs' clay pit section is in the SE¼SW¼ sec. 35, T. 113 N., R. 35 W., both in the Redwood Falls 15-minute quadrangle.

Spruce wood collected from the wood zone in the Ochs' clay pit section has an age of 34,000 years BP (GX-1309); another sample was dated at greater than 39,900 radiocarbon years (I-4932). The inconsistency of these ages has not been resolved.

Sediments that lie below the Granite Falls Till in the clay pit are discontinuous, but even though they may be unimportant areally, their presence attests to several glacial depositional events prior to the glaciation represented by the Granite Falls Till. The variety of nonglacial sediments separating these tills also indicates more than one warm climatic interlude between the deposition of the till sheets.

All the tills found below wood horizons contain limestone and granite fragments, but have sparse siliceous Cretaceous shale. Textures range from sandy loam to clay. Clay-rich till of this same general lithology is widely distributed in Rock and Pipestone Counties, and in parts of eastern South Dakota. Characteristically, the coarse-sand fraction predominantly is composed of granitic rock fragments and quartz (50 to 70 percent), and carbonates (15 to 35 percent). Siliceous shale generally is less than 5 percent of this fraction. Figure VII-22 summarizes the textures of older tills collected over a wide area in southwestern Minnesota and eastern South Dakota.

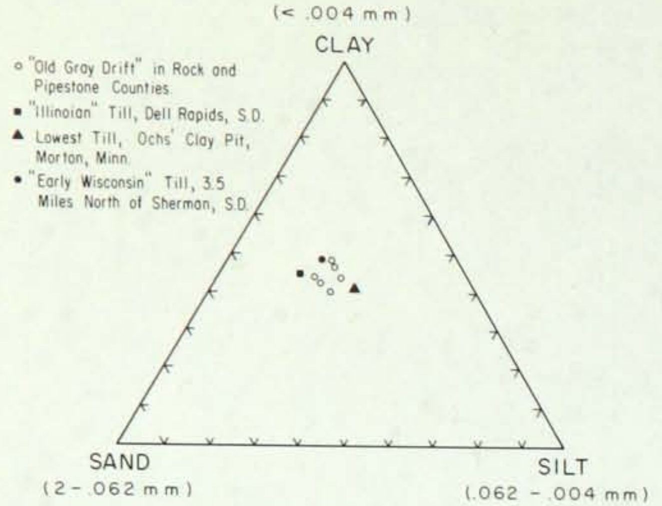


Figure VII-22. Grain size distribution of older tills in southwestern Minnesota and eastern South Dakota.

Except for its clayey texture, this till is similar in lithology to the drifts ascribed to deposition from ice that had not crossed the area underlain by Cretaceous shale in eastern North Dakota. Its fine texture might have resulted from the erosion of clay from the thick regolith developed on the Precambrian crystalline complex, and the high granitic content might reflect grus incorporated from this same source.

Clay-rich deposits that lie between tills at some places in southwestern Minnesota have the characteristics of accretion gleys, as defined by Frye and others (1960). Particularly good exposures are the fine-textured deposits already described from along Hawk Creek and from the vicinity of Morton (Ochs' clay pit) and North Redwood. All these sedimentary units are calcareous, and some are distinctly laminated. Their coarser size fractions contain minerals such as feldspar that are low in the weathering stability series. Mollusk fragments, plant detritus, and pollen grains also have been observed in them. These deposits between tills prove multiple glaciation, but they do not necessarily represent long climatic episodes in the Pleistocene sequence. Deeply weathered horizons developed on till corresponding to gumbotil in other areas have not been observed in the region.

Regional Correlations and Ages

Because of a lack of published detailed lithologic data for formations of Quaternary age in South Dakota and Iowa, the New Ulm Till is the only formation that now can be correlated with certainty over a wide area. Long ago, Upham (in Winchell, 1881) traced this drift sheet from Minnesota into South Dakota and Iowa. In Iowa, it is called the "Cary glacial drift," and its base has been dated at about 14,000 radiocarbon years BP (Ruhe, 1969). In an earlier interpretation of the morainic trends, Ruhe (1952) ascribed this drift to two distinct advances of the Des Moines lobe, and he separated the till sheet into the "Cary drift" and the "Mankato drift."

The New Ulm Till is equivalent to the till at Mankato, Minnesota for which the "Mankato Substage" was named (Leighton, 1933). No type section was designated. Leverett (1932) classified this deposit as "Late Wisconsin." Southwest of Mankato, near Madelia, a date of 12,650 years BP was obtained from basal peat that lies on the New Ulm Till (Jelgersma, 1962). The till can be traced from Mankato into the Twin Cities area, where it overlies the red sandy till of the Late Wisconsin Superior lobe in the St. Croix moraine. This distinctive till is patchily distributed throughout the Alexandria moraine. On the southwest flank of the moraine near Monson Lake, sec. 1, T. 121 N., R. 37 W., spruce wood at the base of a bog on top of the New Ulm Till gives an age of 10,850 years BP (I-5125). The northeastern boundary is known only in a general way.

The southwestern boundary of the New Ulm Till should coincide with the margin of the Des Moines lobe. Although the Bemis moraine has traditionally been accepted as the terminal moraine of the Des Moines lobe, the possibility exists that an extra-morainic belt of loamy shale-rich till is correlative with the New Ulm Till. That extra-morainic till is contiguous with the Tazewell drift of Iowa and dated there as 20,000 radiocarbon years old (Ruhe, 1969).

Direct tracing of the Granite Falls Till south and west of the Minnesota River Valley is not possible with available data. However, till of similar lithology is exposed 40 miles southwest of Granite Falls in deep cuts along the Redwood River near Lynd. It has been traced westward from Big Stone Lake to the foot of the Coteau des Prairies in northeastern South Dakota (Robert H. Rutford, 1970, oral comm.), and may comprise an important volume of the thick Quaternary sequence there. It probably is an extension of the till that comprises drumlins in the Wadena area. If so, this correlation would extend the range of glacial activity of the Wadena lobe far south and west of the limits set by Wright (1962, 1964).

The Granite Falls Till can be traced northeastward from Mankato, where it is the "middle till" of Zumberge and Wright (1956), almost to St. Paul. Gelineau (1959, unpub. M.S. thesis, Univ. Minn.) identified a calcareous till in Dakota County that is characterized by sandy texture and an absence of siliceous Cretaceous shale fragments. In this area, it lies beneath red sandy till of Late Wisconsin age deposited by the Superior lobe. On the basis of lithology, he tentatively ascribed the deposition of this drift to the Wadena lobe. Possibly, the till described by Gelineau is an easternmost exposure of the Granite Falls Till. The so-called silt-capped Cary drift that makes up the Bemis moraine in southeastern Minnesota also is lithologically similar to the Granite Falls Till.

The reddish-brown, sandy Hawk Creek Till in western Minnesota beneath the Granite Falls Till represents an advance of the Superior lobe all the way to the South Dakota border. Discontinuous exposures of this till occur along the Minnesota River Valley from Ortonville to Le Sueur, just north of Mankato, indicating an impressive breadth for this lobe. Reddish-brown till has been noted as far southeast as Owatonna (Joseph Cummins, 1969, oral comm.). Its subsurface presence southwest of the Minnesota River is indicated by scattered concentrations of stones of Lake Superior aspect at various places in the surface drift. Its extent north of the Minnesota River Valley is largely unknown. Soil scientists (Raymond T. Diedrick, 1969, oral comm.) report "salmon-colored" till from a wide area north and east of the valley. In southeast Minnesota (Dakota County), a reddish-brown sandy till mapped as "Illinoian" (Leverett, 1932; Ruhe and Gould, 1954) may be an eastern part of this till sheet.

Tills mapped as "Old Gray Drift" (Leverett and Sardeson, 1919), "Kansan" (Leverett, 1932), and "Iowan" (Ruhe, 1950, *op. cit.*) in southwestern Minnesota, and as "Illinoian" (Tipton, 1959) and "Early Wisconsin" (Tipton and Steece, 1965) in eastern South Dakota have the same general lithologic characteristics—a clay-rich texture and a coarse sand fraction rich in carbonates and granitic rocks, and nearly lacking siliceous shale—as the older tills near the base of the Quaternary sequence exposed along the Minnesota River Valley. These drifts definitely are older than 40,000 radiocarbon years, and probably represent pre-Wisconsin glacial activity.

SUMMARY OF AREAL GEOLOGY

Till, outwash, glacial lake sediments, and loess underlie most of the land surface of southwestern Minnesota. Distribution of most of the glacial sediments is closely related to events associated with the advance and retreat of the Des Moines lobe along the axis of the Minnesota River lowland. Till in the form of end moraines and ground moraines covers the greater part of the area. The stagnating and melting of the Des Moines lobe, which was essentially a valley glacier along this stretch in Minnesota, was accompanied in many places by temporary ponding that resulted in the accumulation of fine-grained lake sediments. The activity of slope processes on a landscape of stagnant ice, coupled with various sedimentary processes in environments associated with disintegrating ice resulted in a variety of features, such as crevasse-fillings, collapsed alluvial fans, and perched lakes. Runoff from the melting Des Moines lobe and discharge from glacial lakes established a network of meltwater channels and lake outlets that have a variety of alluvial sediments associated with them. Lake Agassiz, the grandest glacial lake of all, spilled enough water southward to erode the deep Minnesota River Valley. Wind-swept alluvial flats and till plains bordering the Missouri and Big Sioux Rivers produced silts that blanket almost all the landscape outside of the Bemis moraine. This great variety of Quaternary deposits closely controls man's use of the land surface.

Glacial Tills

Throughout its course in Minnesota (see fig. VII-23), the Des Moines lobe deposited a calcareous, gray (unoxidized) till characterized lithologically by the presence of siliceous shale. This loamy, shale-rich till comprises the surface or near-surface glacial sediments over thousands of square miles in southwestern Minnesota. At places, the till is piled into belts of morainic topography, but more commonly it constitutes gently rolling, poorly drained till plains. The till is thin in most places along the broad flat axis of the Minnesota River lowland, and contains abundant shale.

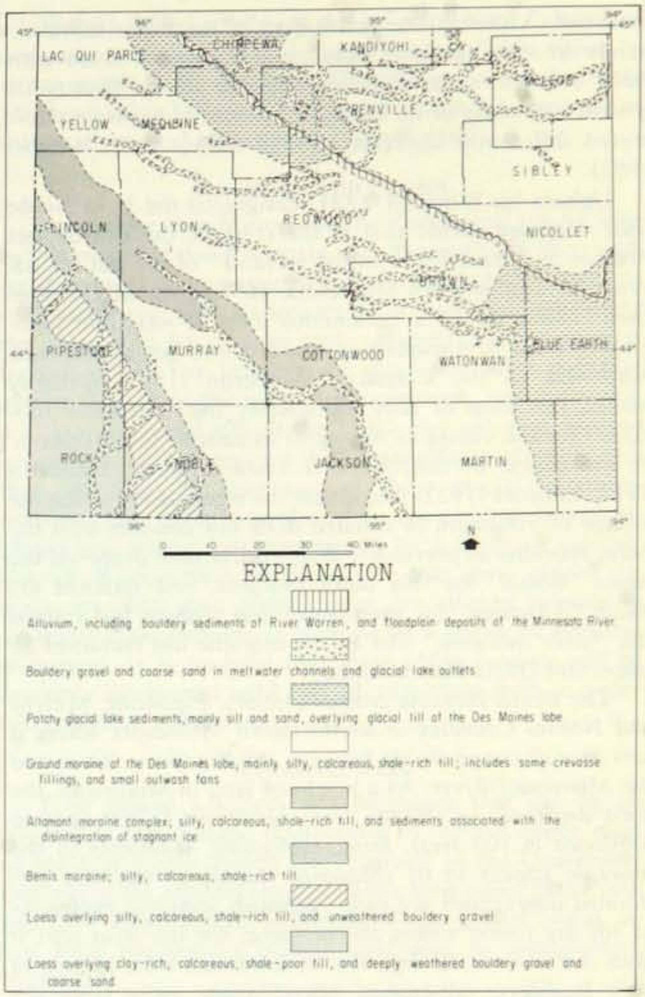


Figure VII-23. Areal geology of southwestern Minnesota.

Away from the axis, the till thickens, and the shale content decreases. On the southwest, a thickness of several hundred feet is attained in the Bemis moraine.

Southwest of the Bemis moraine, the landscape changes to a loess-blanketed, well drained surface that slopes regionally to the central cleft in the Coteau des Prairies that is the Big Sioux River valley. Within this region, shale-rich till continues beneath loess and outwash deposits for some miles beyond the moraine, after which a clay-rich till relatively free of shale is encountered. The boundary between these two tills approximates a division placed long ago between "Old Gray Drift" and "Young Gray Drift" (Leverett and Sardeson, 1919).

Moraines

Upham (in Winchell, 1880, 1881) was the first to interpret hummocky linear belts of hills in southwestern Minnesota as depositional landforms associated with glacier margins. He worked out a sequence of retreatal moraines for the Wisconsin Stage in Minnesota that indicated a general ice recession from southwestern Minnesota toward the northeast. Although the positioning of the major morainic trends on areal geologic maps has not changed much with more detailed work, their significance to the history of growth and decay of ice sheets in Minnesota has been interpreted differently (Leverett, 1932; Ruhe, 1952; Wright, 1962).

Upham (in Winchell, 1881) designated the 1- to 3-mile wide drainage divide that is the crest of the Coteau des Prairies as "the outer terminal moraine" of the last ice advance in southwestern Minnesota. He had earlier (in Winchell, 1880) traced a continuous loop of morainic topography from Minneapolis southward into Iowa, and then northwest to the Coteau. Chamberlin (1883) formally named this loop of hilly topography the "Altamont moraine" for the village of Altamont in eastern South Dakota. In the course of field work in eastern South Dakota in 1912, Leverett (1922) found that the moraine on which the village of Altamont is situated does not connect with the outer moraine as previously mapped. He then proposed the name "Bemis" for this outer moraine, and retained the name "Altamont" for the feature that Upham had termed the "inner moraine." The Bemis moraine has remained an important Quaternary boundary.

The Bemis moraine crosses Lincoln, Pipestone, Murray and Nobles Counties in southwestern Minnesota where it acts as a drainage divide between the Big Sioux River and the Minnesota River. As a height of land in Minnesota, the crest decreases in altitude from northwest (1,950 feet) to southeast (1,700 feet). Stratigraphically, sediments in the moraine appear to be composed mainly of one till. Undrained depressions are rare. Although scattered exposures of silt are found within the moraine, for the most part it does not have a silt cap. On its southwest side, aprons of very bouldery, silt-capped gravel grade away from the moraine.

The moraine is breached in several places by impressive gorges which served as outlets for ponded waters between the moraine and receding glacier ice. The outlets were abandoned as the ice melted away from the Bemis moraine, and drainage became established down the slope of land toward the axis of the Minnesota River Valley. These capacious valleys, at places, have lakes at their heads, as for example Lakes Benton, Hendricks, and Shokatan, which resulted from damming by fans as the abandoned outlets were filled with silt.

On its northeast side, the Bemis moraine grades into a poorly drained area of low relief that is underlain by till, fine sand and gravel, and scattered lake silts. At places, this belt of ground moraine grades into a broad stagnant-ice complex characterized by flat-topped hills, ridges, depressions, and meltwater channels. Many of the round, flat-topped hills are underlain by lake silts and cross-bedded sand. These flat-topped hills were the floors of ice-walled lakes that formed within the thin stagnant margin of the Des Moines lobe as it melted back into the Minnesota River lowland.

Broad belts of hummocky terrain elsewhere on the Coteau des Prairies have been designated the "Altamont" or "Altamont-Gary moraine" (Leverett, 1922, 1932). The rugged relief and poor drainage of this moraine complex contrast strongly with the well drained Bemis moraine. All of this deadice moraine on the northeastern flank of the Coteau is here designated the Altamont moraine complex.

In southwestern Minnesota the Altamont moraine complex is 5 to 12 miles wide. It parallels closely the southeastward trend of the Bemis moraine until the vicinity of Lake Shetek, where it narrows and takes a more easterly course. Near Windom the complex turns south to cross into Iowa in the vicinity of Spirit Lake. The moraine was formed along the receding margin of the Des Moines lobe.

Some features previously mapped as end moraines are no longer considered of that origin. The Antelope moraine (Upham, 1896; Leverett, 1932), composed mainly of sorted sediments, is a large crevasse filling, whereas the Marshall moraine (Leverett, 1932) is a trend of higher relief resulting from erosion along a meltwater channel system.

Outwash Deposits

Sorted sediments that were deposited by running water are widely distributed in the region, and most of them are related to the stagnation and retreat of the Des Moines lobe. These deposits consist mainly of meltwater channels, crevasse fillings, terrace gravels, and outwash fans.

A system of meltwater channels that apparently flowed along successive margins of the receding ice sheet vein the entire region (fig. VII-23). Major rivers, such as the Yellow Medicine and Cottonwood, follow the courses set by these earlier ice-melt streams. Some branches of this channel network head into flat terrain underlain by silts that must have been glacial lake bottoms. Although the channel sediments are variable in texture, the deposits typically consist of well-sorted coarse sand interlayered with poorly-sorted bouldery gravel.

Another type of sorted glacial sediment occurs as long ridges as much as 50 feet high composed of shale-rich pea gravel, cobble gravel, and till interbedded in deformed layers. These crevasse fillings are most abundant below altitudes of 1,200 feet. Their frequency increases toward the axis of the Minnesota River lowland.

Small discontinuous patches of extremely shale-rich sand and pea gravel are randomly distributed throughout the same geographic area characterized by the crevasse fillings. Commonly, these patchy deposits are less than 10 feet thick, and they appear locally to blanket the rolling topography. This type of deposit may be collapsed colluvium that was deposited by slope wash in shallow depressions on the ice-cored surface of the late-glacial landscape. Alternatively, these patches of sediment might represent strands of shallow lakes developed on the ice-cored terrain.

Terrace segments preserved at various heights above the floodplain of the Minnesota River fall within three major categories that relate to the history of the valley (fig. VII-23). The highest surface, only slightly inset into the till plain, is underlain by flat-bedded coarse sand and cobbly, well- to poorly-sorted gravel 10 to 40 feet thick. These sediments are remnants of an extensive braided stream system that drained the margins of the retreating Des Moines lobe. The master stream followed the axis of the regional topographic sag that had been such an important control on ice movement during glaciation.

Another set of terrace surfaces at intermediate heights is distinguished by a veneer of lag boulders that lie atop older Quaternary sediments or bedrock. These boulder-armored surfaces are remnants of successively lower channel bottoms of Glacial River Warren, a highly competent stream that discharged water from Lake Agassiz in late-glacial and early postglacial time.

A third type of sediment is found both slightly higher than the modern floodplain and locally buried beneath the floodplain sands and silts. These alluvial deposits are boulder-gravel beds composed of well-rounded boulders and cobbles in a matrix of coarse gravel and sand. Commonly, the dominant size reaches as much as 12 inches in diameter. These deposits were once part of the bedload of River Warren and lagged during the waning stages of its discharge through the present Minnesota River Valley.

Glacial Lakes

As the Des Moines lobe retreated from the trough-shaped Minnesota River lowland, ice and moraine barriers combined to pond water into a number of lakes. The former extent of these lakes is known chiefly from the distribution of laminated clay, silt, and sand and from the topographic position of lake outlet channels.

Upham (1896) proposed the name "Lake Minnesota" for an ice-marginal lake that he believed extended from Waseca to Ortonville. Later, Leverett (1932) proposed that Lake Minnesota was of much more limited extent. Their disagreement is based on alternate interpretations of the deployment of the retreating ice margin. No continuous strand features mark the limits of this lake. However, laminated fine-textured sediment and patches of well-sorted sand blanket much of the terrain below an altitude of 1,120 feet in Blue Earth, Faribault, and Watonwan Counties. The determination of the true areal extent of Lake Minnesota and a better definition of its history await more detailed field studies.

On the basis of extensive soils surveys, Diedrick (1967) concluded that a glacial lake of approximately 1,500 square miles covered parts of Swift, Chippewa, Big Stone, and Lac qui Parle Counties in western Minnesota. For this lake he proposed the name "Glacial Lake Benson." The basin rim had an altitude of approximately 1,050 feet. The lake was fed mainly by meltwater streams from the north and east that built deltas at their points of entry.

The rapid drainage of this lake may have produced some streamlined landforms in the vicinity of Montevideo and Granite Falls, which previously have been interpreted as drumlins (Matsch and Wright, 1966).

QUATERNARY HISTORY

Quaternary deposits in southwestern Minnesota are largely glacial tills and outwash, and include minor amounts of nonglacial deposits. Thick exposures along the Minnesota River and its tributaries reveal a complicated stratigraphic sequence that is characterized by superimposed till sheets separated by a variety of nonglacial sediments. The entire sequence is a dramatic testament to a long history of climatic fluctuations in middle-western United States.

Wright (1964) proposed the informal term "phase" to designate a time of glacial activity, identifiable either by stratigraphy or morphology. The history of southwestern Minnesota that follows is recounted within a framework of informally named phases where appropriate.

Pre-Wisconsin Events

In southwestern Minnesota, the Pleistocene glacial stages that preceded the Wisconsin are represented by thin, scattered deposits of iron-stained gravels, leached silts, and several weathered tills. Exposures at several localities along the Minnesota River show such sediments resting on Cretaceous clays. The lithology of these sediments indicates glacier flow from the north and northwest, and the superposition of several tills at the Ochs' clay pit near Morton suggests more than one pre-Wisconsin glacial advance from that direction. At least one of these glacial intrusions reached as far south as northwestern Iowa and southwestern Minnesota, where it is represented by a clay-rich calcareous till. Radiocarbon dates indicate that these glacial deposits are older than 40,000 years. No attempt is made to designate these deposits as Nebraskan, Kansan, or Illinoian at present. It is doubtful that these older deposits will be correctly interpreted until careful subsurface work is undertaken over very large areas.

Wisconsin Stage

Four glacial advances and recessions that can be attributed to the climatic fluctuations of the last major stage of the Pleistocene Epoch are recorded in the Quaternary deposits of the region. The distinctive lithologies of the tills, and their separation by sediments of non-glacial origin, allow their recognition over a wide area in southwestern Minnesota.

Earliest Phase

The first phase of ice activity recorded in the region is an advance into the area of a glacier from the north and northwest. The extent of this ice sheet is largely unknown; however, it deposited a loamy, calcareous, shale-free till in the vicinity of Granite Falls. This till unit is the fourth, or lowest, deposit exposed in the composite Hawk Creek section. The ice retreated, and subaerial erosion produced some lag deposits on slopes, resulting in the accumulation of silts and clays in depressions. Vegetation was established on the deglaciated landscape, and plant detritus joined inorganic sediments in the depressions.

Hawk Creek Phase

Before the development of a significant soil profile on this lowest calcareous till, a glacial advance from the Lake Superior area buried the landscape under a blanket of reddish-brown sandy till. This advance of an early Superior lobe is documented by the occurrence of the distinctive Hawk Creek Till all the way to the foot of the Coteau des Prairies in eastern South Dakota. Therefore, it must have left a girdle of red till across the entire midsection of Minnesota. The history of retreat of this glacial lobe is completely obscured by later glacial deposits. Subsequently, sediments accumulated in depressions on the exposed till sheet, and some of the lakes contained freshwater mollusks.

Granite Falls Phase

Renewed ice activity in the north and northwest resulted in another glacial advance and the subsequent deposition of a thick layer of loamy, calcareous, shale-free till and associated outwash throughout most of south-central and western Minnesota. This glacial depositional event may have taken place after 34,000 radiocarbon years ago. The maximum extent of the ice sheet that deposited the Granite Falls Till has not yet been determined; however, ice during this phase of activity extended at least as far as Lynd in Lyon County.

The ice lobe that deposited the Granite Falls Till retreated to an unknown northerly position, possibly as far as the Alexandria moraine. Upon the landscape laid bare by the melted ice sheet, subaerial erosion developed an extensive lag deposit of boulders, even on very gentle slopes. The environment at the time this boulder pavement accumulated may have been one of arid climate because neither soils nor organic deposits developed.

New Ulm Phase

The last glacier to advance across the area moved southward from the Winnipeg lowland and became lobate against the topographic buttress called the Coteau des Prairies. On the east side of the Coteau, the Des Moines lobe moved along the Minnesota River lowland and eventually covered almost all of southern Minnesota. This ice lobe spread a broad sheet of distinctive shale-rich calcareous till throughout the entire region, and it carried limestone and shale from the Winnipeg area as far south as Des Moines, Iowa. The timing of the ice movement through the New Ulm region is not precisely known. It had reached central Iowa by 14,000 radiocarbon years BP. A northeastern offshoot, the Grantsburg sublobe, crossed the St. Croix moraine in the Minneapolis area not long after that region had been vacated by the Superior lobe, sometime after 15,000 years BP (Wright, 1971). Till, peripheral to the Bemis moraine, dated tenuously as 20,000 years old in Iowa, may represent the southwest margin of this advance.

In its advance through southwestern Minnesota, the Des Moines lobe assumed the geometry of a broad valley glacier, confined on the southwest by a flank of the Coteau des Prairies and on the northeast by ice-cored terrain of the Alexandria moraine. Within the confines of this broad topographic channel, the ice moved generally southeastward along its axis, with a component of flow toward the margins. Along its axial part, the glacier moved across the boulder pavement that had developed on the Granite Falls Till sheet, but did not destroy it. Rather, the basal ice faceted and striated individual stones. On either side of the axis, however, subglacial erosion was active, and the originally abundant siliceous Cretaceous shale was diluted by the entrainment of locally derived material. Eventually, an impressive lateral moraine marked the southwest margin of the Des Moines lobe; deposition along the northeast side of the ice was less impressive.

By 13,000 years ago, the Des Moines lobe had melted back almost to the Minnesota border from its maximum stand in central Iowa (Ruhe, 1969), and in the next thousand years the deterioration had completely cleared active ice from southern Minnesota. A boreal forest was established at Madelia by 12,650 years BP. However, much of the terrain remained cored with deadice. Water and sediments were ponded behind ice-cored moraine dams and within unstable basins on the stagnant-ice surface. The deposition of these sediments was most impressive along the line of contact between the lateral moraines and the ice margin as it melted back toward the glacial axis. Lake waters breached and eroded sharp outlet gorges across the moraines in a few places, resulting in their rapid drainage. Other outlets no doubt were cut into the stagnant-ice dams, but their record has melted away.

This time of rapid ablation of the Des Moines lobe is represented in the region by a great network of meltwater channels, and by lacustrine and other sediments associated with ice disintegration, especially elongate crevasse fillings along the broad axis of the Minnesota River lowland.

Lake Agassiz Phase

As the Des Moines lobe retreated from the Big Stone moraine between Browns Valley and Ortonville, water became ponded in the newly exposed Red River basin. Eventually, this lake, called Glacial Lake Agassiz, expanded across an area of about 200,000 square miles (Elson, 1967). During its early stages, Lake Agassiz had just one outlet, the Glacial River Warren, a high-volume stream that discharged southeastward along the axis of the Minnesota River lowland where it followed a course previously occupied by a braided meltwater stream. The highly competent outlet stream entrenched itself into the landscape and continued to deepen and widen its valley as Lake Agassiz expanded (Matsch and Wright, 1967).

Periodically, the channel bottom became armored with large boulders and the river stabilized. Subsequent increases in competence resulted in renewed downcutting and the river eventually exposed the Precambrian bedrock that had been deeply buried by several till sheets. Eventually, the ice melted back sufficiently far to expose other outlets for Lake Agassiz, and Glacial River Warren was beheaded. Consequently, the wide and deep channel now carries only a small fraction of its former discharge, and the present-day Minnesota River is a classic example of an underfit stream.

The leading edge of the boreal forest migrated northward as the active ice retreated, and by 10,850 years BP the entire region had been invaded. Some time later, the forest gave way to prairie vegetation, all the buried ice was melted, and the modern drainage pattern established. At the present time, erosion appears vigorous, all the streams tributary to the Minnesota River are extending themselves headward, and the master stream seems to be regrading itself to the postglacial regime by deposition.

PHYSIOGRAPHY OF MINNESOTA

H. E. Wright, Jr.

Minnesota has the largest area of any of the middle western states, and in many respects it has the most diverse landscape, especially when one includes as landscape elements not only the landforms but also the vegetation. The diversity results from three factors—the geologic framework, the glacial history, and the climatic setting. Although the bedrock almost everywhere is mantled by glacial drift, it exerts a strong topographic influence in the northeastern part of the state, where the crystalline rocks of the Canadian Shield form highlands and hills, and in the southeastern part, where the Mississippi River and its tributaries have cut through flat-lying Paleozoic rocks to produce sharp valleys with rock bluffs. Elsewhere in the state the glacial landforms—rugged and massive lake-dotted moraines, broad, pitted outwash plains, smoothly ridged drumlin fields, vast lake plains, or rolling loess-covered plains—are dominant. The distribution of lakes gives a clue to the landforms (fig. VII-24).

The climatic effect is manifested by the vegetation. Temperature lines are roughly latitudinal (fig. VII-25), and precipitation lines longitudinal (fig. VII-26), so the resultant vegetational boundaries trend diagonally (fig. VII-27). In the northeast are dense coniferous forests, which give way toward the center of the state to deciduous forests. These in turn grade to prairie, which dominates the western and southwestern parts of the state. The vegetational zones transect the glacial subdivisions, so the soils, which are controlled in their formation by both vegetation and geology, have an even more complex distribution.

In the northeast, the physiographic subdivisions here shown (fig. VII-28) follow in large part those earlier presented (Wright, 1956), and in the north-central part of the state they are consistent with those recently made for maps of soils and landforms (Dept. Soil Science, 1969, 1972). Elsewhere the boundaries of the subdivisions are generalized or locally modified from Leverett (1932). The assistance of C. L. Matsch in describing some of the southern regions is appreciated.

1. BORDER LAKES AREA

The Border Lakes area of bedrock lakes occupies a belt about 25 miles broad extending about 130 miles westward along the Canadian border from the northeastern end of the North Shore Highland. The eastern third of this area is traversed by the Gunflint trail. Here, glacial activity was largely confined to differential erosion of bedrock, producing patterns of lakes and ridges that delicately reflect the rock structure. The patterns are particularly refined on air photos, but they are noticeable on any drainage map of the area.

Several distinct patterns can be distinguished (Zumberge, 1952, p. 24). In the east, the lakes are dominantly linear and trend east-west (fig. VII-29). Within about 5 miles of the Canadian border, the lakes and other valley features are etched in the relatively weak slates of the Rove Formation, and the ridges are formed by southward-dipping sills of diabase (fig. VII-30). A pattern of subsequent streams undoubtedly existed here in preglacial time (ver Steeg, 1947), but valleys were locally overdeepened by the erosion of glacier ice, which passed transversely across the valleys and rode up over the ridges (Zumberge, 1955).

South of the belt of Rove Formation and diabase sills, the lakes also have east-west linearity, but they are generally narrower and shallower and are symmetrical in section. The bedrock is the Duluth Complex, and the lakes are localized by weak zones parallel to the layering in the gabbro.

Northwest of the belt of linear east-west lakes, in the broad northward protuberance of the Minnesota boundary, is an area of irregular lakes with some linear and rectangular segments. The bedrock is Saganaga Tonalite, a granitic rock in which the joint pattern has had some control on glacial erosion.

Southwest of the Saganaga area through the Knife Lake area to Lake Vermilion, the pattern shows a stronger linearity, with dominant trend to the east-northeast, but with segments transverse and oblique. The bedrock in this belt consists of metamorphic rocks cut by numerous faults and dikes.

Finally, the large area north of Lake Vermilion is underlain by jointed and faulted granite, and the lake pattern is less regular.

2. NORTH SHORE HIGHLAND

The North Shore Highland, underlain mostly by southeastward-dipping Keweenawan basalt and diabase, overlooks Lake Superior from a height of 900-1,500 feet all the way from Duluth to the Canadian border. The shore itself is relatively straight, although in detail it is interrupted by points and bays that reflect the differential resistance of the igneous rocks. Short streams, 10-15 miles long, lead from the highland directly to the lake. Most of them have falls in their lower reaches. The falls started in late-glacial time as the level of the Glacial Great Lakes fell, and they have retreated upstream several hundred feet. The linear pattern produced by the streams reflects a finer pattern of till ridges and grooves (fig. VII-31).

Although the North Shore Highland as here delineated appears as a distinct highland from the lake shore, it is much less prominent from the interior. Its inner edge is

Figure VII-24. Map of Minnesota showing distribution of lakes. Major drainage basins are outlined by dashed lines.

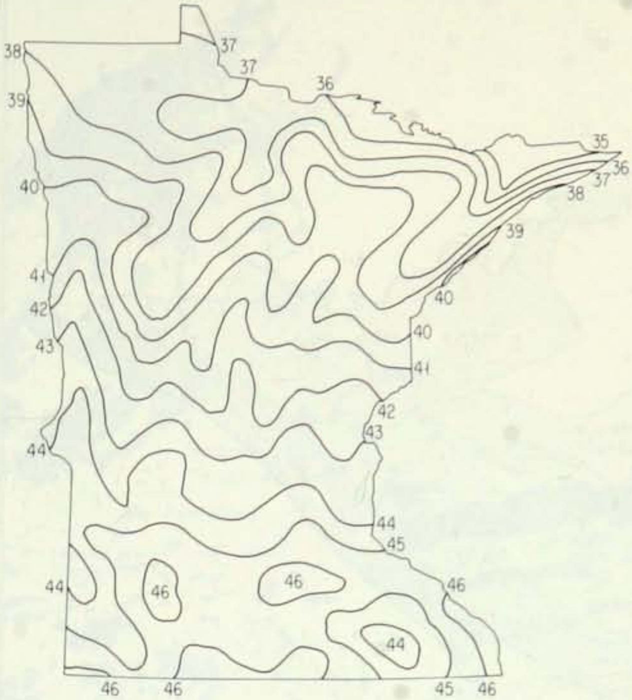


Figure VII-25. Average annual temperature in Minnesota (F°) (Baker and Strub, 1965).

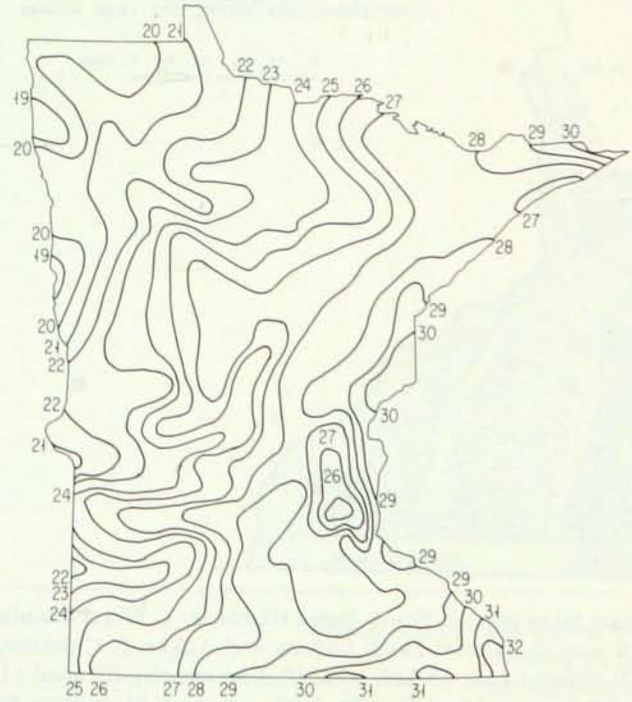


Figure VII-26. Average annual precipitation in Minnesota in inches (Baker and others, 1967).

Figure VII-27. Vegetation of Minnesota before extensive land settlement (Upham, in Winchell, 1884b).

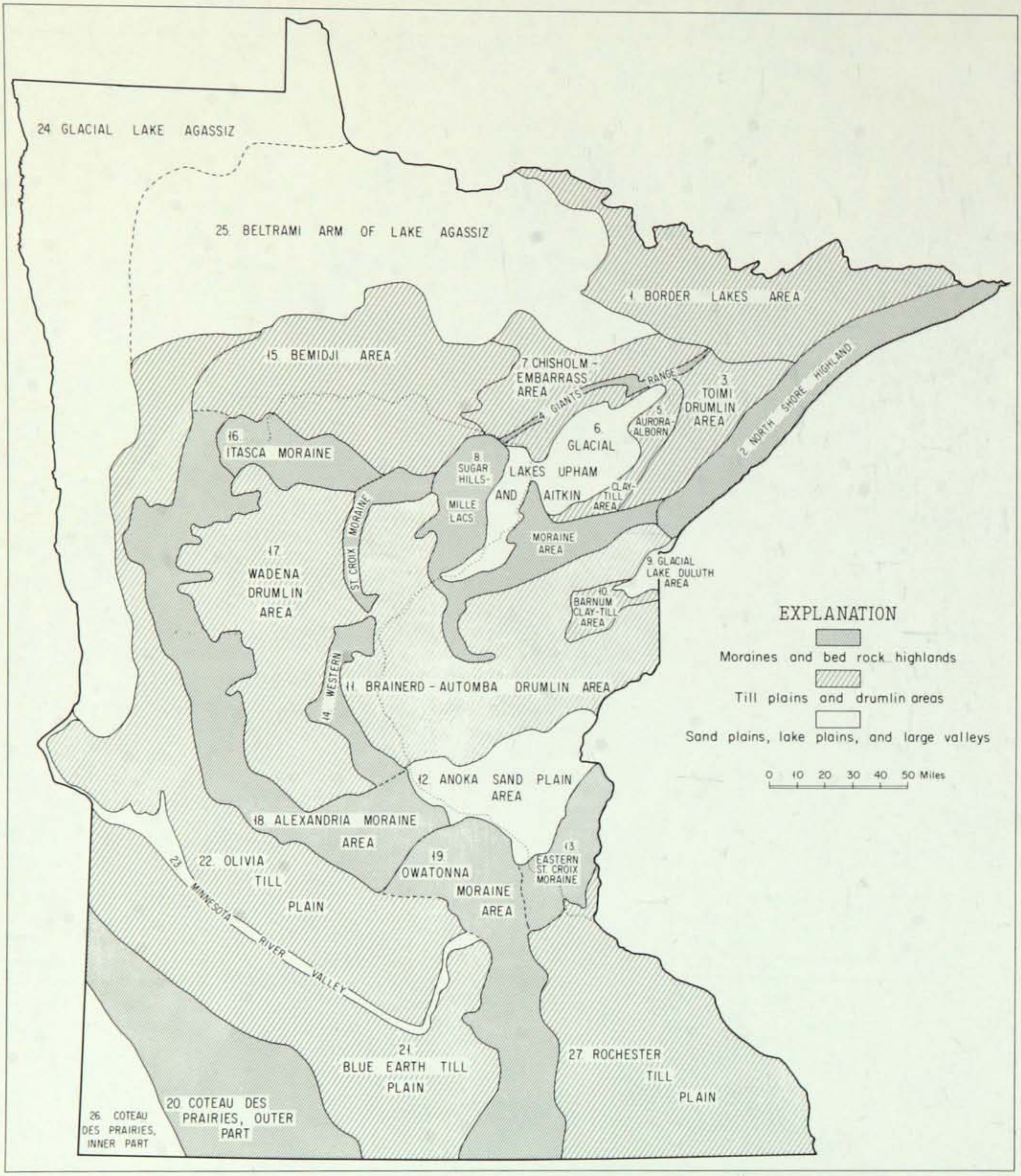


Figure VII-28. Map of physiographic areas in Minnesota. 1, border lakes area; 2, North Shore Highland; 3, Toimi drumlin area; 4, Giants range; 5, Aurora-Alborn clay-till area; 6, Glacial Lakes Upham and Aitkin; 7, Chisholm-Embarrass area; 8, Sugar Hills-Mille Lacs area; 9, Glacial Lake Duluth area; 10, Barnum clay-till area; 11, Brainerd-Automba drumlin area; 12, Anoka sandplain area; 13, eastern St. Croix moraine; 14, western St. Croix moraine; 15, Bemidji area; 16, Itasca moraine; 17, Wadena drumlin area; 18, Alexandria moraine area; 19, Owatonna moraine area; 20, Coteau des Prairies, outer part; 21, Blue Earth till plain; 22, Olivia till plain; 23, Minnesota River Valley; 24, Glacial Lake Agassiz; 25, Beltrami arm of Lake Agassiz; 26, Coteau des Prairies, inner part; 27, Rochester till plain.

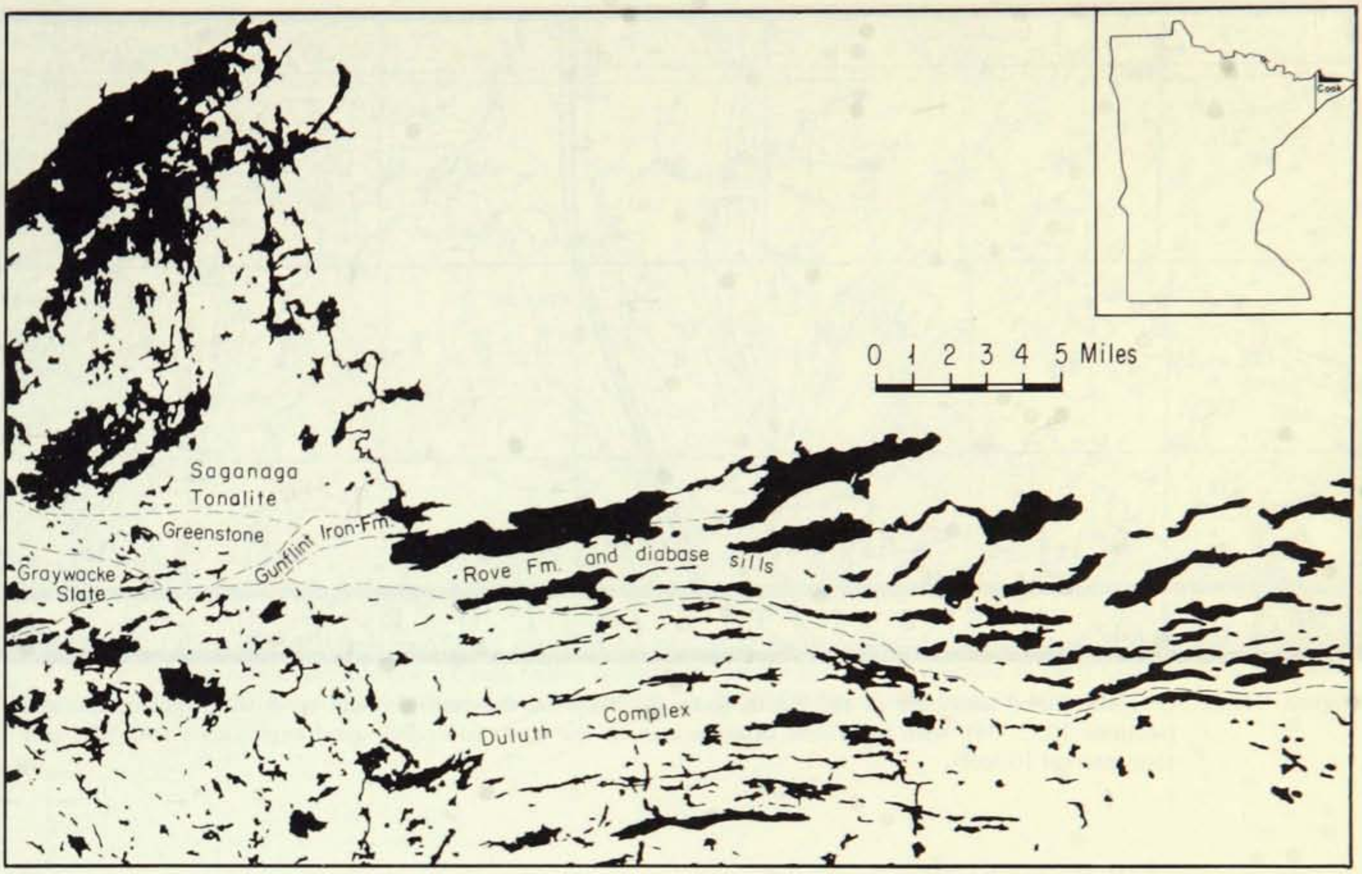


Figure VII-29. Map of much of Border Lakes area to show the relations between lake patterns and bedrock type (modified from Zumberge, 1952).

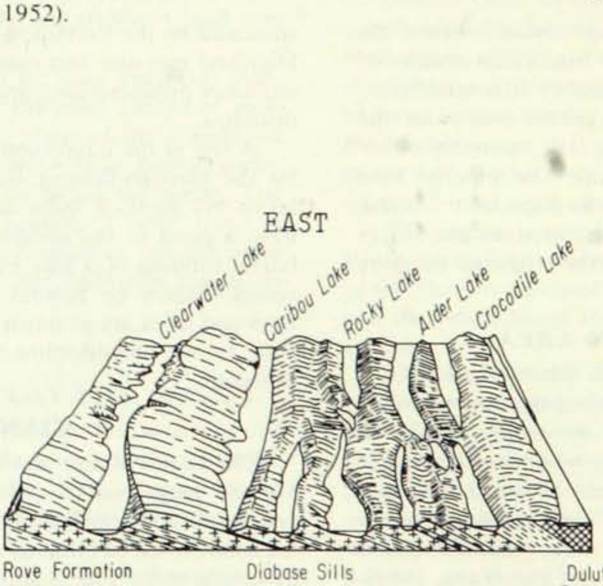


Figure VII-30. Block diagram of part of Border Lakes area showing the bedrock structure. The ice sheet moved from left to right across the area and deepened the pre-existing stream valleys, which now hold the long, narrow lakes characteristic of this region (from Zumberge, 1952).

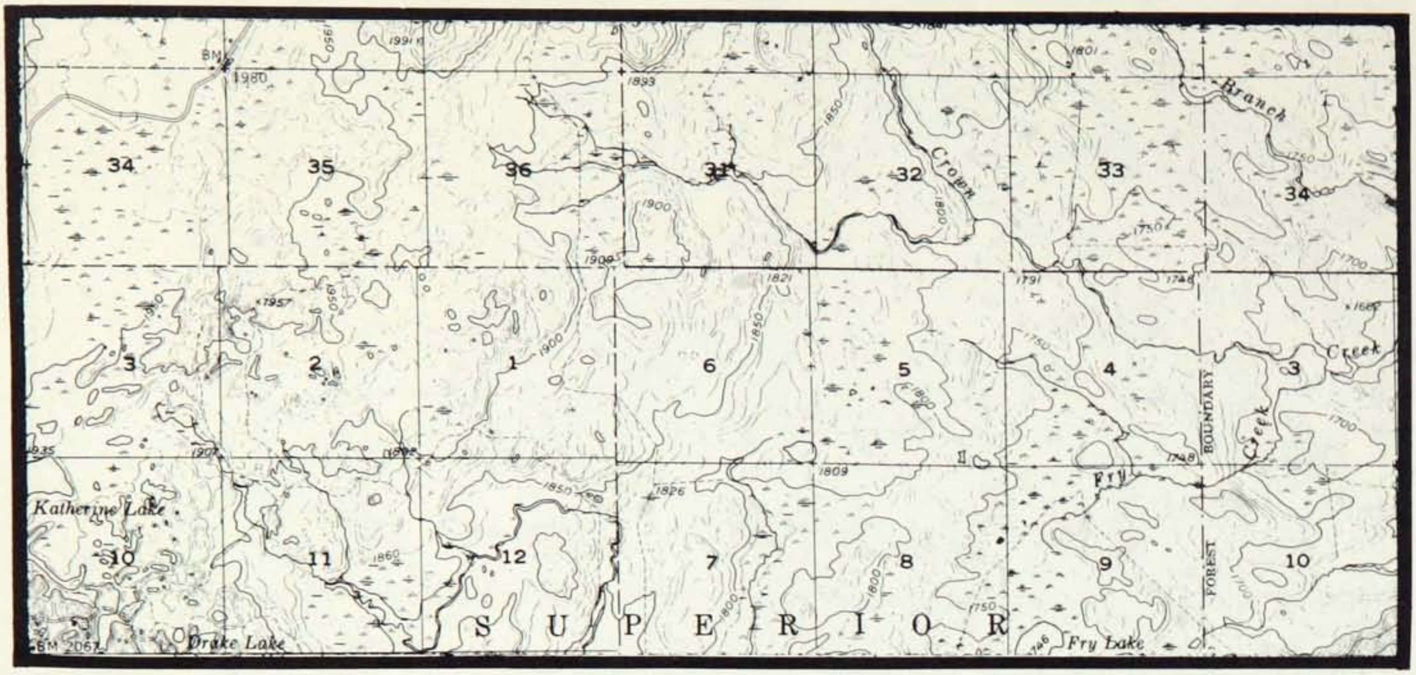


Figure VII-31. Glacially fluted landscape of the North Shore Highland on the east, terminating in the Highland moraine (sections 10, 2, 36), with the Toimi drumlin area on the west; Silver Bay quadrangle (scale 1:62,500; contour interval 10 feet).

here taken principally as the drainage divide between the short coastal streams and the linear headwater streams of the St. Louis River system. This boundary is controlled by glacial features but coincides in a general way with the northwestern limit of the lava flows. The boundary is actually the toe of the Highland moraine. The Superior lobe at one phase of glaciation just filled the Superior basin and spread up the steep slope, forming the grooves and ridges mentioned above and terminating at the Highland moraine (Wright and Watts, 1969).

3. TOIMI DRUMLIN AREA

Northwest of much of the North Shore Highland is a triangular area marked by southwestward-trending drumlins and a linear stream pattern. Most of the area is within the Superior National Forest; it is heavily wooded and sparsely inhabited, and the dominating pattern of ovoid hills and linear drainage was not appreciated until recent air-photo studies. The drumlins are about 1-2 miles long, one fourth mile broad, and 30-50 feet high (Wright and Watts, 1969).

The region is drained to the southwest by the Whiteface and Cloquet Rivers, which join the St. Louis River near where the latter turns abruptly southeastward toward the head of Lake Superior. The St. Louis River approximately delimits the region on the south, where younger drifts obscure the drumlin pattern. On the west, the Toimi drumlins are overlapped by red clayey drift of the St. Louis sublobe in the Aurora-Alborn area (fig. VII-32). On the north, it is truncated by the Vermilion moraine, and on the east by the Highland moraine, and outwash sediments from both these moraines produced long gravelly plains winding among the drumlins.

A few of the interdrumlin swales that were not affected by the through-flowing outwash streams are filled with lakes, but most of them contain bogs. Some of the bogs have a pond in the middle, indicating the progression of lateral infilling of a lake by bog growth after the lake becomes shallow by bottom filling. The sediments in such bogs and lakes are as much as 40 feet thick, so when originally formed the drumlins had almost twice as much relief as today.

4. GIANTS RANGE

The Giants Range is a highland of granite flanking the Mesabi range on the north from Hibbing to Babbitt, and rising 200-400 feet above the plains to the north and the south. West of Hibbing the granite belt is largely buried by glacial drift. The Giants Range contains the three-way divide between drainage to Hudson Bay, the Great Lakes, and the Gulf of Mexico. Some streams transect the range. The Embarrass River has a course about 25 miles long before it cuts southward through the range near Aurora as a major tributary of the St. Louis River. This river once carried a large volume of glacial meltwater from a lake on the Embarrass plain north of the range to Glacial Lake Upham to the south. Its course is marked by a string of lakes that

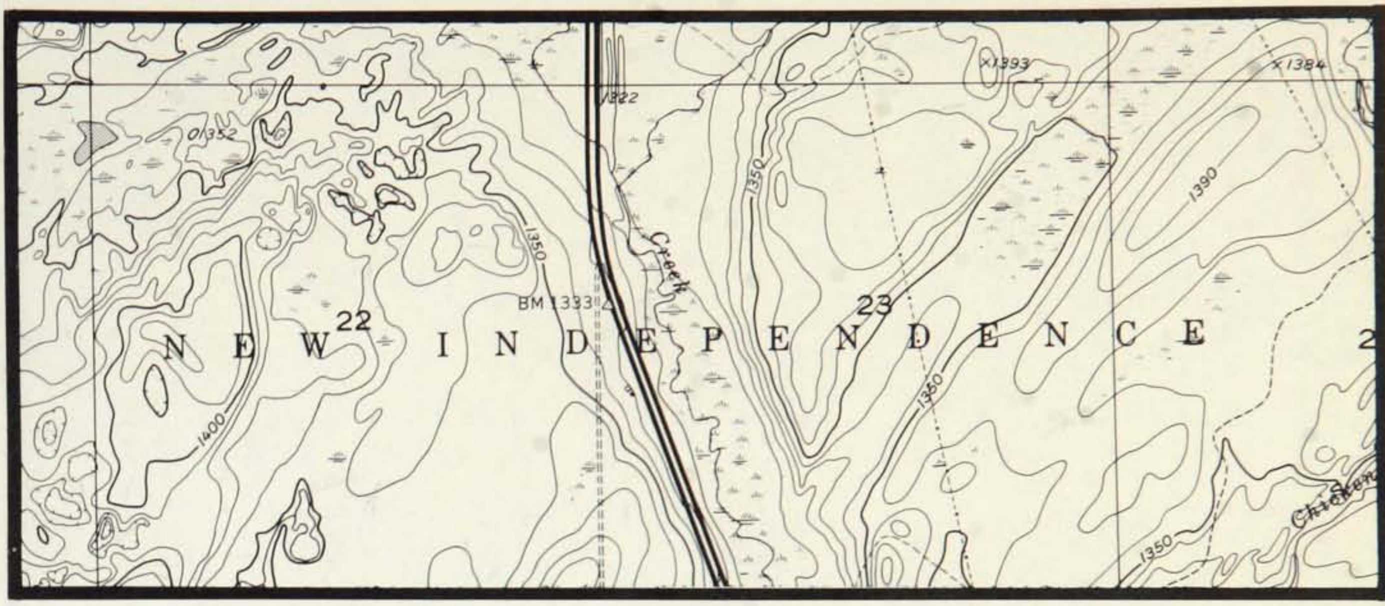


Figure VII-32. Southwestward-trending Toimi drumlins on east, with the irregular clay-till moraine of the Aurora-Alborn area on the west. Hellwig Creek valley through the center is an outlet channel of early Glacial Lake Upham; U.S. Highway 53 on Independence quadrangle (scale 1:24,000; contour interval 10 feet).

represent buried blocks of glacier ice; one of these lakes was drained in the development of the Lake mine west of Aurora, and thick deposits of sand are visible in road cuts and iron ore pits in the vicinity.

The Mesabi range is marked by huge open pits hundreds of feet deep, bordered by towering dumps of rock waste that can be seen for miles. Many of the abandoned pits now have clear-water lakes at the bottom, but the rocky slopes are largely bare of vegetation, as are the dumps. The older mines are located on local pockets of "soft" iron ore along the continuous belt of iron-formation, but the more recent pits are on the unaltered iron-formation (taconite) itself.

5. AURORA-ALBORN CLAY-TILL AREA

Burying the Toimi drumlin field on the west and the Giants Range on the south is a fringe of red-brown clay till deposited by the St. Louis sublobe in its advance to the northeast and east. The ice picked up clay from the sediments of an earlier Glacial Lake Upham and redeposited it at the ice lobe margin, generally as a veneer less than 25 feet thick (Wright and Watts, 1969). Some of the buried landforms, such as Toimi drumlins, are still visible beneath the till cover. The till is well exposed as the surface material in many open pits of the Mesabi range.

Along the southeastern part of the area the till forms a distinct moraine (fig. VII-32), which crosses the St. Louis River. The clay till here not only buries the Toimi drumlins but laps up onto a Superior-lobe moraine as well.

6. GLACIAL LAKES UPHAM AND AITKIN

Glacial Lake Upham is south of the Giants Range in southwestern St. Louis County. The lake plain and marginal sand plains consist of a broad expanse of swamp-covered silt and sand. The northern part, crossed for 20 miles by the highway leading north to Virginia, is mostly sand, and near the Mesabi range it contains ice-block lakes and sand terraces that finger into Embarrass gap and other glacial sources on and north of the range.

This region is now drained southward by the St. Louis River, which turns abruptly southeastward at the south end of the plain to flow past the end of the Toimi drumlin field and the North Shore Highland and enter Lake Superior at Duluth.

The Glacial Lake Aitkin plain is transected by the intricately meandering Mississippi River. It is separated from Glacial Lake Upham to the east in part by a high morainic ridge and in part by an alluvial fan deposited on the lake plain by the glacial Mississippi River, which entered the northwest end of the Aitkin plain. At the time of its full development, however, Lake Aitkin was not simply an expansion of the Mississippi River, for the outlet gorge through the moraine on the south between Aitkin and Brainerd (fig. VII-33) is too small to have accommodated the flow from a large glacier-fed lake (Farnham and others, 1964). This outlet only developed in postglacial time, after glacial retreat had reduced the volume of water. During glacial time, the main lake was confluent eastward with Glacial Lake Upham, which produced a substantial outlet gorge eastward down the St. Louis River.

Figure VII-33. Glacial Lake Aitkin plain, contained on the south and west by moraines of the Sugar Hills-Mille Lacs area. The Mississippi River flows in irregular incised meanders through the plain. The moraine in this area consists mostly of clay-rich till composed of reworked lake beds; Cuyuna quadrangle (scale 1:62,500; contour interval 10 feet).

7. CHISHOLM-EMBARRASS AREA

Between the Giants Range and the eastern arm of Glacial Lake Agassiz is a wedge-shaped area of low moraines and outwash plains here called the Chisholm-Embarrass area. It is bounded on the north by the moraines that excluded Lake Agassiz. These moraines, which trend roughly to the east, were deposited by the Rainy lobe. They are stony and dominated by crystalline rocks. The western part of the area was subsequently overridden by the St. Louis sublobe from the west, so the surface drift there is fine grained and calcareous. Outwash plains are common between the moraines.

8. SUGAR HILLS-MILLE LACS MORAINE AREA

The Sugar Hills-Mille Lacs moraine area includes several moraines from Mille Lacs Lake to the Grand Rapids area—not all closely related in genesis. The most distinctive is the arcuate moraine that bounds Mille Lacs Lake on the south and west. It consists mostly of sandy till and outwash related to the Superior lobe, but on its inner side it bears a cap of clay till deposited when the St. Louis sublobe spread out of the Glacial Lake Aitkin basin (fig. VII-33).

The morainic topography extends both northward along the west side of the Lake Aitkin basin to the Sugar Hills, and northeastward along the east side of the Aitkin basin to the Sandy Lake and Jacobson areas, and eastward to connect with the North Shore Highland. Much of this area also has a cap of St. Louis sublobe till on top of stony or sandy moraine.

9. GLACIAL LAKE DULUTH AREA

Near the head of the Lake Superior basin is a partly dissected clay plain that marks the former bed of Glacial Lake Duluth, at an altitude of about 1,000 feet above sea level, or 400 feet above the present level of Lake Superior. The plain is deeply dissected by the St. Louis River and its tributary the Nemadji River (fig. VII-34). Deep river and highway cuts (especially Minnesota Highway 23) show great landslides of the homogeneous red clay down even relatively gentle slopes.

Lake Duluth filled much of the Lake Superior basin and drained southward into the Brule River in Wisconsin and thence to the St. Croix. Its strand line along the north shore of Lake Superior has been traced by wave-cut cliffs, stream deltas, and local deposits of red clay. The clay is sufficiently widespread in some places to favor different forest cover, but elsewhere the terrain is bedrock or stony till.

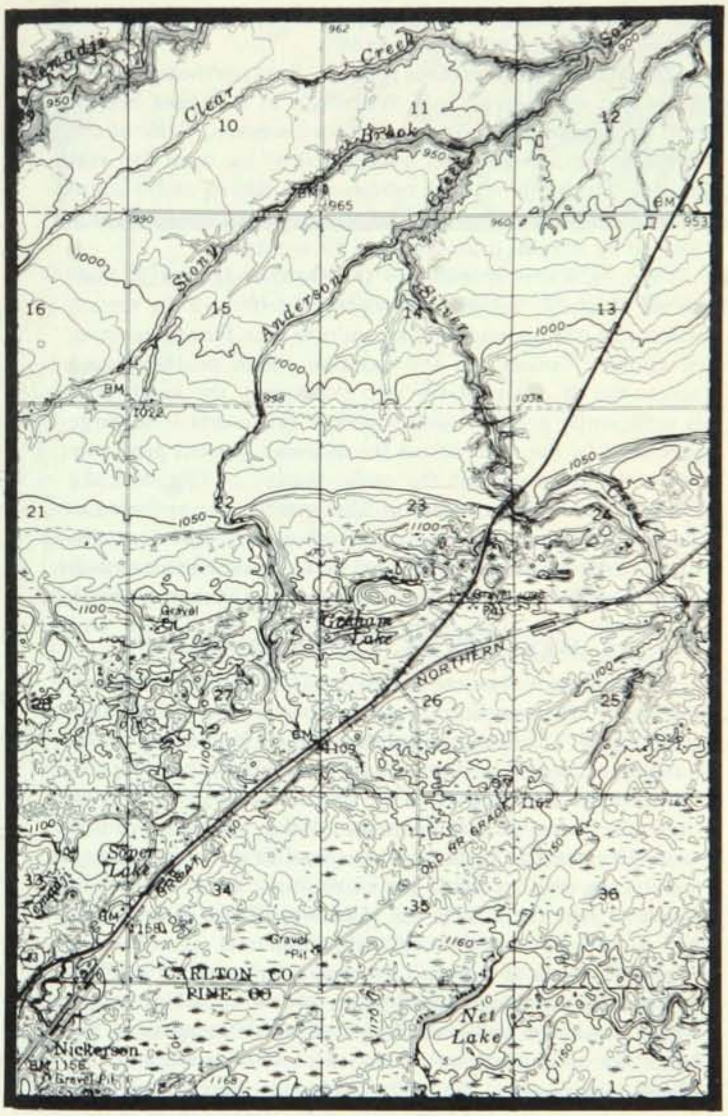


Figure VII-34. Nickerson moraine and Glacial Lake Duluth plain southwest of Duluth along State Highway 23. Lake Nemadji strand line is at an altitude of about 1,050 feet, and Lake Duluth about 1,000 feet. The streams dissecting the clay plain are tributaries of the Nemadji River; Holyoke quadrangle (scale 1:62,500; contour interval 10 feet).

The clay plain of Lake Duluth at the head of the lake is fringed by a narrow sandy plain at an altitude of about 1,060 feet. This is the record of the slightly older Glacial Lake Nemadji, which drained westward through a boulder-paved channel to the Moose River and thence to the St. Croix (Wright and Watts, 1969).

10. BARNUM CLAY-TILL AREA

The Glacial Lake Duluth area at the west end of the Lake Superior basin is rimmed by an area of red clay till and associated outwash, formed during at least two short advances of the Superior lobe, in which proglacial lake clays were overridden and redeposited. The most conspicuous landform in the area is the Nickerson moraine, which represents the southeast flank of the ice lobe at one time (fig. VII-34). The northwest flank at the same time apparently was marked not by a till moraine but rather by a series of frontal outwash plains. Beyond the Nickerson moraine, for an additional 20 miles to the southwest, about as far as the bedrock divide that separates the Lake Superior basin from the Minneapolis lowland, is a thin veneer of clay till.

11. BRAINERD-AUTOMBA DRUMLIN AREA

The large Brainerd-Automba drumlin area constitutes most of the ground moraine of the Rainy and Superior lobes inside the arc of the St. Croix moraine and not buried by the Anoka Sandplain on the south or by younger drift on the north.

Much of the Brainerd-Automba area is marked by drumlin fields. The largest is the Pierz area, south of Mille Lacs Lake (fig. VII-35). This shows a nice fan-shaped pattern to west and southwest, interrupted by the Mississippi River but identifiable on the west side just inside the St. Croix moraine.

A second drumlin field in the area is the Brainerd, which has a southwest trend, west of the Mille Lacs moraine. This is also interrupted by the Mississippi River, and northwest of Brainerd a few additional drumlins are present.

A third drumlin field is east of Mille Lacs Lake. This is the Automba field, which has a pattern fanning from northwest to west and southwest. It can be traced north of the St. Louis River beyond Cloquet.

The entire area is interrupted in numerous places by outwash plains. The largest single one is the Mississippi River valley train, which enlarges north of Brainerd into a great complex of pitted plains, mostly leading to moraines of the Sugar Hills-Mille Lacs area. The Mississippi River valley train also received major contributions of outwash from the Crow Wing River gap in the St. Croix moraine west of Brainerd.

The area south of Mille Lacs Lake is also interrupted by sharp erosional valleys containing swamps, lakes, or underfit streams (fig. VII-35). These are considered to be tunnel valleys, formed by subglacial streams flowing under very great hydrostatic pressure.

12. ANOKA SANDPLAIN AREA

North of the eastern arm of the St. Croix moraine, and between the Mississippi and St. Croix Rivers, is a broad sandplain, formed largely by glacial drainage from the north and west that was held back by the moraine (fig. VII-36). The area had been covered by the Grantsburg sublobe, which advanced northeastward up the Minneapolis lowland over the ground previously filled by the Superior lobe. But with the wastage of the Grantsburg sublobe the meltwater streams shifted across the vacated ground, until they found outlets to the south, first by way of the St. Croix River, then directly to the Mississippi.

The Anoka sandplain is not featureless, by any means. Low regions of upland represent areas of till that were not buried by the outwash sand. Other features of positive relief are patches of sand dunes, formed by southwesterly winds after the sandplain was abandoned by the outwash streams. Landscape features of negative relief include numerous lakes and marshes, representing ice blocks originally buried by the outwash sand. Many of the depressions are long troughs trending to the southwest. These are remnants of tunnel valleys formed beneath the Superior lobe by major subglacial streams draining to the ice margin (Wright, in press). When the ice lobe retreated, stagnant ice remained behind in the valleys and was subsequently buried by the outwash sand.

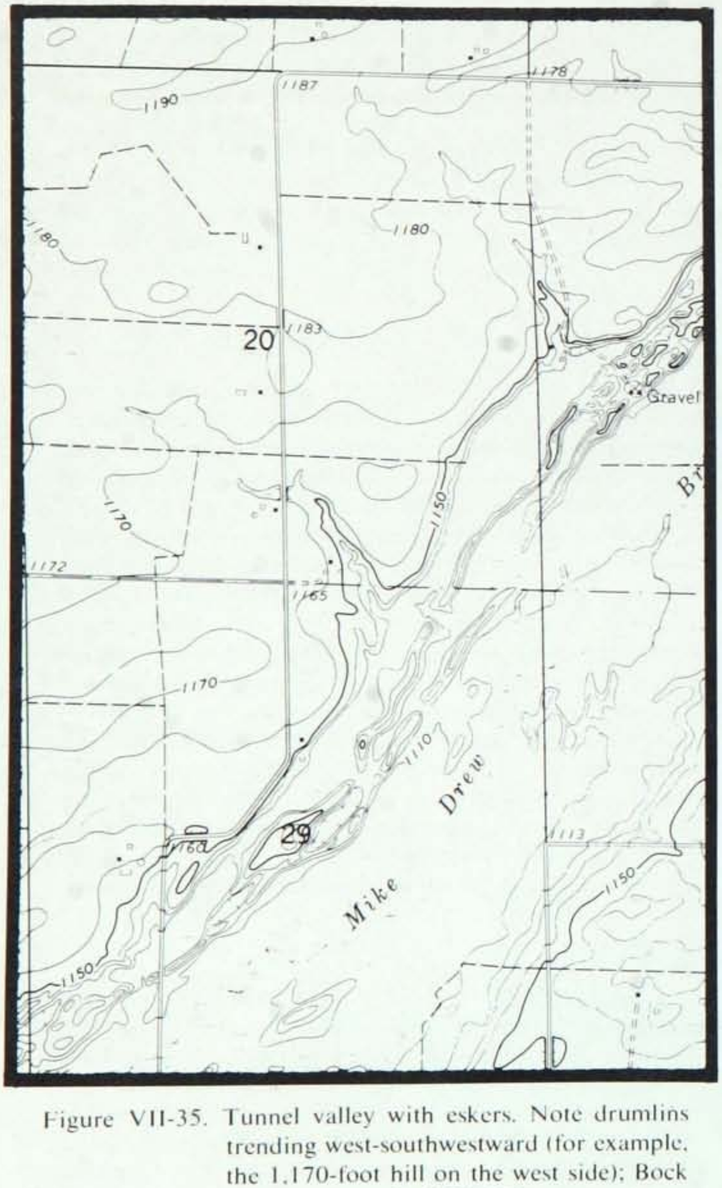


Figure VII-35. Tunnel valley with eskers. Note drumlins trending west-southwestward (for example, the 1,170-foot hill on the west side); Bock quadrangle (scale 1:24,000; contour interval 10 feet).

13. EASTERN ST. CROIX MORAINE

One of the sharpest moraines in Minnesota is the St. Croix, which marks the limit of the combined Superior and Rainy lobes during the St. Croix phase of Wisconsin glaciation. The central section of the moraine was later overridden by another ice lobe and partly obscured; the remainder can be considered in two segments.

The eastern part extends from St. Paul northeastward to Stillwater and beyond into Wisconsin as a rugged belt of hills and depressions. The inner (northwestern) flank of the moraine was subsequently overlapped by ice from the west (Grantsburg sublobe). The moraine is bordered on the south in Dakota County by a broad outwash plain, presumably formed by streams draining from the ice front.

The St. Croix moraine in the eastern segment is composed of stony, reddish-brown glacial drift, and the terrain is less suitable for intensive agriculture than for scenic siting of country houses. As the moraine passes northeastward across the St. Croix River into Wisconsin, the local relief is sufficiently great to sustain several ski resorts. The gorge produced by the river across the moraine was cut primarily when the river carried the outlet waters of Glacial Lake Duluth and its predecessors. The spectacular potholes at Taylors Falls were excavated in the basaltic bedrock at this time by the turbulent sand-laden water—some are as much as 10 feet in diameter and 17 feet deep (Alexander, 1932).

14. WESTERN ST. CROIX MORAINE

The western segment of the St. Croix moraine borders the upper Mississippi River on the west for about 100 miles from the St. Cloud area north to Walker. It averages about 6 miles in breadth and presents a particularly sharp face to the west, where it is fringed for part of its length by frontal outwash plains that bury parts of the Wadena drumlin field. The moraine is transected west of Brainerd by a broad water gap, which carried the Crow Wing River and all its outwash from the west to the Mississippi River after the ice had withdrawn from the St. Croix moraine.

In the segment west of Little Falls (Schneider, 1961) the moraine is cut longitudinally by several broad drainageways, which are well graded in their northern portions but break up southward into a series of blind ends and ice-block depressions. The drainageways were apparently formed by southward-flowing outlet streams from a proglacial lake during ice retreat, and then were subsequently overridden at their southern ends by an ice lobe from the southwest.

15. BEMIDJI AREA

The Beltrami arm of Glacial Lake Agassiz was contained on the south by a complex of moraines and outwash plains here called the Bemidji area. This is a heavily for-

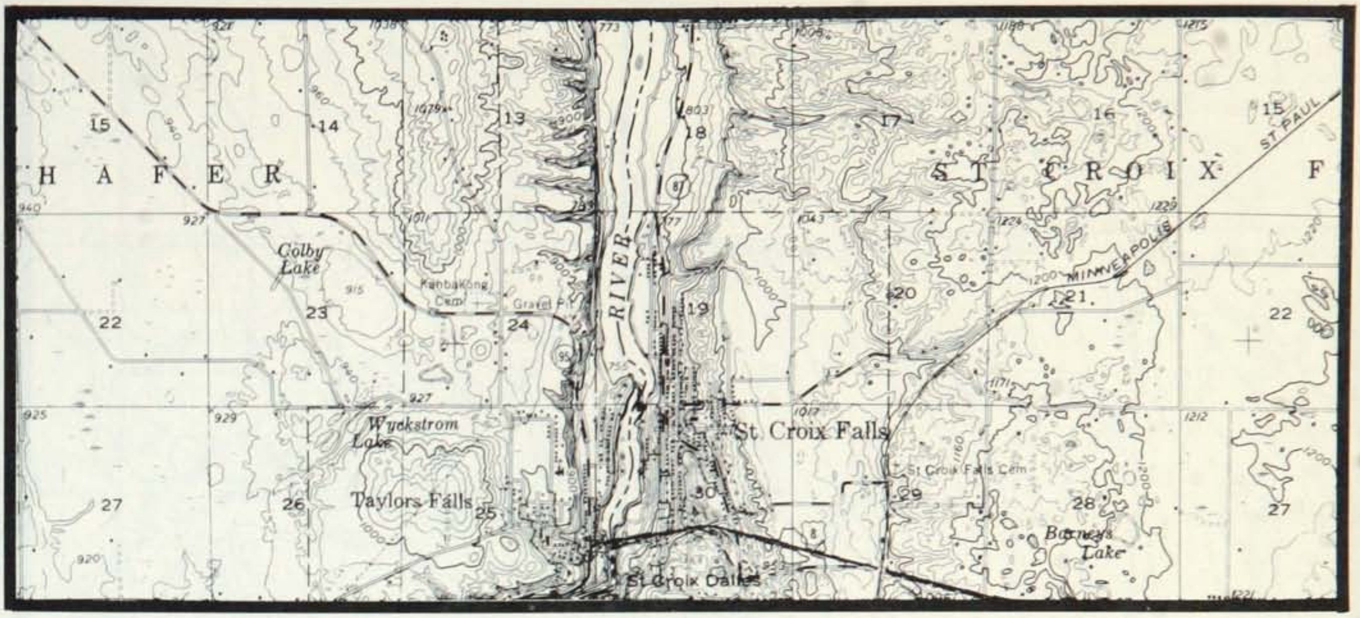


Figure VII-36. Eastern St. Croix moraine cut by St. Croix River gorge and bordered on the west by the Anoka sandplain. St. Croix Falls quadrangle (scale 1:62,500; contour interval 10 feet).

ested and poorly known region whose glacial landforms probably relate both to the movements of the Wadena lobe and to subsequent invasion by the St. Louis sublobe. The principal outwash area extends eastward as a long, broad, pitted plain from Bagley to Bemidji and Lake Winnibigoshish, a plain now followed in part by the Mississippi River. The outwash streams continued eastward through the Grand Rapids area into Glacial Lake Aitkin.

Between this principal outwash area and the Itasca moraine to the south is a series of deep north-south troughs that in some cases continue through the Itasca moraine. Some of them are filled with sand, but others have long bogs or lakes. The Mississippi River flows northward as an underfit stream in one of the bog-filled troughs and then shifts over to another. Lower La Salle Lake, located in another trough, is more than 200 feet deep.

These troughs were probably eroded as tunnel valleys by powerful southward-flowing subglacial streams, which were under great hydrostatic pressure when the front of the Wadena lobe stood at the Itasca moraine. The northern ends of the troughs are partly obscured because they were overridden later by the St. Louis sublobe.

16. ITASCA MORAINE

South of the Bemidji area is the prominent east-west Itasca moraine, which is a deposit of the Wadena lobe. The moraine is characterized by numerous north-south lake-filled trenches that continue southward the pattern of tunnel valleys described for the area to the north. The two arms of Lake Itasca are in two such trenches.

Within the heart of the Itasca moraine some of the tunnel valleys can barely be traced as a row of small lakes, whereas others are broad, sweeping trenches. The difference probably reflects the amount of rock debris that collapsed into the subglacial troughs when the ice melted.

The tunnel valleys end abruptly at the south edge of the moraine, where the streams emerged and disgorged their great loads of gravel to form the Park Rapids outwash plain.

Some of the tunnel valleys contain small eskers, which were formed when the much diminished subglacial streams deposited sand and gravel after the ice had thinned so much that the hydrostatic pressure was lost.

Although the tunnel valleys were formed beneath the Wadena lobe, they were occupied later, perhaps still as tunnels under stagnant ice, by outwash streams draining southward from the overriding St. Louis sublobe, for telltale fragments of Cretaceous shale (the index to St. Louis lobe drift) have been found in terrace deposits near the Lake Itasca tunnel valley.

17. WADENA DRUMLIN AREA

South of the Itasca moraine and west of the St. Croix moraine is the Wadena drumlin field, buried next to these moraines by outwash plains. The Wadena drumlin field contains about 1,200 conspicuous drumlins in a fan-shaped pattern, formed by the Wadena lobe spreading to the west and south (fig. VII-37). The drumlin field is buried on the western edge by younger till and outwash from the Des Moines lobe.

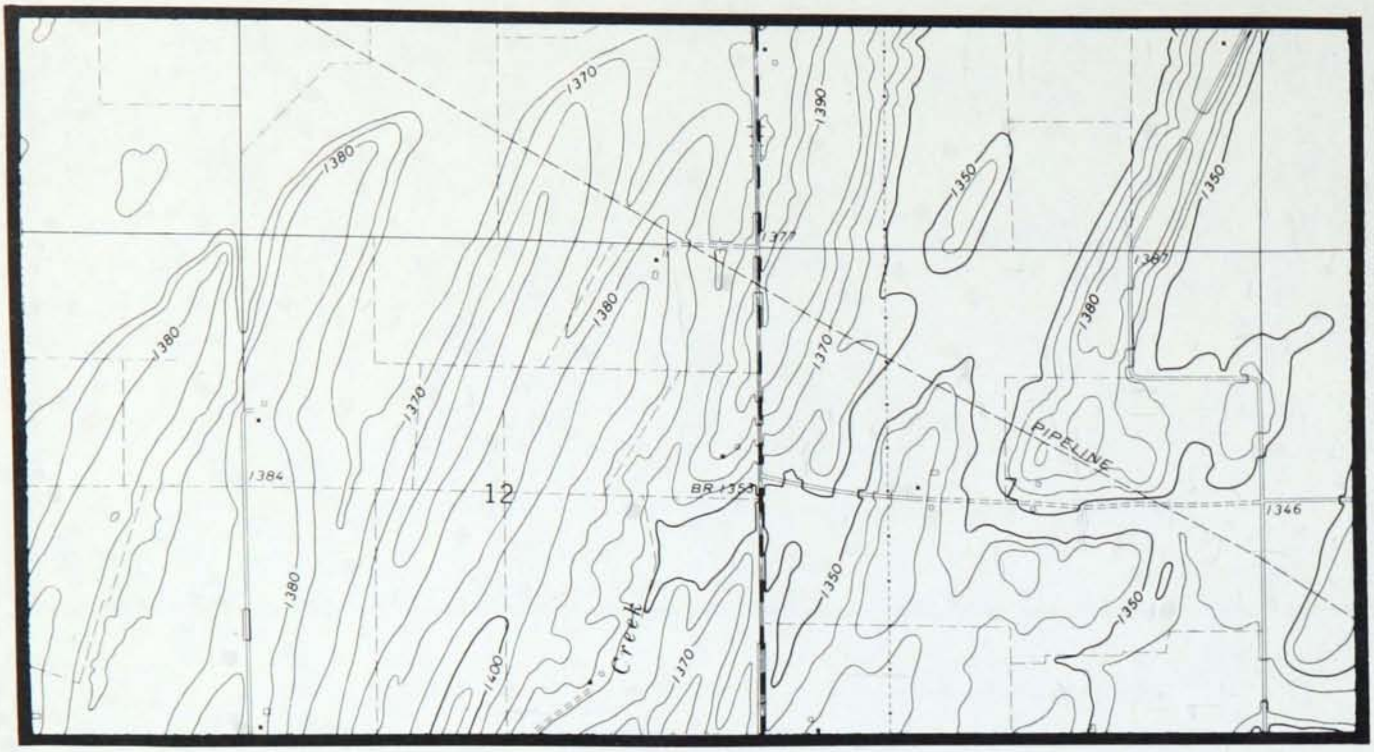


Figure VII-37. Wadena drumlins near Hewitt. Drumlins on northwest were truncated by outwash stream, and swales on east are largely drowned by outwash deposition; Bertha quadrangle (scale 1:24,000; contour interval 10 feet).

The various outwash plains that obscure the Wadena drumlins are partly of different generations. The Park Rapids plain on the north clearly was formed by streams emerging from the Wadena lobe as it stood at the Itasca moraine, and the outwash plain on the east was formed in front of the Rainy lobe at the St. Croix moraine in the same way and at the same time. The combined outwash streams led southward down the Long Prairie River along the front of the St. Croix moraine. This course was subsequently blocked by advance of the Des Moines lobe across the southern part of the Wadena drumlin area. But by this time the Rainy lobe had withdrawn eastward from the St. Croix moraine. The Long Prairie River then reversed its course and the meltwaters escaped through a gap in the St. Croix moraine at Pillager, feeding into the Mississippi River near Brainerd.

As the Des Moines lobe retreated, new infloods of meltwater came from the Des Moines lobe to the west, contributing to the flow through Pillager gap and forming new outwash plains and valley trains, one of which led through another gap in the St. Croix moraine, at Cold Spring west of St. Cloud.

18. ALEXANDRIA MORAINE AREA

The great belt of lake-dotted moraine extending northward in an arc through west-central Minnesota is the Alexandria moraine complex—a complex because it is 10-20 miles broad, is interrupted by extensive areas of outwash, and contains the drifts of two different ice lobes. The bulk of the moraine is believed to have been produced at the terminus of the Wadena lobe, concurrent with formation of the Wadena drumlin field. The moraine was subsequently overridden from the west by the Des Moines lobe.

The Alexandria moraine complex contains the thickest glacial drift in the state and reaches the highest altitudes in western Minnesota—the Leaf Hills are 1,700 feet above sea level. The relief is rugged and the slopes heavily wooded, so the area has much greater value as recreational land than as agricultural land (fig. VII-38).

Northward the moraine complex gives way to the Itasca moraine and the moraines of the Bemidji area. Southeastward it merges with the St. Croix moraine; any boundary drawn in this area is arbitrary, for both moraines were overridden and are partly obscured. The western and southwestern base of the moraine is relatively sharp and straight, and the rise in elevation here is enough so that the base essentially marks the forest/prairie border throughout much of its length.

19. OWATONNA MORAINE AREA

Extending southward from the Minneapolis area to the Iowa border and beyond is a series of moraines that formed along the eastern edge of the Des Moines lobe. These moraines have been traced around the Des Moines lobe to the west side, most recently by Ruhe (1969), where they carry the names Bemis and Altamont, but for the purpose of gross physiographic subdivisions of Minnesota, the moraines on the east side are here grouped as the Owatonna moraine

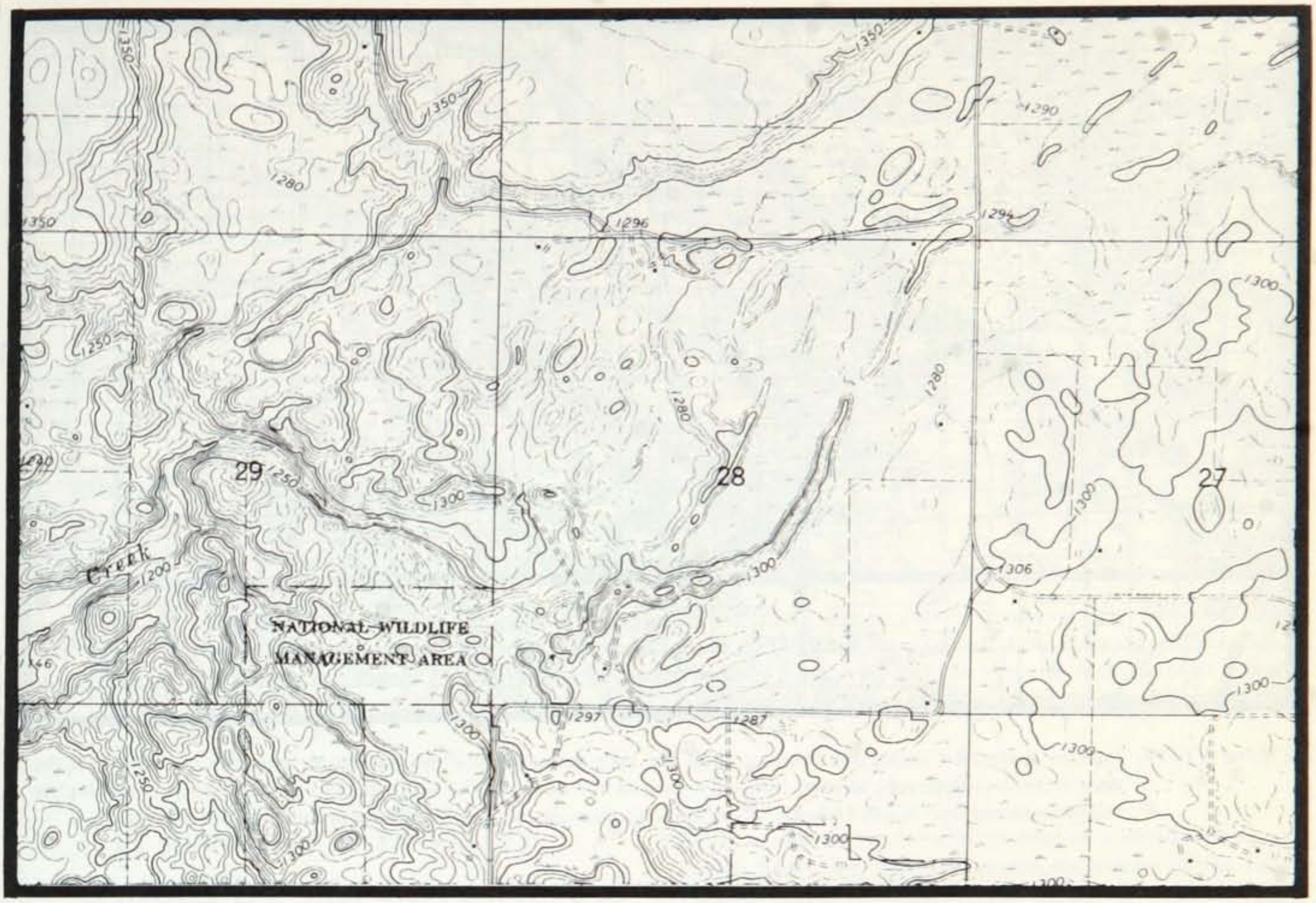


Figure VII-38. Rugged collapse topography in Alexandria moraine in Pope County south of Glenwood. Note eskers in a possible tunnel valley; Lake Johanna quadrangle (scale 1:24,000; contour interval 10 feet).

area. The eastern edge terminates abruptly beside the featureless Rochester till plain. The western edge grades into the Blue Earth till plain.

Relief in this area is rugged in the northern part, which is largely forested—part of the Big Woods. But southward the relief decreases, farming is more widespread today, and the prehistoric vegetation was prairie rather than forest. Thus the northern part of the section carries a peninsula of the Big Woods southward into prairie.

The linear character of the moraine area is emphasized by the presence of some marginal streams such as the Straight River. A prominent transverse valley, filled with ice-block depressions, may record either a pre-Wisconsin bedrock valley or a subglacial erosional valley.

20. COTEAU DES PRAIRIES, OUTER PART

Between the lowland of the Minnesota River and the lowland of the James River in South Dakota is a wedge-shaped upland pointing north—the Coteau des Prairies. This upland has a remarkably straight and steep eastern escarpment, trending southeast. The upland and its scarp have the appearance of a structurally controlled plateau, but no exposures of bedrock have been found along the scarp, and well borings show only several hundred feet of glacial deposits. Nonetheless, it seems most reasonable to postulate some kind of bedrock upland, presumably of Cretaceous sedimentary rocks, that separated the preglacial Minnesota and James River lowlands.

When the Des Moines lobe filled the Minnesota River lowland during Wisconsin glaciation, it rose on its western flank up over the escarpment and onto the crest of the coteau, producing the Bemis moraine at its terminus and the Altamont and other moraines as it withdrew from its most advanced position and retreated down the escarpment (fig. VII-39). This system of linear moraines on the scarp and crest of the prairie coteau gives the appearance of a set of lateral moraines bounding a valley glacier, but the Des Moines lobe was really not like a valley glacier, because flow of the ice here was probably toward the lateral margin rather than as shear along the side.

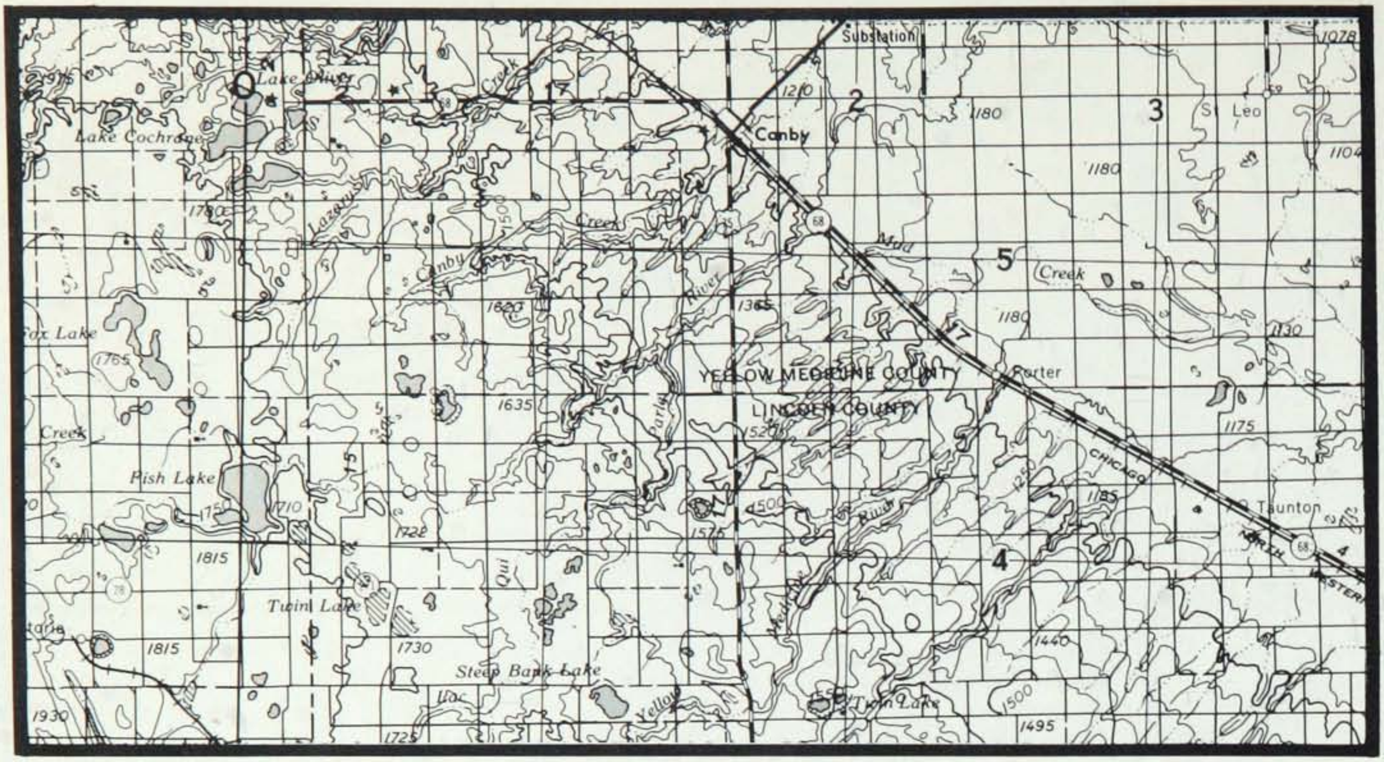


Figure VII-39. Eastern scarp of Coteau des Prairies, with the lake-dotted Bemis and Altamont moraines on the crest; the area is here collectively termed the outer part of the coteau. The scarp becomes much steeper and higher northwestward into South Dakota; Watertown quadrangle (scale 1:250,000; contour interval 50 feet).

The steep scarp of the Coteau des Prairies is marked by numerous gullies that carry patches of deciduous woods, including oak, elm, ash, and basswood. These trees are protected from the strong winds (and formerly the prairie fires) that sweep across the uplands, and they are nourished by meltwater from snow that accumulates there in the winter.

21. BLUE EARTH TILL PLAIN

South of the Minnesota River to the Iowa border, the area covered by the interior part of the Des Moines lobe is a generally featureless till plain (fig. VII-40). The western part, at the foreslope of the Coteau des Prairies escarpment, has a certain linearity that in some cases reflects weak "lateral" moraines formed during shrinkage of the ice lobe; in other cases the lineations are the channels of former ice-marginal meltwater streams. The courses of the Redwood, Cottonwood, and Watonwan Rivers follow these old channels. Linear ridges of gravel up to 3 miles long and 50 feet high are common features of the till plain. Chains of lakes in Martin County probably reflect buried preglacial valleys. The southern part of the area is particularly flat, because this was the region of Glacial Lake Minnesota.

The Blue Earth till plain is the heart of the productive Minnesota cornbelt. It was entirely long-grass prairie before settlement, except for small patches of woodland along lake-basin slopes and on river flood plains.

22. OLIVIA TILL PLAIN

North of the Minnesota River a till plain exists that is comparable to the Blue Earth till plain. The linearity, however, is weaker, at least through most of its area (fig. VII-40), than that of the Blue Earth till plain. In the northwestern part, where it is narrower as it becomes constricted between the Minnesota River Valley and the Alexandria moraine complex, this till plain is crossed obliquely by the Chippewa and Pomme de Terre Rivers. The latter cut a long, relatively straight trench through the plain, perhaps when it carried outflow water from Glacial Lake Pelican in the Alexandria moraine complex southward to the Minnesota River near Watson. Both the Pomme de Terre and the Chippewa Rivers fanned out on sand plains before reaching the Minnesota River, and they built multiple sand-filled channels on the plain before merging with the Minnesota River, which at that time had barely begun its dissection as the Lake Agassiz outlet.

The Olivia till plain extends far to the northwest to form a narrow band between the Alexandria moraine and the Lake Agassiz plain.

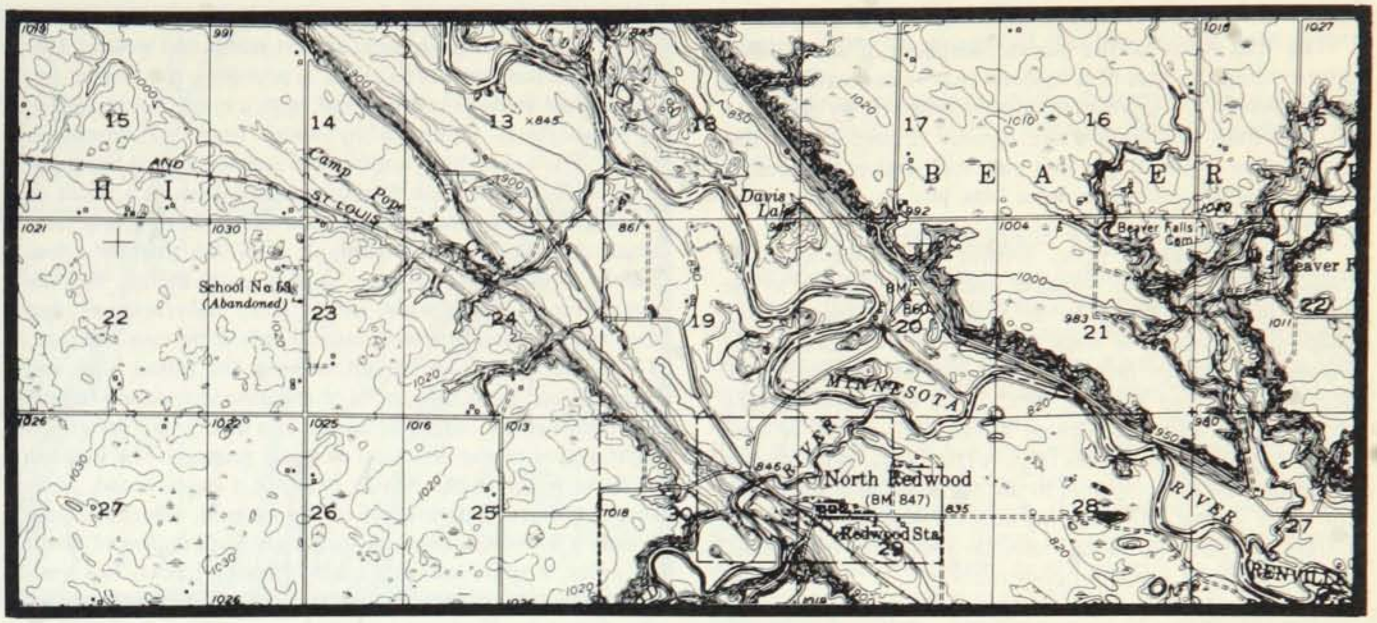


Figure VII-40. Minnesota River Valley near Redwood Falls, with the Olivia till plain to the northeast and the Blue Earth till plain to the southwest. Note the smooth terrace at 990 feet, the steep scarps of the terrace and the valley, the slightly fan-shaped colluvial footslopes at the base of the scarps, and the irregular rock-knobbed center of the valley; Redwood Falls quadrangle (scale 1:62,500; contour interval 10 feet).

23. MINNESOTA RIVER VALLEY

The Minnesota River Valley extends through the till plains in a straight course for 180 miles from its head at Browns Valley to the big bend at Mankato, and then for an additional 40 miles to the moraine complex west of Minneapolis (fig. VII-40). The river itself is grossly underfit, for it flows in a sharp, wide valley cut by the Glacial River Warren, a much larger river that drained Glacial Lake Agassiz. In its upstream portion it is marked by two long river lakes—Lac qui Parle, dammed by a fan from the Chippewa River, and Big Stone Lake, dammed by a fan from the Whetstone River. The River Warren channel continues upstream to the Lake Agassiz plain and is occupied by still another river lake—Lake Traverse, which actually drains northward into the Red River; the fan of the Little Minnesota River at Browns Valley, between Big Stone and Traverse Lakes, forms here the continental divide (Matsch and Wright, 1967).

Downstream from the river lakes and fans, which largely obscure the valley bottom, the Minnesota River Valley shows extensive outcrops of crystalline rocks—a long and narrow window to the Precambrian geology of southwestern Minnesota. These outcrops continue as far as the big bend at Mankato, where Paleozoic rocks appear. The change in bedrock here accounts indirectly for the sharp right-angle bend, which marks the intersection of the Minnesota River lowland (basically in Precambrian rocks with a shallow filling of Cretaceous sediments) and the Minneapolis lowland (which follows the northeastern strike of the soft Cambrian sandstones on the western edge of a broad structural basin). These lowlands had guided the course of the Des Moines lobe as it turned northeast to form the Grantsburg sublobe. The lowlands were not filled by glacial drift, so the incipient glacial Minnesota River during the time of ice retreat simply followed the same general course.

The Minnesota River Valley throughout its length is characterized by glacial outwash terraces along the flanks. For most of the length of the river, these terraces are largely confined to the valley itself, but upstream from Granite Falls there are several subparallel channels beside the main valley, representing river levels either during the time of active outwash deposition as the ice retreated or during the beginning of downcutting by the Lake Agassiz outlet stream, which was not localized to a single channel when it was first formed. Watson Sag is the most conspicuous of the now-abandoned channels. Others were initially occupied by the Pomme de Terre and Chippewa Rivers, which join the Minnesota River in this area.

The Minnesota River Valley is truly the most striking and scenic feature of all of south-central Minnesota. It is a narrow sliver of wooded hill slopes in the vast plains to north and south, and it holds within it a diversity of geologic features such as rugged granite knobs on the valley floor, boulder-gravel river bars, broad sandy terraces, gentle colluvial slopes—and a stream along the axis that is almost tiny in the context of these major features. When the valley was filled from side to side with the Glacial River Warren, as it must have been much of the time in the days of Lake Agassiz, it must have been an impressive feature indeed.

24. GLACIAL LAKE AGASSIZ

The Red River of the North meanders in an intricate pattern along the axis of a lowland that is impressive in its great expanse. The lowland is underlain largely by clays and silts deposited in Glacial Lake Agassiz (Elson, 1967). The combination of flat slopes and clay soils makes the area so poorly drained that agriculture was late to develop here (Warkentin, 1967).

The Lake Agassiz plain, formed as the ice sheet retreated northward into Canada, had its first outlet to the south, into the Glacial River Warren. The outlet river eroded progressively in steps through the low moraine that bounds the lake on the south, forming boulder-paved terraces (Matsch and Wright, 1967). The lake level thereby fell in stages, and wind-driven waves built ridges of sand and gravel along the shores (figs. VII-41, 42). With further ice retreat, the outlet shifted to the northeast and north, and still lower strand lines were formed around the basin. The strand lines serve as natural sites for roads and farmsteads, and as sources of gravel for construction.

The central part of the Red River plain is underlain by clays and silts, formed in water 300 feet deep. Toward the marginal beaches sand was generally deposited, but in some areas in the shallow water the waves were apparently strong enough so that no sediment was deposited at all, and glacial till is exposed on the surface.

25. BELTRAMI ARM OF LAKE AGASSIZ

Glacial Lake Agassiz, which left its principal mark in Minnesota in the flat expanses of the Red River Valley, extended in its early stages eastward across northern Minnesota almost as far as Ely. Most of this area is a vast wetland, locally interrupted on the east by unburied moraines and on the north by the dissecting tributaries of the Rainy River. The central, undissected part north and west of Upper Red Lake constitutes what is probably the largest uninterrupted wetland in the world, with a magnificent display of raised bogs separated by water tracks (Heinselman, 1963). The raised bogs are tree-covered areas that are topographically higher than the surrounding fens, as a result of rapid growth of sphagnum mosses, heath shrubs, and tamarack and black spruce trees, all of which tolerate a low supply of mineral nutrients. As they grow higher, they lie out of range of the water inflow from mineral areas, and they receive their nutrients instead only from rain and dust. The water tracks, which run between the raised bogs and "streamline" them into ovoid or teardrop shapes, are largely fens, marked by sedges arranged in a string-like pattern at right angles to the direction of water seepage. The greatest of these water tracks, which is about 4 miles broad, leads eastward from the west end of the wetland, where the peat is only a foot or so thick. After an eastward course of about 15 miles, it splits into two gently curving tracks, one leading south toward Upper Red Lake, the other north toward the Rapid River. In the angle between them is the large complex of raised bogs previously mentioned.

26. COTEAU DES PRAIRIES, INNER PART

In the southwestern corner of Minnesota beyond the Bemis moraine of Wisconsin age is a small triangle of drift largely covered with loess. The loess, which buries drifts of both Wisconsin and pre-Wisconsin age, thickens toward the southwest; it probably originated as wind-blown silt from the outwash deposits of the Big Sioux River. The plain is characterized by a well developed drainage system, and thus by the absence of depressions. Major streams, for example

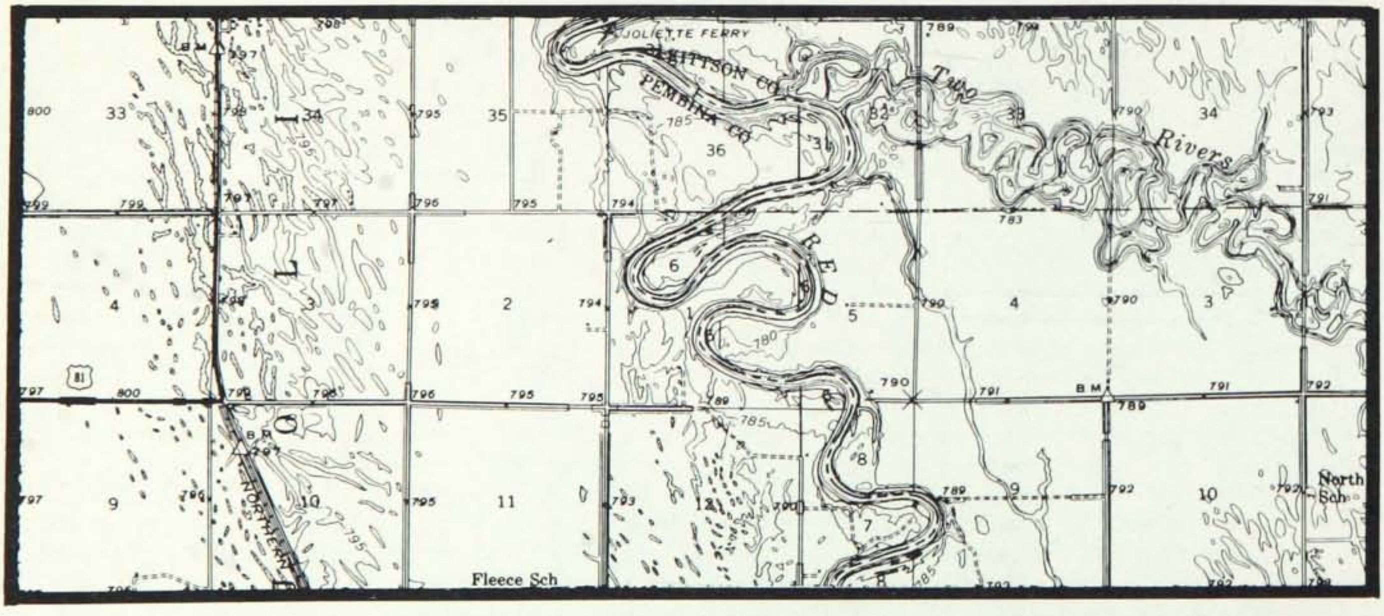


Figure VII-41. Glacial Lake Agassiz plain in Kittson County, northwestern Minnesota, and in adjacent North Dakota. Note intricately incised meanders of the Red River and its tributary. Linear pattern of 5-foot contours represents the micro-relief variously ascribed to frost-action, ice-floe tracks, or reflection of bedrock fractures; Pembina quadrangle (scale 1:62,500).

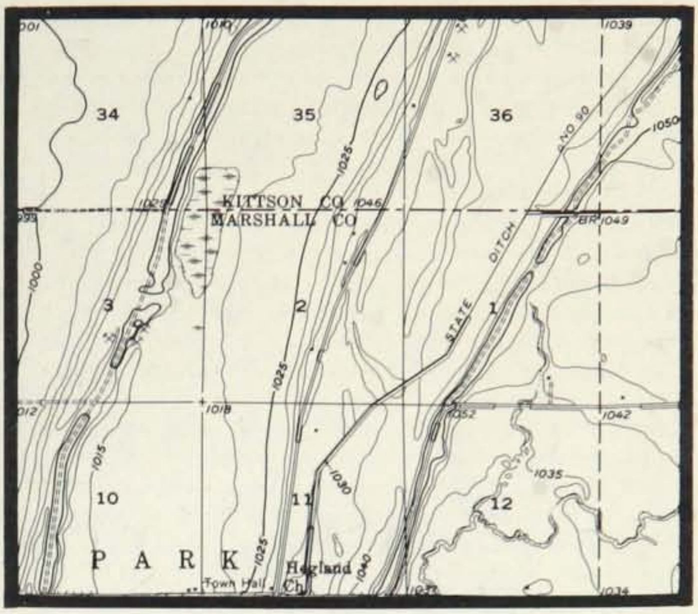


Figure VII-42. Three Lake Agassiz beaches south of Karlstad, Kittson County, northwestern Minnesota. Beach crests at 1,055, 1,045, and 1,030 feet. Note beach pond (now a marsh) behind the 1,030-foot beach. Also note the roads and gravel pits along the beaches; Karlstad quadrangle (scale 1:62,500; contour interval 5 feet).

the Flandreau, Rock, and Kanaranzi, carried meltwater and sediment from the margin of the Des Moines lobe across the region.

Numerous outcrops of Sioux Quartzite bear the polish and striations of several ice advances from different directions.

27. ROCHESTER TILL PLAIN

In southeastern Minnesota beyond the Wisconsin moraines of the Owatonna area is a nearly featureless pre-Wisconsin till plain with a partial cover of loess that thickens eastward toward the Mississippi River (fig. VII-43). The eastern part of the area is deeply dissected by tributaries of the Mississippi—the Cannon, Zumbro, Whitewater, and Root Rivers. The till has been mapped as Iowan, but Ruhe (1969) has shown in adjacent Iowa that Iowan till is probably Kansan till from which the weathering profile was largely removed before burial by loess. In the eastern part of the plain the loess seems to rest directly on the bedrock, although few exposures exist, and this area has often been mapped as "driftless," like the portion of southwestern Wisconsin directly across the Mississippi River.

The area of thick loess takes on a slight relief of its own, resulting from local deposition of loess in long hill-like forms (paha). The loess presumably dates from the main Wisconsin glaciation, having been derived from the Mississippi River outwash or from tributary outwash plains (Foss and Rust, 1962). The upland of till and loess is intensively farmed.

Figure VII-43. Rochester till plain north of Rochester, locally dissected by the South Fork Zumbro River. Paleozoic bedrock is overlain by pre-Wisconsin till and thin loess; Rochester quadrangle (scale 1:62,500; contour interval 20 feet).

The deep dissecting valleys give the eastern edge of the area almost a mountainous aspect, with a relief of 500 feet. Flat-lying Paleozoic sedimentary rocks crop out in the valley slopes or are mantled thickly with block-filled colluvium or with loess. The valley floors are generally flat and cleared of forest for farming. They abut abruptly against the heavily forested hill slopes. Downstream the flat floors are underlain in many cases by laminated silts and fine sands, deposited in backwater lakes that extended upstream from the Mississippi River valley for several miles during the time of deposition of glacial outwash in the main valley. This depositional flooding even topped some low divides between tributaries, and in the Mississippi valley itself it caused the separation of large islands of upland, as at Frontenac and Red Wing.

The old valley floors now stand as terraces, for the depositional cycle ended when Glacial Lake Agassiz formed and its outlet stream (River Warren) began an epoch of downcutting. This cycle of dissection extended several miles up the tributaries from the Mississippi River, but it was brought to an end by the reversal in base level, as the Lake Agassiz outlet shifted to the north, and the Mississippi River once again began to alluviate. In the smaller tributaries an epicycle of cut and fill can be related to soil erosion accompanying agricultural land clearance. In places the well developed alluvial soil of the main valley floor is overlain by a few feet of fresh sediment eroded from cultivated hill slopes.

The Mississippi River flood plain today is a complex of channels, lakes, and marshes. Dams for flood-control and navigation have produced some water pools, and dredging has changed the river channels, so the river is hardly in a natural condition today. But the basic features of the flood plain are not completely obscured. Lake Pepin results from fan deposition by the Chippewa River. Practically all other tributary streams have also placed at least a partial dam across the flood plain, either blocking the river or pushing it over to the far side. Many of the river towns are built on such fans, including Winona and Lake City. The river channel itself, which is commonly multiple, is bordered by tree-covered natural levees that show as a pair of ribbons separating the channel from the backswamp, which itself may contain lakes.

Chapter VIII

GRAVITY AND MAGNETICS

REGIONAL GRAVITY FIELD

P. K. Sims

Gravity surveys have proved useful in the study of the geology of Minnesota, particularly when used in conjunction with magnetic data and, where available, seismic data. In this stable region, gravity anomalies having steep gradients are caused by lateral density differences in the rocks of the upper crust. Accordingly, local anomalies mainly reflect compositional differences in the Precambrian rocks at or near the surface, and gravity maps provide a basis for projecting the boundaries of these rocks from localities of surface outcrops into areas covered by Phanerozoic deposits.

Gravity surveys have been carried out intermittently in Minnesota since early in the century, and the history of these is discussed by Craddock and others (1970) in the text that accompanies the Bouguer gravity map of Minnesota and northwestern Wisconsin. Systematic gravity surveys in the state were started in the early 1960's, and are being continued on a one- or two-mile grid network. The maps are being published at a scale of 1:250,000. The status of the detailed surveys is shown in Figure VIII-1. These surveys are intended to aid in the delineation of geologic units as well as to contribute to a better understanding of the tectonic history.

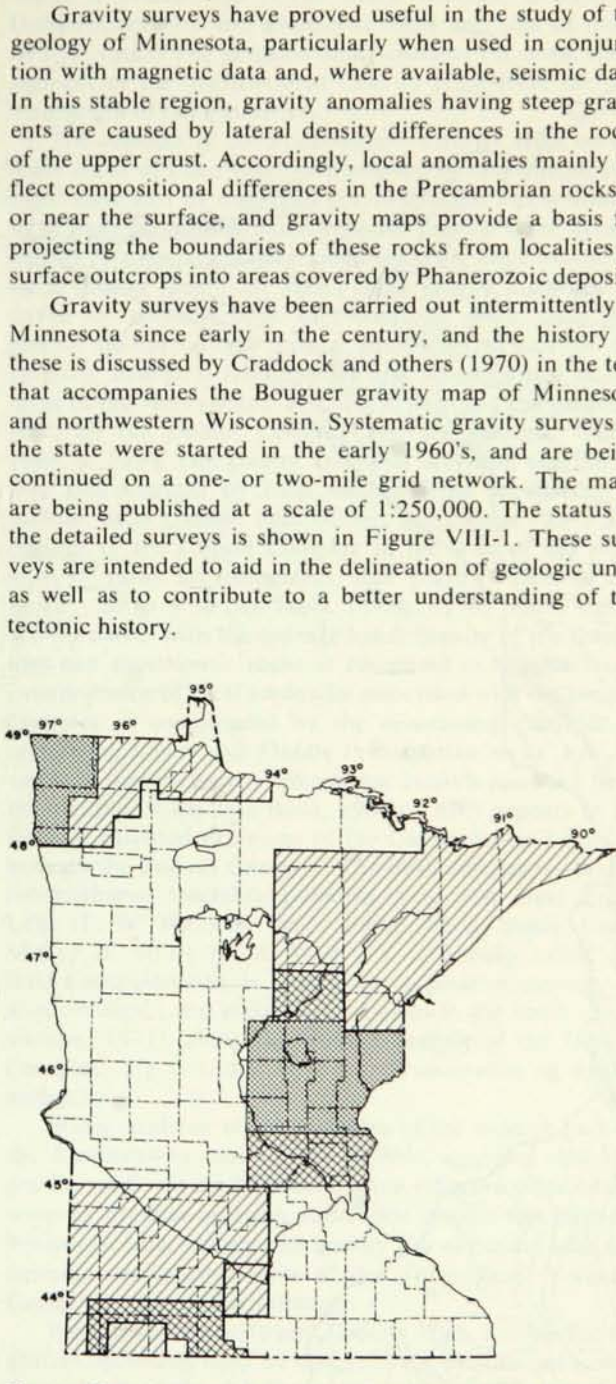


Figure VIII-1. Status of gravity surveying in Minnesota (Lined area surveyed 1966-69; stippled area surveyed 1970; cross-ruled area surveyed 1971).

In Minnesota, the mean altitude is about 1,100 feet above sea level, and according to Woollard's (1959) empirical curve that relates the Bouguer anomaly to surface altitude, the approximate predicted or normal Bouguer anomaly for the region is about -45 milligals. According to the theoretical estimates calculated from the plate equation, the normal Bouguer anomaly for a mean elevation of 1,100 feet is -37.5 milligals (L. D. McGinnis, 1971, written comm.). Anomalies that are significantly higher or lower than the predicted anomaly suggest the presence of relatively high- or low-density rocks in the underlying Precambrian sequence.

The most prominent gravity anomaly in Minnesota is the Midcontinent Gravity High, a broad, positive feature (fig. VIII-2) that trends northward across the eastern counties into Wisconsin and Michigan (Thiel, 1956; Craddock and others, 1963) and extends southwestward across Iowa into Kansas (Woollard and Joesting, 1964; Lidiak, 1964; Goldich and others, 1966, p. 5405). Because of the tectonic significance of this feature, it has been studied by a variety of geophysical methods, and the broad aspects of the geology now are known (see section on Late Precambrian, this volume).

South of Lake Superior, the Midcontinent Gravity High is 10 to 40 miles wide and is marked along its length by conspicuous gravity highs and intervening saddles. The maximum Bouguer gravity values (+50 mgal) on the feature are south of Duluth and southwest of Minneapolis-St. Paul. Along most of its length, the Midcontinent Gravity High is flanked by linear gravity lows, which reach a minimum of -89 mgal adjacent to the Wisconsin line. It has been shown (Thiel, 1956; Craddock and others, 1963) that the gravity high is caused by a raised block of thick Upper Precambrian mafic lavas, termed the St. Croix horst, and the flanking gravity lows reflect thick successions of downdropped Upper Precambrian sedimentary deposits. The lines of inflection along the steep gravity gradients that separate the horst from the flanking basins coincide with steeply-dipping faults that have vertical displacements of as much as several thousand feet (Craddock and others, 1970; Mooney and others, 1970a and b). Two significant bends in the anomaly are apparent on Figure VIII-2. One of these is south of Minneapolis-St. Paul, and possibly is caused by an older Precambrian fault that controlled the boundary of the horst (G. B. Morey, 1972, oral comm.). A northwest-trending fault that offsets Paleozoic strata near Belle Plaine (Sloan and Danes, 1962), along the Minnesota River, lies a few miles to the northeast of the inferred basement fault. Another, lesser bend northeast of the Twin Cities probably is related to northwest-trending, sinistral faults that offset the horst (G. B. Morey, 1971, oral comm.).

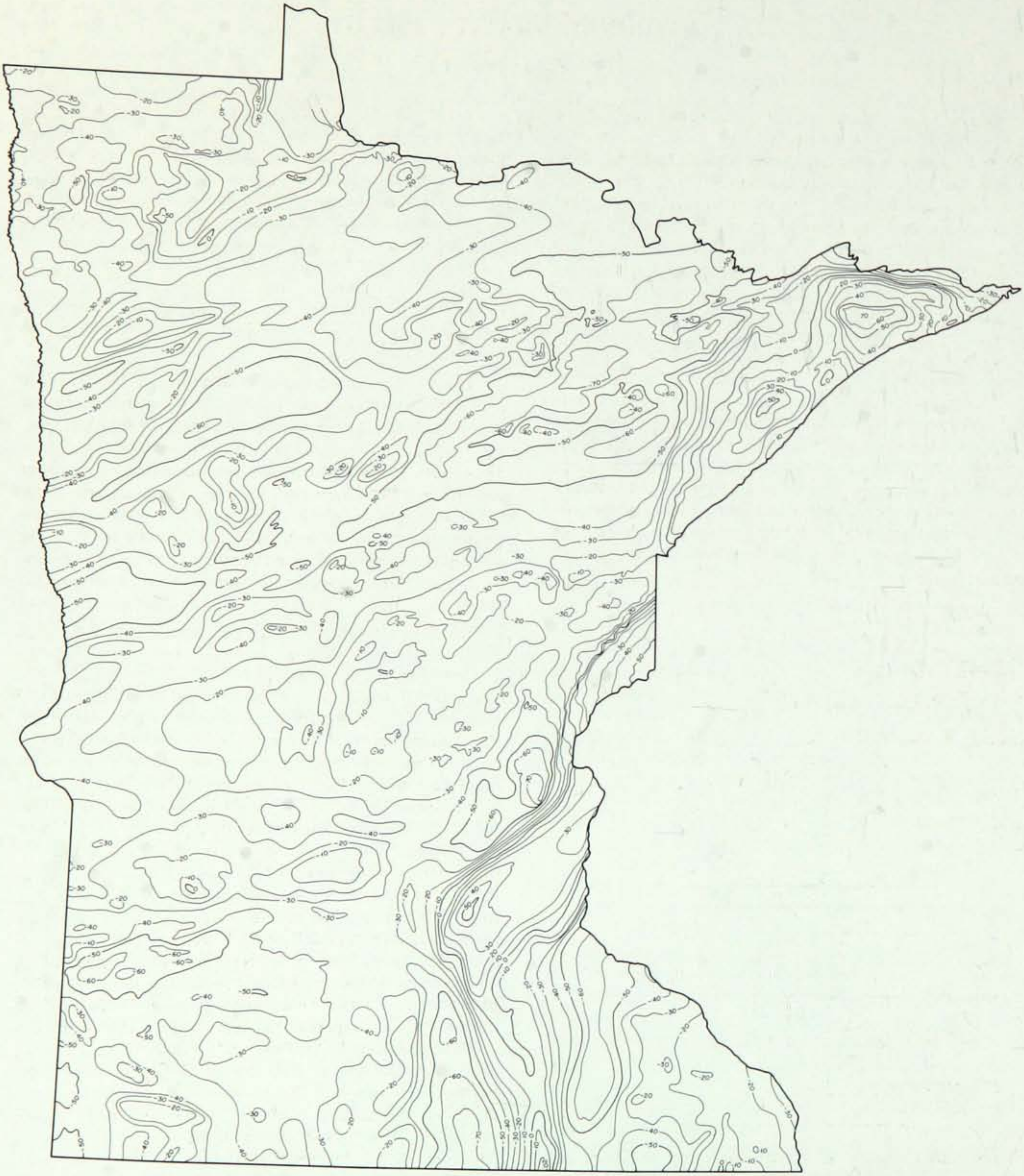


Figure VIII-2. Simple Bouguer gravity map of Minnesota, compiled by Robert Ayers and P. K. Sims in 1972 from various sources including published reports and open file maps by R. J. Ikola and L. D. McGinnis.

The gravity high bifurcates in northwestern Wisconsin, south of Lake Superior (Thiel, 1956; Craddock and others, 1970; White, 1966), where it gives way to a nearly circular gravity low centered on the Bayfield Peninsula. The southern branch follows an outcrop belt of Middle Keweenawan mafic igneous rocks, including both lava flows and gabbroic intrusive rocks, eastward into Michigan. The northern branch extends along the north shore of Lake Superior, where it overlies the North Shore Volcanic Group, the Duluth Complex, and associated dikes and sills of mafic igneous rocks. The Douglas fault separates the north shore segment of the Midcontinent Gravity High from the segment south of Lake Superior.

The segment of the Midcontinent Gravity High on the north shore of Lake Superior is marked by two conspicuous highs separated by a marked gravity trough. The northernmost high has a maximum Bouguer gravity value greater than +70 mgal; the southernmost anomaly has a maximum value greater than +50 mgal. These gravity highs are inferred to represent thick successions of mafic lavas, but precise interpretations remain equivocal because of the complex structural relationships and intertonguing of the lavas of the North Shore Volcanic Group and the gabbroic rocks of the Duluth Complex (see for example White, 1966, p. E19). The gravity low, west of Tofte, overlies an area characterized by large inclusions of anorthosite in diabase, as at Carlton Peak (Grout and others, 1959). Exposures of the Duluth Complex to the west of the main mass of flows are coincident with gravity values in the range −20 to +30 (see Ikola, 1968b, 1970). These values are consistent with the average lower density of the troctolites and anorthositic rocks as compared to basaltic lavas. Interpretation of local anomalies associated with the Duluth Complex is complicated by the contrasting densities of underlying Lower and Middle Precambrian rocks. For example, a narrow gravity trough that extends eastward from the vicinity of Ely (see Ikola, 1968b, 1970) appears to reflect an eastward extension of the Giants Range batholith beneath the Duluth Complex. The differentiated, probably funnel-shaped troctolitic-gabbroic body near Bald Eagle Lake (P. W. Weiblen, 1965, unpub. Ph.D. thesis, Univ. Minn.) (T. 60-61 N., R. 9-10 W.), informally called the Bald Eagle intrusion, is reflected by a positive anomaly of about 8 mgal.; and small dunite bodies to the south (Bonnichsen, 1971), along the western margin of the Duluth Complex, are reflected by positive anomalies of a few milligals.

In his synthesis of the tectonics of the western part of the Keweenawan basin, White (1966) suggested that the gravity saddle on the Minnesota shore reflects a ridge or upwarp of pre-Keweenawan rocks that crosses the Bayfield Peninsula. Interestingly, the gravity low coincides with uppermost stratigraphic units of the North Shore Volcanic Group (see Green, this volume).

West of the Midcontinent Gravity High, sources for the gravity anomalies must be sought in the Middle and Lower Precambrian rocks. A belt of moderately high gravity values about 75 miles wide, which trends west-southwestward from the edge of the Keweenawan province at the southwest tip of Lake Superior, coincides with Middle Precambrian rocks (see Morey, this volume). The higher Bouguer values in this belt coincide with outcrops of the Thomson Formation in Carlton County and with possibly equivalent biotite schists in an area north of St. Cloud. Judged from available density data (Mooney and Bleifuss, 1953, table 2), these rocks have an average density 0.1 to 0.2 gm/cm^3 greater than other rocks in the Middle Precambrian stratigraphic succession, except perhaps the unaltered iron-formations of the Cuyuna district, and can account for the higher anomalies. The intrusive rocks of the Central Minnesota batholith coincide with somewhat lower Bouguer gravity values, which is consistent with their having an average lower density than the graywacke-slate succession. It is of interest to note, however, that these intrusive rocks do not cause conspicuous gravity lows of the same order of magnitude as do the batholithic masses of Early Precambrian age.

The remainder of Minnesota to the west of the Midcontinent Gravity High is characterized by irregular gravity highs and lows that reflect lateral density differences in the Lower Precambrian rocks. The three major rock types in this terrane—mafic volcanic rocks of Ely Greenstone-type, graywacke-slate and associated dacitic pyroclastic deposits of Knife Lake-type, and granitic rocks of the Algoman batholiths—have markedly different average densities. The granitic rocks have an average density about 0.3 gm/cm^3 less than the mafic metavolcanic rocks and about 0.1 gm/cm^3 less than Knife Lake-type strata. These rock-density differences together with the universally steep dips of the rocks, which accentuate mass differences, result in distinctive anomalies. The anomalies trend northeastward in northern Minnesota and eastward in southwestern Minnesota, in general agreement with the average structural trends of the Lower Precambrian rocks in the respective regions.

The most conspicuous anomalies in the Lower Precambrian terranes are the lows over the major Algoman batholiths. The prominent negative anomaly in northern Minnesota, which extends from the vicinity of Ely southwestward for a distance of 150 miles (Ikola, 1968a), coincides with the Giants Range batholith. North of Hibbing and Virginia it attains a minimum Bouguer gravity value of −75 mgal, and throughout most of its length it is less than −60 mgal. At its eastern end, the gravity effect of the granitic rocks is "damped out" by the mass effect of the Duluth Complex (Ikola, 1970; Craddock and others, 1970). The gravity low in the vicinity of the Red Lakes also is known to coincide with a granitic body, and correspondingly low values in western Minnesota are presumed to overlie sizeable granitic bodies, as shown on the state geologic map (pl. 1, this volume). In southwestern Minnesota, the strongly negative (−60 mgal) anomaly centered near Marshall and extending northeastward to the Minnesota River Valley reflects a large body of granitic rocks (the Sacred Heart Granite of Lund, 1956).

The Vermilion batholith in northern Minnesota, termed the Vermilion granite-migmatite massif by Southwick (this volume), because of the vast amount of metavolcanic and metasedimentary rocks within it, does not produce conspicuously low Bouguer gravity values. Apparently the average density of the rocks constituting the massif is not significantly less than that of adjacent metamorphosed graywacke-slate.

The gravity highs in the Lower Precambrian terrane of northern Minnesota coincide with mafic metavolcanic rocks, such as those that make up the Ely Greenstone in the Vermilion district. These rocks cause Bouguer gravity anomalies in the range −40 to +5 mgal. The anomalies are accentuated at places by the presence of dense ultramafic bodies, as in the vicinity of Deer Lake, 30 miles west of Cook (Sims and others, 1970).

Intermediate gravity values characterize the rocks of Knife Lake-type and large-scale intercalations of granite and greenstone. Because of the similarity in densities and magnetic susceptibilities of graywacke-slate and felsic (dacitic) volcanic strata, these rock types cannot be distinguished separately by gravimetric and magnetic data.

The cause of many gravity anomalies south of latitude 46° N. in southwestern Minnesota is uncertain because of the sparse exposures and fragmentary knowledge of the geology. It is inferred, however (pl. 1, this volume), that many of the common rock types that characterize the Early Precambrian of northern Minnesota also occur in southwestern Minnesota. The apparent differences in the patterns of gravity (and magnetic) anomalies in the two regions can be accounted for perhaps by differences in structural fabric and metamorphic grade.

Additional gravity data in parts of the state not now covered adequately will further increase our knowledge of its major geologic features. In addition to solving local problems of economic or geologic importance, detailed gravity surveys are needed to delineate the outlines of the principal geologic rock bodies in areas of thick cover and to aid in solving such problems as the nature and location of the boundary between the 2,700 m.y.-old terrane of northern Minnesota and the still older terrane of southwestern Minnesota. Advancement in knowledge of the geology of the basement rocks in Minnesota will proceed hand in hand with the continued application of geophysical techniques.

MAGNETIC DATA AND REGIONAL MAGNETIC PATTERNS

P. K. Sims

Knowledge of the geology of Minnesota has been enhanced substantially by information obtained from magnetic surveys. Prior to development of the airborne magnetometer, dip-needle surveys delineated the major iron-formations of the state and disclosed several strong magnetic anomalies in areas where the bedrock is concealed by thick glacial drift, as can be seen by reference to the 1932 edition of the state geologic map (Grout and others, 1932). From 1947 to 1966, the U.S. Geological Survey in cooperation with the Minnesota Geological Survey carried out aeromagnetic surveys in the state. The separate surveys were published by the U.S. Geological Survey at scales of 1:250,000 or greater (fig. VIII-3), and later were recompiled (Zietz and Kirby, 1970) at a scale of 1:1,000,000. In conjunction with the airborne mapping, magnetic properties of the Precambrian rocks were determined (Mooney, 1952; Mooney and Bleifuss, 1953; Bath, 1960, 1962; Jahren, 1963, 1965; Bath and others, 1971) to aid in interpreting the magnetic anomalies. The aeromagnetic maps and data together with existing gravity maps were used extensively in preparing the 1970 state geologic map (pl. 1, this volume).

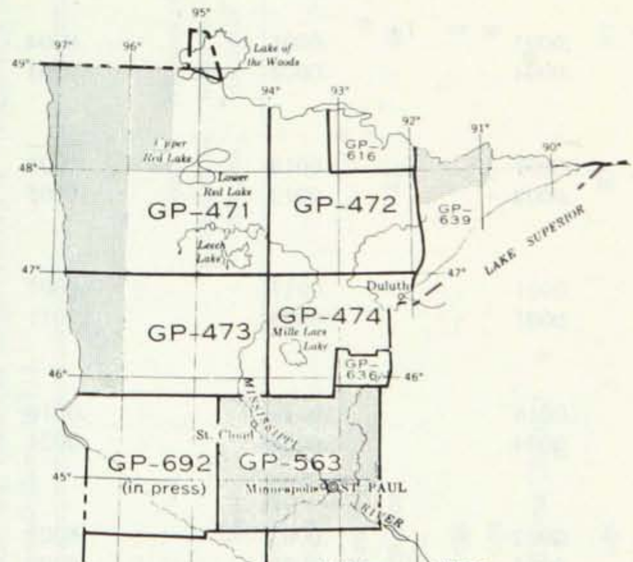


Figure VIII-3. Index to published aeromagnetic maps of Minnesota (Aeromagnetic map of Minnesota: Zietz and Kirby, 1970 is published in color at scale of 1:1,000,000).

MAGNETIC DATA ON PRECAMBRIAN ROCKS

Data on the magnetic properties of most of the common Precambrian rocks in Minnesota are given in Tables VIII-1 and VIII-2. Table VIII-1 summarizes measurements made by Mooney and Bleifuss (1953) on many important rock types; additional data on the rocks, including density and magnetite and ilmenite content, were included in the published report. To facilitate comparison of the data with those in Table VIII-2, the magnetic susceptibility measurements of Mooney and Bleifuss were converted to induced magnetization by G. D. Bath of the U.S. Geological Survey. Table VIII-2, compiled from data of Jahren (1965) and Bath and others (1971), includes measurements of remanent magnetization as well as magnetic susceptibility. As the iron-formations and to a lesser extent other rock units have both magnetic and nonmagnetic facies, the measurements were restricted to those rock samples having a total magnetization greater than 0.0001 gauss. The magnetic properties of three important rock types having moderate magnetizations—serpentinized peridotite, lamprophyres, and the Linden syenite of Grout, 1926—have not been determined quantitatively.

As a generalization, the iron-formations have total magnetizations (Bath, 1962) that are an order of magnitude greater than magnetic mafic igneous rocks of Late Precambrian age and two orders of magnitude greater than magnetic phases of the granitic rocks of Early and Middle Precambrian age.

The iron-formations and the Upper Precambrian mafic igneous rocks have a dominant remanent magnetization and a wide range of average total magnetization. The direction of remanent magnetization is along the bedding for both the gently-dipping Biwabik Iron-formation and the steeply-dipping Lower Precambrian banded iron-formations (Jahren, 1963) and along the fossil geomagnetic field (Jahren, 1965) for the Upper Precambrian igneous rocks. For the Upper Precambrian rocks that were measured, the remanent magnetization has an approximate mean declination of azimuth 290° and a downward inclination of about 40° (fig. VIII-4), which conforms generally to the mean direction of remanent magnetization determined for Middle Keweenawan rocks in the Lake Superior region by Dubois (1962) and others (Beck, 1970; Beck and Lindsley, 1969; Books, 1968; and Palmer, 1970). The same investigators have shown that some of the rocks of Late Precambrian age have a steep upward or reversed magnetization, which is nearly opposite to the present geomagnetic field. In Minnesota, the basal 5,000 feet of mafic lavas (North Shore Volcanic Group) in the Grand Portage area, possibly the thin wedge of basaltic lavas west of Duluth that

Table VIII-1. Induced magnetization of Precambrian rocks in Minnesota (calculated by G. D. Bath from magnetic susceptibility data of Mooney and Bleifuss, 1953).

Rock	Total number of samples	Number of magnetic[1] samples	Induced magnetization[2] of magnetic samples, in gauss: Minimum value	Maximum value	Average value
Soudan Iron-formation					
Large sample[3]	3	3	.025	.069	.051
Small sample	3	3	.031	.069	.058
Biwabik Iron-formation					
Large sample	2	2	.033	.034	.034
Small sample	2	2	.038	.055	.046
Upper Precambrian rocks					
Basalt flows					
Large sample	40	31	.0001	.0058	.0022
Small sample	40	30	.0002	.0050	.0022
Rhyolite flows					
Large sample	5	4	.0004	.0018	.0008
Small sample	5	3	.0003	.0004	.0004
Duluth Complex					
Large sample	36	34	.0001	.0014	.0006
Small sample	36	32	.0001	.0036	.0007
Diabase					
Large sample	18	18	.0008	.0031	.0016
Small sample	17	17	.0005	.0072	.0023
Granophyre					
Large sample	8	8	.0001	.0044	.0014
Small sample	8	7	.0007	.0057	.0020
Granodiorite					
Large sample	3	2	.0003	.0005	.0004
Small sample	3	2	.0004	.0006	.0005
Igneous rock (Pre-Keweenawan)					
Giants Range Granite					
Large sample	16	8	.0001	.0018	.0005
Small sample	16	7	.0002	.0012	.0005
McGrath Gneiss of Woyski (1949)					
Large sample	7	5	.0001	.0011	.0005
Small sample	7	4	.0001	.0026	.0011
Stearns Magma series of Woyski (1949)					
Large sample	8	1[4]	.0016	.0016	.0016
Small sample	8	1[4]	.0024	.0024	.0024
Ely Greenstone of Grout and others (1951)					
Large sample	14	1	.0002	.0002	.0002
Small sample	14	1	.0005	.0005	.0005

[1] Induced magnetization > .0001 gauss
[2] Magnetic susceptibilities (K) as measured by Mooney and Bleifuss (1953) are converted to induced magnetization (M_i) by the equation $M_i = \kappa T$, where the intensity of the earth's magnetic field (T) is taken as 0.600 oersteds
[3] See Mooney (1952) for method used in determining susceptibilities
[4] Rock type of magnetic sample not known

Table VIII-2. Induced and remanent magnetizations of Precambrian rocks in Minnesota (determined by measurements on cylindrical samples; compiled by Bath and others, 1971).

Rock	Total number of samples	Number of magnetic[1] samples	Induced magnetization[2] of magnetic samples (in gauss)			Remanent magnetization of magnetic samples (in gauss)			Ratio of average remanent to average induced magnetization
			Minimum value	Maximum value	Average value	Minimum value	Maximum value	Average[5] value	
Biwabik Iron-formation									
Lower Cherty Member									
Virginia Horn area	41	41	.0070[3]	.070[3]	.034[3]	.019	1.11	.29	8.5
Ridge area									
K > .01 cgs	8	8	.0068[3]	.14[3]	.041[3]	.030	1.88	.65	16.
K < .01 cgs	6	6	.0030[3]	.0050[3]	.0040[3]	.0091	.088	.039	9.8
Members above Lower Cherty									
Ridge area									
K > .01 cgs	30	30	.0061[3]	.070[3]	.036[3]	.0068	3.38	1.08	30.
K < .01 cgs	9	9	.0018[3]	.0056[3]	.0042[3]	.0027	.51	.073	17.
Argo Lake area	11	11	.011[3]	.18[3]	.073[3]	.0037	.65	.14	1.9
Banded Iron-formation									
Vermilion district	39	39	.0084[4]	.23[4]	.054[4]	.0013	1.51	.15	2.8
Bear Lake area	16	16	.0080[4]	.26[4]	.10[4]	.017	.19	.089	.9
Upper Precambrian igneous rocks									
Diabase									
Remanent intensity > .01 gauss	27	27	.0003	.0042	.0016	.010	.082	.030	19.
Remanent intensity < .01 gauss	22	22	.0001	.0083	.0021	.0002	.0096	.0045	2.1
Basalt flows (Duluth area)									
Lower sequence	24	18		.0019	.0003		.0087	.0016	5.3
Upper sequence	22	22	.0004	.0032	.0018	.0017	.033	.011	6.1
Duluth Complex (Duluth area)	63	63	.0001	.0017	.0004	.0004	.017	.0028	7.0
Bardon's Peak area only	42	42		.0032	.0004	.0001	.064	.0069	17.
Granophyre	10	10	.0001	.0041	.0026	.0007	.0044	.0024	.9
Granodiorite	2	2	.0015	.0074	.0044	.0012	.0042	.0027	.6

Table VIII-2. Continued. Induced and remanent magnetizations of Precambrian rocks in Minnesota (determined by measurements on cylindrical samples; compiled by Bath and others, 1971).

Rock	Total number of samples	Number of magnetic[1] samples	Induced magnetization[2] of magnetic samples (in gauss)			Remanent magnetization of magnetic samples (in gauss)			Ratio of average remanent to average induced magnetization
			Minimum value	Maximum value	Average value	Minimum value	Maximum value	Average[5] value	
Igneous rocks of intermediate composition									
Tonalite (Giants Range batholith) in Birch Lake area	32	32	.0005	.0022	.0012		.0038	.0008	.7
Snowbank Lake stock	39	39	.0001	.0032	.0011		.0013	.0005	.4
Mafic phase (actinolitic gabbro) of Warman Quartz									
Monzonite of Woyski (1949) in Mora area	4	4	.0009	.0011	.0010				
Hillman tonalite of Woyski (1949) in Rum River area	5	5	.0007	.0020	.0014				
Freedhem Tonalite of Woyski (1949)	19	11	.0005	.0033	.0017				
Igneous rocks of felsic composition									
Giants Range Granite	51	23		.0018	.0004		.0005	.0002	.5
McGrath Gneiss of Woyski (1949)	4	4	.0001	.0006	.0004				
Ely Greenstone of Grout and others (1951), including wall rock of Soudan Iron-formation	46	14		.0003	.0001	.0001	.0056	.0016	16.

[1] Arbitrarily defined as total magnetization > .0001 gauss
[2] Geomagnetic field = .600 oersteds
[3] Measured across the layers
[4] Measured along the layers
[5] Assumes a constant direction
K = magnetic susceptibility

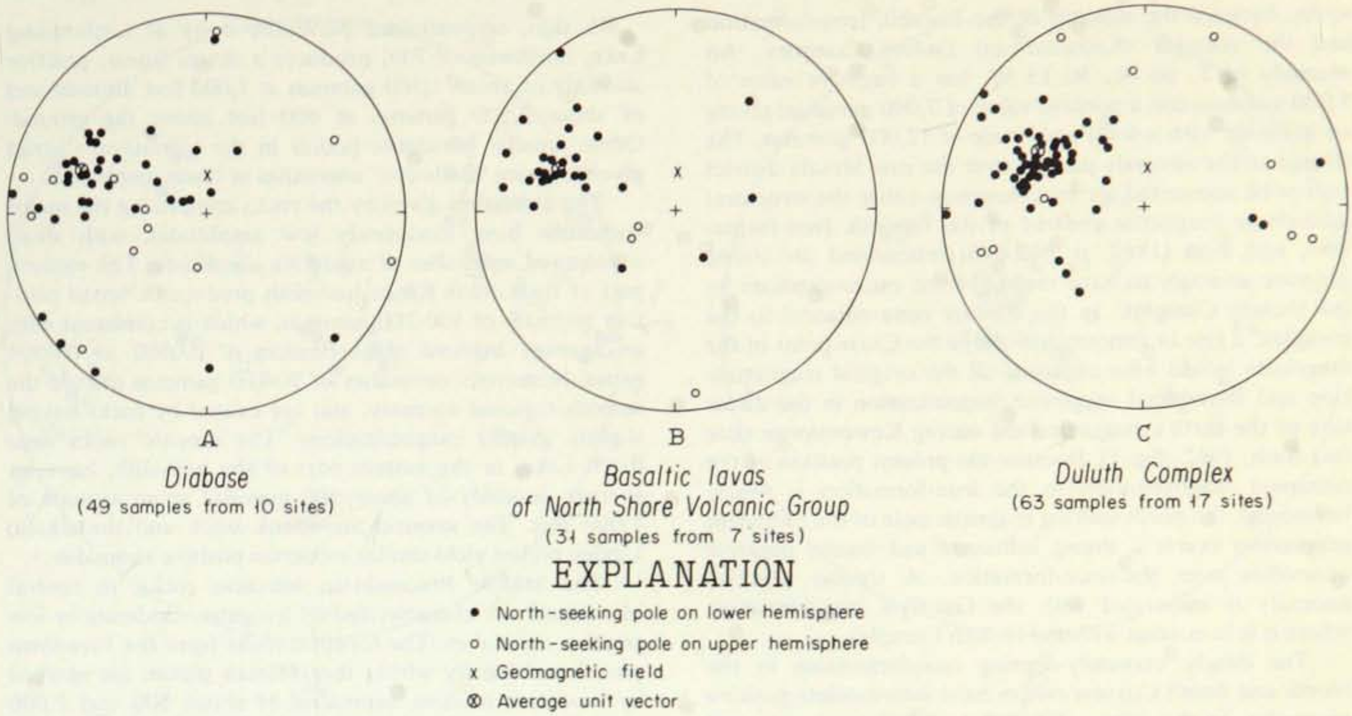


Figure VIII-4. Directions of remanent magnetization of Keweenawan rocks in vicinity of Duluth (after Bath and others, 1971).

underlies the Duluth Complex, and the so-called "Logan sills" of Lawson (1893) in the Rove Formation in Cook County (see Green, this volume) have reversed magnetic polarities. The diabase that gives the Esko anomaly, west of Duluth, probably also has a reversed magnetization.

All measured rocks other than the iron-formations and the Upper Precambrian mafic igneous rocks have a dominant induced magnetization, which is dependent on the amount of magnetite in the rock (see Mooney and Bleifuss, 1953; Jahren, 1963). Intermediate intrusive rocks—such as the dioritic facies of the Giants Range Granite in the Birch Lake area south of Ely, the Snowbank stock, part of the Hillman tonalite of Woyski (1949), and the Freedhem tonalite of Woyski (1949)—have moderate magnetizations, whereas most felsic intrusive rocks have weak magnetism. The metamorphosed basaltic lavas and associated metadiabases of the Ely Greenstone and of other Lower Precambrian units have low magnetizations because primary magnetite was destroyed by the pervasive low-grade metamorphism. An exception is the mafic volcanic rock included within the large granitic batholiths and in the contact zone of the Duluth Complex (Schwartz, 1924). These rocks contain magnetite and have moderate magnetizations. All samples of graywacke and slate from the Lower and Middle Precambrian units (Mooney and Bleifuss, 1953, table 2) and of the Hinckley Sandstone and the Fond du Lac Formation are nonmagnetic, as are the Phanerozoic rocks of the region.

MAGNETIC ANOMALIES

The factors determining magnetic anomalies in Minnesota have been discussed by Bath (1960, 1962) and Bath and others (1971). They have shown that the induced magnetism cannot explain all the anomalies, and that both induced and remanent magnetization must be considered.

The Middle Precambrian iron-formations give magnetic anomalies that vary in amplitude and character. Except where magnetite has been oxidized to hematite and limonite and where the magnetite content of unaltered formations is negligible, the iron-formations produce anomalies ranging from a few hundred to several thousand gammas in aeromagnetic traverses 1,000 feet above the ground. The Biwabik Iron-formation of the Mesabi range has been studied in detail (Bath, 1962). The western end of the range —west of Calumet—has anomalies ranging from 50 to 200 gammas, which are consistent with a very low magnetite content in the Biwabik. The main Mesabi district is overlain by irregular highs and lows that reflect changes in the magnetite content of the iron-formation, and which are interpreted (Bath, 1962, p. 635) to indicate that the dominant magnetization (about 0.012 gauss) is induced and across the layers (in the direction of the geomagnetic field). The total magnitude of these anomalies is about 1,000 gammas. The anomalies associated with the East Mesabi district are markedly different. East of Mesaba a strong negative anomaly overlies the iron-formation and irregular highs occur to the south, between the subcrop of the Biwabik Iron-formation and the younger (Keweenawan) Duluth Complex. An anomaly in T. 60 N., R. 13 W. has a negative value of 5,000 gammas and a positive value of 7,000 gammas, giving an anomaly with a total amplitude of 12,000 gammas. The change in the anomaly pattern over the east Mesabi district cannot be accounted for by changes in either the structural attitude or magnetite content of the Biwabik Iron-formation, and Bath (1962, p. 642-646) interpreted the strong negative anomaly to have resulted from metamorphism by the Duluth Complex. In the contact zone adjacent to the complex, a rise in temperature above the Curie point of the magnetite would have removed all the original magnetization and introduced remanent magnetization in the direction of the earth's magnetic field during Keweenawan time (see Bath, 1962, fig. 1). Because the present position of the remanent magnetization in the iron-formation is nearly horizontal, the north-seeking magnetic pole of the remanent magnetism exerts a strong influence and causes negative anomalies over the iron-formation. A similar negative anomaly is associated with the Gunflint Iron-formation where it is in contact with the Duluth Complex.

The nearly vertically-dipping iron-formations in the North and South Cuyuna ranges have intermediate positive anomalies in the range of 2,000 to 3,000 gammas; most likely the dominant magnetization is induced.

The Lower Precambrian iron-formations have strong positive anomalies, even where folded, except in areas where magnetite has been oxidized, as at the Soudan mine and at Ely. The anomalies typically have amplitudes of about 10,000 gammas and are elongate parallel to the strike. The largest recorded positive anomaly in Minnesota (20,000 gammas at 1,000 feet above the ground) occurs over a vertical, folded iron-formation near Bear Lake, north of Nashwauk (Bath and others, 1965). The anomalies can be accounted for by a dominant remanent magnetization along the bedding, which is nearly parallel to the induced magnetism (fig. VIII-5). Bath and others (1971) believed that most of the remanence is viscous magnetization acquired from the action of the earth's present magnetic field.

The mafic igneous rocks of Late Precambrian age give irregular local anomalies of low to high amplitudes, many of which are negative. Because of the juxtaposition of positive and negative anomalies, the average (or regional) anomaly over the rocks depends markedly on the volume of rock affecting the magnetic field measurements. Because the remanent magnetization direction of the Upper Precambrian mafic rocks differs markedly from the direction of the present geomagnetic field, the anomaly patterns of these rocks differ from the induction anomaly. Bath (1960) has shown that the Keweenawan lava flows on the northwest limb of the Lake Superior syncline can be expected to have lower magnetic anomalies than those on the southeast limb because of southeastward tilting of the northwest limb since the remanent magnetism was imposed. Tilting of the northwest limb has rotated the direction of remanent magnetization to a nearly horizontal position, and as a consequence the north-seeking pole of the remanent magnetism exerts a strong influence and can produce negative anomalies on this limb.

A thin, serpentinized peridotite body at Little Long Lake, northwest of Ely, produces a sharp, linear, positive anomaly of about 1,500 gammas at 1,000 feet altitude and of about 2,500 gammas at 400 feet above the ground. Other, smaller peridotite bodies in the Vermilion district give moderate "bulls-eye" anomalies of lesser amplitude.

The anomalies given by the rocks comprising the major batholiths have moderately low amplitudes, with small superposed anomalies of moderate amplitude. The western part of the Giants Range batholith produces a broad positive anomaly of 100-200 gammas, which is consistent with an average induced magnetization of 0.0002 to 0.0004 gauss. Numerous anomalies of 50-100 gammas disrupt the smooth regional anomaly, and are caused by rocks having slightly greater magnetizations. The dioritic rocks near Birch Lake, in the eastern part of the batholith, have an average anomaly of about 500 gammas at an altitude of 1,000 feet. The syenitic Snowbank stock and the alkalic Linden pluton yield similar moderate positive anomalies.

The Middle Precambrian intrusive rocks, in central Minnesota, are characterized by irregular, moderate or low positive anomalies. The tonalitic rocks from the Freedhem pluton and locally within the Hillman pluton are marked by moderate positive anomalies of about 500 and 1,000 gammas respectively, which are consistent with the measured induced magnetizations (table VIII-2). The more felsic Warman quartz monzonite and Stearns magma series of Woyski (1949) give low-amplitude anomalies comparable to the McGrath Gneiss.

REGIONAL MAGNETIC PATTERNS

Because the Phanerozoic rocks are thin and essentially non-magnetic, the near-surface fabric of the Precambrian rock units is strikingly apparent in the magnetic patterns (*cf.* plate 1 and the aeromagnetic map by Zietz and Kirby, 1970).

The Lower Precambrian rocks of northern and western Minnesota are characterized by irregular, mainly northeast-trending magnetic anomalies that accurately reflect the fabric of the greenstone-granite complexes. The most prominent anomalies overlie the thin, steeply-dipping, banded iron-formations that are common in the metabasaltic units of the greenstone belts. Many of these are curvilinear, and indicate the complex fold patterns in these rocks. A few anomalies of nearly comparable amplitude result from small bodies of serpentinized peridotite. The batholith plutonic rocks, as for example the Giants Range batholith, produce low-amplitude, crudely linear anomalies or irregular, moderate-sized anomalies that are slightly higher than those over the felsic volcanic rocks and the graywacke-slates. The major fault in the Precambrian terrane, the Vermilion (see Sims, this volume), is expressed in northwestern Minnesota as a linear feature that separates terranes having contrasting magnetic trends and patterns.

The magnetic anomalies over the Lower Precambrian rocks of southwestern Minnesota reflect the easterly trend of these rocks, but are difficult to interpret because of the sparse exposures. The strong, relatively small linear and elliptical anomalies are interpreted as iron-formations and mafic igneous rocks. The differences in fabric and intensity

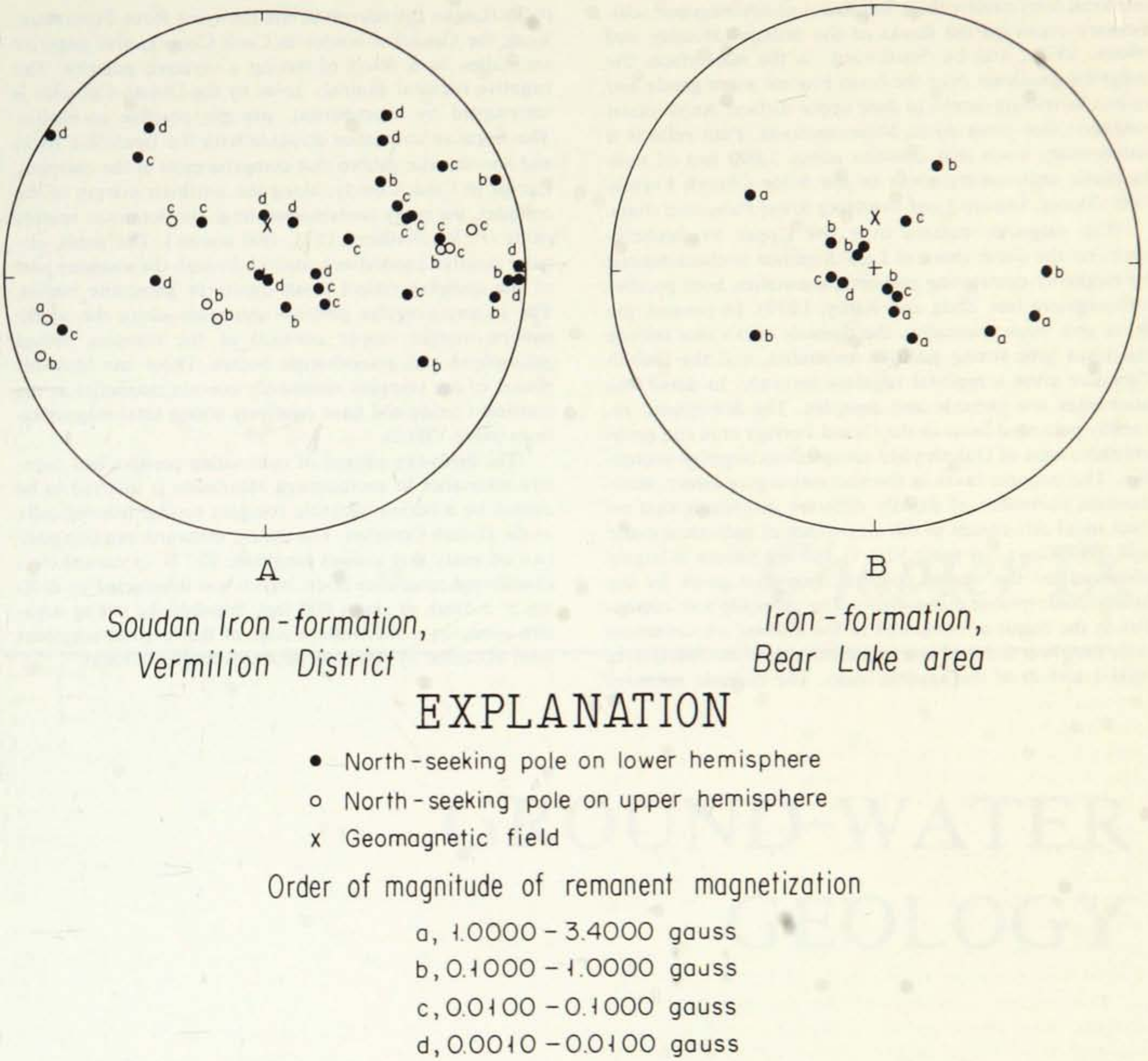


Figure VIII-5. Directions of remanent magnetization of Lower Precambrian iron-formations (after Bath and others, 1971).

of the anomalies may indicate a generally higher grade of metamorphism than in northern Minnesota.

The magnetic patterns over the Middle Precambrian rocks are more diverse and generally less distinctive than those of the Lower Precambrian rocks. Strong, linear magnetic anomalies coincide with the Mesabi range (Bath, 1962) and the iron-formations of the Cuyuna district (Schmidt, 1963). The lows between segments of strong positive anomalies over the Mesabi range reflect the important hematite-limonite direct-shipping ores of the range. In the vicinity of St. Cloud (Zietz and Kirby, 1970), numerous, scattered, small anomalies that are conspicuous but somewhat weaker than those produced by iron-formations mainly reflect intermediate and mafic intrusive rocks of Middle Precambrian age. The graywacke-slate and related strata of the various sedimentary rock units of Middle Precambrian age give a neutral but heterogeneous magnetic background.

The dominant magnetic feature of the state, given by Upper Precambrian rocks, coincides precisely with the Midcontinent Gravity High (Craddock and others, 1963; Sims and Zietz, 1967; King and Zietz, 1971). In east-central Minnesota and adjacent areas in Wisconsin, south of Lake Superior, a broad positive anomaly with short-period highs and lows overlies mafic lavas that are at or near the surface and local lows overlie thick sequences of nonmagnetic sedimentary rocks on the flanks of the feature (Mooney and others, 1970a and b). Southward, in the subsurface, the magnetic gradients over the lavas become more gentle and reflect increasing depths to their upper surface. An elliptical magnetic low centered at Minneapolis-St. Paul reflects a sedimentary basin that contains about 5,000 feet of nonmagnetic sedimentary rocks of the Solor Church Formation (Morey, in prep.) and overlying lower Paleozoic strata.

The magnetic pattern over the Upper Precambrian rocks on the north shore of Lake Superior is characterized by markedly contrasting magnetic anomalies, both positive and negative (see Zietz and Kirby, 1970). In general, the lavas give weak anomalies, the diabasic rocks that intrude the lavas give strong positive anomalies, and the Duluth Complex gives a regional negative anomaly. In detail, the anomalies are variable and complex. The lowermost, reversely polarized lavas in the Grand Portage area and probably also west of Duluth yield conspicuous negative anomalies. The younger lavas in the succession give linear, intermediate anomalies of slightly different magnitude that reflect small differences in the magnetism of individual mafic and felsic flows (see table VIII-1), but the pattern is largely obscured by the strong positive anomalies given by the nearly conformable diabase sills. The generally low anomalies in the basalt as compared to the diabase are consistent with the lower induced and remanent magnetizations (tables VIII-1 and 2) of the basaltic rocks. The diabasic intrusive rocks (Logan Intrusions) in the flat-lying Rove Formation, along the Canadian border in Cook County, give negative anomalies, as a result of having a reversed polarity. The negative regional anomaly given by the Duluth Complex is interrupted by short-period, strongly positive anomalies. The negative anomalies coincide with the troctolitic rocks and anorthositic gabbro that comprise most of the complex. Except in Cook County, along the northern margin of the complex, the rocks contain ilmenite as the dominant opaque oxide (P. W. Weiblen, 1971, oral comm.). The small, circular positive anomalies scattered through the southern part of the complex reflect small dunite or peridotite bodies. The larger, irregular positive anomalies along the southeastern margin (upper contact) of the complex reflect granophyre and granodioritic bodies. These late intrusive phases of the complex commonly contain magnetite as the dominant oxide and have relatively strong total magnetizations (table VIII-2).

The birds-eye pattern of contrasting positive and negative anomalies in southeastern Minnesota is inferred to be caused by a buried gabbroic complex similar lithologically to the Duluth Complex. The strong, eastward-trending positive anomaly that crosses longitude 92° N. is caused by a titaniferous magnetite body, which was intersected by drilling at a depth of about 800 feet. Possibly the strong negative anomaly at the western edge of the inferred subjacent body is caused by reversely polarized magnetic rocks.

Chapter IX

GROUND-WATER GEOLOGY

GROUND-WATER RESOURCES IN MINNESOTA

Rudolph K. Hogberg

Except for local areas in the northeast and west, Minnesota has adequate resources of good quality ground water. Most rural areas and towns obtain their water supplies from wells. Minneapolis and St. Paul use large volumes of ground water for air conditioning. However, the Twin Cities metropolitan area and many other large communities in the state obtain nearly all their water supplies from streams and lakes.

In the state, reliable data on the position, structural configuration, and areal extent of the bedrock aquifers generally are available only for the major urban centers. However, even in the urban areas specific information on the physical and chemical environments of the geologic units generally is poorly known.

STRATIGRAPHIC AND STRUCTURAL CONTROLS

In Minnesota, ground water occurs in both the unconsolidated rocks, mainly glacial drift, and the bedrock, and is available locally from one to several aquifers depending upon particular hydrodynamic and geologic conditions.

Glacial Drift

As discussed previously (Wright, this volume), four ice lobes flowed across the land surface of Minnesota during the latest, or Wisconsin, glaciation. Each lobe included from one to four phases of advance and recession (Wright and Ruhe, 1965). The composite effect of the Pleistocene glaciation is the several types of constructional features that now form Minnesota's varied landscape; only the extreme northeastern part of the state has landforms resulting from glacial erosion (fig. IX-1). Approximately 60 percent of the state's surface, other than lakes, is immediately underlain by morainal deposits, about 25 percent by lacustrine deposits, and the remainder by outwash deposits. Postglacial erosion, during the past approximately 10,000 years, has somewhat modified the surface glacial features.

The glacial deposits are composed of rock materials that were available to the ice lobes at the land surface. Materials carried from a north and northeastern provenance (fig. IX-1) are generally red and sandy and are composed predominantly of sand-size particles; the pebble-size fraction consists mostly of crystalline rocks. In contrast, the materials that were carried from a northwestern provenance are yellow or gray and calcareous; they are mostly composed of sand-size particles but have considerable quantities of silt- and clay-size particles; in some areas a considerable part of the sand- and pebble-size fraction is shale.

Inasmuch as the glacial deposits are heterogeneous, a knowledge of their stratigraphic relationships is needed to determine the environments most suitable for the occurrence of ground water. In the Minneapolis quadrangle, for example, studies of the surficial deposits indicate that the size, shape, and sequence of the glacial deposits depended upon the environment that existed during the time the ice lobes occupied the region. In this area, the moraines are composed mostly of till, and form sheet-like deposits, each layer of which consists of a heterogeneous mixture of particles ranging in size from silt to boulders. Many of the till deposits are composed dominantly of sands that are in a matrix of silt- and clay-size particles. Outwash plains, the composite aprons of several alluvial fans that were deposited by meltwater streams, tend to consist of several outwash deposits, each of which is lens-shaped in section, fan-shaped in plan, and composed of stratified sands. Ice-contact deposits—kames, eskers and ice crevasse fillings—are limited to small parts of the area and are transitional in lithology between till and outwash deposits. The primary glacial deposits commonly were reworked by waters that formed small glacial lakes and by late- and postglacial streams. The resulting tabular bodies of stratified sand commonly contain silty sand layers. In some parts of Minnesota, large glacial lakes such as Agassiz, Aitkin, Duluth, Nemadji, and Upham (fig. IX-1) provided basins for the accumulation of silty and clayey sediments from 5 to 80 feet thick. These extinct glacial lake basins as well as most smaller basins in northwestern, north-central, and east-central Minnesota now have a surface cover of peat. Thin

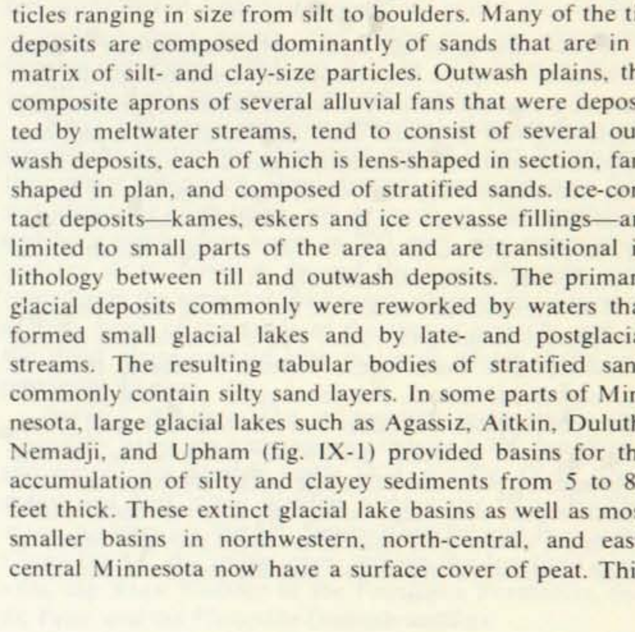


Figure IX-1. Surficial (glacial) geologic map of Minnesota.

layers of loess, composed mostly of silt-size particles, cover the bedrock and glacial drift in parts of southeastern Minnesota (Wright, this volume).

Little or no definitive quantitative information is available in the state on either recharge of surface waters to the ground-water system or addition of ground waters to the surface water system. Judged from the sparse data, it seems probable that tills form confining beds, lake deposits form impermeable to semi-permeable barriers to water flow, and ice-contact deposits provide environments similar to outwash and till deposits. Most of the data now available concern the state's shallow-drift aquifers from which most of the ground water is pumped.

Bedrock

Rocks ranging in age from Early Precambrian to Late Cretaceous comprise the bedrock of Minnesota (pl. 1). The bedrock surface commonly is buried beneath approximately 100 to 200 feet of glacial drift, and locally more, and it is exposed in parts of northeastern and southeastern Minnesota. Altitudes on the bedrock surface range from a low of about 500 feet above sea level beneath the Red River Valley, in the northwest, to 2,301 feet at Eagle Mountain in Cook County, in the northeast (fig. IX-2).

The Precambrian rocks are dominantly impervious igneous and metamorphic rocks; Upper Precambrian arkosic sandstones and shales occur beneath Paleozoic rocks in east-central Minnesota. Paleozoic rocks crop out in southeastern Minnesota and occur beneath Mesozoic and Cenozoic strata in extreme northwestern Minnesota. In the southeastern part, the Paleozoic rocks comprise the Hollandale embayment (see Austin, this volume). Small basins and faults within the Precambrian rocks provided minor to major controls on deposition and the present structure of the overlying Paleozoic succession (Austin, 1969). The Twin City basin, the only formally named basin in the state, is about 2,000 square miles in area and contains a maximum of about 1,100 feet of Paleozoic and 200 feet of Upper Precambrian rocks that comprise the water-bearing strata of the basin. A maximum of 1,500 feet of Paleozoic rocks (Austin, 1970b) occur within the southern part of the Hollandale embayment. In the northwestern corner of the state, as much as 450 feet of Ordovician rocks are overlain by about 40 feet of Jurassic "red beds" at the eastern margin of the Williston basin.

Upper Cretaceous sandstones and shales, which range from 50 to 400 feet in thickness, unconformably overlie the older bedrock in western Minnesota (Austin, this volume).

GROUND-WATER PROVINCES

To describe the ground-water resources of Minnesota, it is convenient to divide the state into four provinces—southeastern, east-central, northeastern, and western (fig. IX-3). The boundaries of each province are drawn as closely as possible to the limits of the controlling geologic features, which generally are various stratigraphic units within the glacial drift. The southeastern province is characterized by near-surface glacial deposits, mainly of northwestern provenance, which overlie Paleozoic bedrock within the Hollandale embayment. The east-central province is underlain by drift of northwestern provenance which averages about 100 feet in thickness. Tills comprise most of the surficial deposits, and extensive areas of outwash sands occur in the western part of this province. The principal bedrock aquifer, the Hinckley Sandstone, underlies the drift in the eastern part. The northeastern province has a thin discontinuous cover of drift of northeastern and northern provenances. The western province has the most varied

Figure IX-2. Topographic map of bedrock surface.

Figure IX-3. Ground-water provinces of Minnesota.

geologic environment of the four ground-water provinces. In most of this area, impervious crystalline rocks of Precambrian age are directly overlain by 50 to 400 feet of Cretaceous shales and sandstones. In the northwestern corner of the province, as much as 500 feet of Ordovician and Jurassic rocks overlie the basement rock surface. Glacial Lake Agassiz, which covers most of the northwestern part, contains clayey lake plain sediments and small local sand bodies. Drift, dominantly of northwestern provenance, and associated surface and buried outwash deposits, overlies the bedrock in the southern part of the province.

Southeastern Province

The state's largest yields of ground water are from the southeastern province. The combined Mt. Simon-Hinckley and the combined Prairie du Chien-Jordan are the most prolific aquifers. Glacial drift aquifers yield low (5-100 gpm) to moderate (100-695 gpm) amounts of ground water.

The ground-water supplies in the southeastern province are adequate for present and foreseeable needs. In only a few areas are sufficient quantities of ground water of acceptable quality absent. In much of the province, sand and gravel aquifers (fig. IX-4) are capable of yielding much more water than is currently withdrawn; some of them have potential yields of more than 500 gpm to wells less than 300 feet deep. The bedrock aquifers are among the highest yielding aquifers in the United States. Yields from the Prairie du Chien-Jordan and the Mt. Simon-Hinckley aquifers range from 500 to 3,000 gpm to wells 400 to 1,800 feet deep. The ground water from the southeastern province is suitable for most uses with moderate treatment. Generally, the hardness and high iron content need to be corrected. Other chemical characteristics such as sulfate, chloride, nitrate, and dissolved solids meet drinking water, irrigation, and many industrial standards.

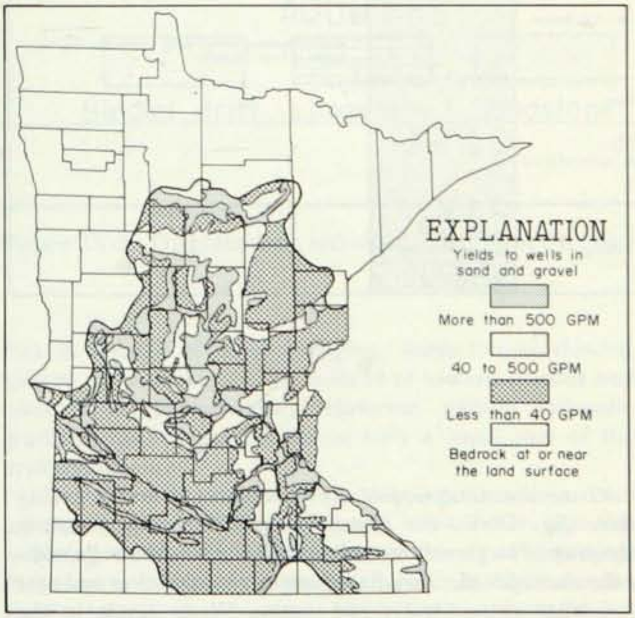


Figure IX-4. Yields to wells in sand and gravel deposits, upper Mississippi River watershed.

Twin Cities Metropolitan Area

Ground water is stored within the bedrock, glacial drift, and post glacial alluvium (table IX-1) in the Twin Cities metropolitan area.

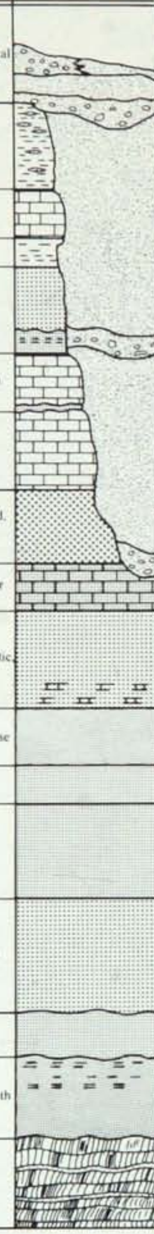
Glacial Drift and Alluvium Aquifers. A sequence of glacial deposits from 50 to 500 feet thick overlies the eroded Paleozoic bedrock surface in the area. Shallow—less than 125 feet deep—aquifers provide water for most suburban domestic wells, and some of the wells have been polluted by the effluents from soil absorption systems. Numerous lenses of gravelly sand are known to occur in the drift at medium depths, but information on their water quality and yields is poorly known. Probably the highest yielding drift aquifers are bodies of sand that are confined within the network of buried bedrock valleys. The valleys that have been delineated (see Payne, 1965) are 0.25 to 1.25 miles wide, 50 to 450 feet deep, and extend for tens of miles.

River terraces containing abundant deposits of sand parallel the Mississippi, Minnesota, and St. Croix Rivers and their major tributaries. These deposits are as much as 200 feet above the river levels, are as much as 150 feet thick, and have yields of as much as 500 gpm. The water table surfaces slope, however, toward the river channels, and accordingly the saturated thicknesses attain only a few tens of feet. Some of the sands of the terrace deposits and of the flood plains contain silty materials whose yields approximate 40 gpm.

Bedrock Aquifers. The sedimentary rocks within the Twin City basin are a maximum of about 1,300 feet thick, and contain six aquifers (table IX-1 and fig. IX-5). Two major and four minor aquifers yield water to wells. The major aquifers are the Mt. Simon-Hinckley and Prairie du Chien-Jordan; the minor ones include the Ironton-Galesville, the Reno Member of the Franconia Formation, the St. Peter, and the Platteville-Decorah aquifers.

The Platteville-Decorah, the shallowest of the Twin City aquifers, yields low quantities of water from fractures. The underlying St. Peter Sandstone, which is approximately 140 feet thick, is capable of yielding moderate quantities of moderately hard (61-120 ppm) water, but the quality of the water is questionable in heavily urbanized areas. Beneath the St. Peter is the combined Prairie du Chien-Jordan aquifer. Of the two units, the Jordan Sandstone has the greater porosity, but the Prairie du Chien Group provides the highest yields. The 5-foot-thick, very silty dolomite beds of the St. Lawrence Formation separate the Jordan Formation from the 130-foot-thick Reno Member of the Franconia Formation. Small quantities of moderately hard water are available from the Reno. Below the Reno, separated by silty to dolomitic sandstone beds of the Franconia confining bed, is the approximately 65-foot-thick Ironton-Galesville aquifer. Intakes for wells in the Ironton-Galesville aquifer range in depth from 500 to 1,200 feet. Because of the higher yields from aquifers above and below, few wells are completed in the Reno and the Ironton-Galesville, and accordingly only sparse data are available on their yields. However, water levels are lower than in the Jordan Sandstone, showing that recharge is received from above. The Mt. Simon-Hinckley aquifer, the deepest and second most important aquifer of the Twin City basin, yields moderate

Table IX-1. Water-bearing characteristics of geologic units, Twin City basin.

System	Rock Unit	Approx. thickness (in feet)	General Description	Graphic Column	Water - Bearing Characteristics
Quaternary	Undifferentiated glacial deposits	0-500	Glacial till, outwash, and valley train sand and gravel, lake deposits, and alluvium of several ages and several provenances; vertical and horizontal distribution of units is complex		Distribution of aquifers and confining beds is poorly known; sand and gravel aquifers that yield moderate to large amounts of water are common in buried bedrock valleys
Ordovician	Decorah Shale	90	Shale, greenish-gray, fissile to blocky; includes thin discontinuous lenses of fossiliferous limestone that increase in abundance upward		Aquifer: Low yields from fractures in shale and solution cavities in dolostone
	Platteville Formation	to 35	Dolostone, light-gray to buff, thin- to med.-bedded, shaly		
	Glenwood Formation	to 5	Shale, greenish-gray, fissile, sandy		Confining bed
	St. Peter Sandstone	150	Sandstone, light-gray, massively bedded, well sorted, med.-gr., poorly cemented, quartzose; approx. 20-ft.-thick silty to shaly bed near base		Aquifer: moderate yields Confining bed
	Prairie du Chien Group: Shakopee Formation	50	Dolostone, buff, thin- to thick-bedded, silt- and sand-rich, med.-gr. thin sandstone beds near base		Aquifer: high yields from fractures in dolostone and from poorly cemented sandstone; principal aquifer of the Twin City basin
	Prairie du Chien Group: Oneota Dolomite	100	Dolostone, buff, thin- to thick-bedded, vuggy, med.-gr., silt-size dolomite matrix		
Cambrian	Jordan Sandstone	90	Sandstone, light-gray, massively bedded, med.- to coarse-gr., well sorted, poorly cemented, quartzose		
	St. Lawrence Formation	50	Dolostone, gray to tan, silty or sandy, argillaceous; glauconitic in upper part		Confining bed
	Franconia Formation	155	Sandstone, greenish-gray, thin-bedded, fine- to coarse-gr., silty to dolomitic, commonly glauconitic; an upper aquifer (Reno) is a fine-gr. sandstone		Aquifer: low yields Confining bed
	Ironton Sandstone	30	Sandstone, light-gray, poorly to well sorted, med.-gr., silt-rich, quartzose		Aquifer: moderate to high yields
	Galesville Sandstone	35	Sandstone, light-gray, well sorted, fine- to med.-gr., quartzose		
	Eau Claire Formation	to 130	Sandstone, red, fine- to med.-gr., silty, glauconitic; interbedded with grayish-green to red, fissile shale		Confining bed
	Mt. Simon Sandstone	160	Sandstone, light-gray, fine- to coarse-gr., quartzose; thin shale beds in upper part		Aquifer: moderate to high yields; second most important aquifer of Twin City basin
Keweenawan	Hinckley Sandstone	75	Sandstone, tan, med.- to coarse-gr., arkosic		
	Fond du Lac Formation and older sedimentary rocks	to 4,000	Sandstone and siltstone, fine-gr., well cemented, arkosic; interbedded with red to green micaceous shale		Confining bed
	Metamorphic and Igneous Rocks	to 20,000	Mostly mafic, lava flows with thin interflow sediments		

to high quantities of relatively soft (<60 ppm) water. It is underlain by the sandstone and shales of the Fond du Lac Formation. Although less permeable than the Prairie du Chien-Jordan aquifer, the Mt. Simon-Hinckley aquifer is thicker—about 235 feet—and yields about as much water to wells. However, the long-term yields probably will be less because of its slower rate of recharge.

Ground-water Supplies. At most places in the Twin City basin (fig. IX-5), the water is under artesian pressure. Pumping of large volumes of ground water from wells within the metropolitan area has changed the direction and rate of normal ground-water circulation. Water levels in the bedrock and glacial aquifers have declined appreciably since the first records in 1885. Piezometric surface maps show

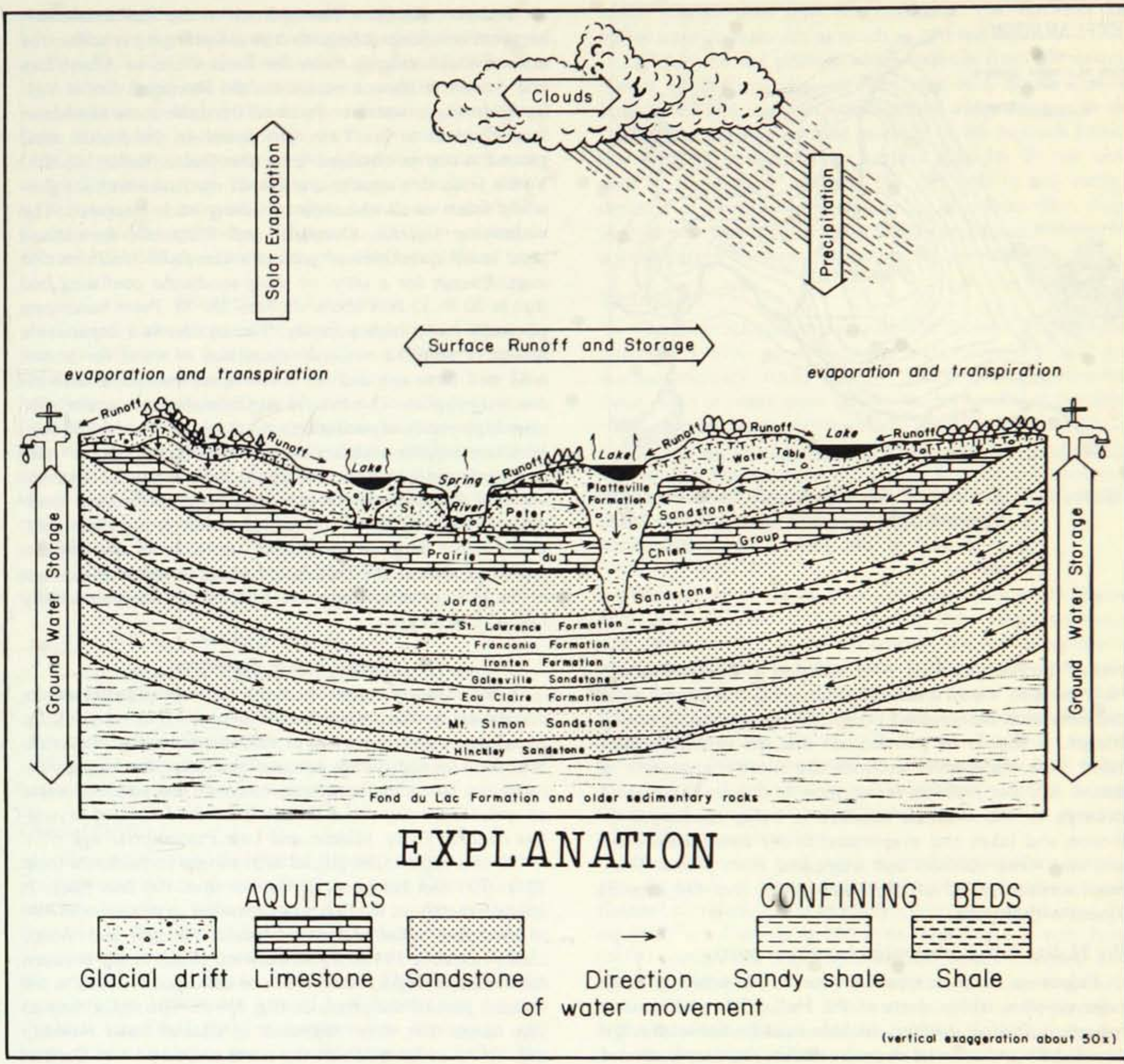


Figure IX-5. Diagrammatic section through Twin City basin showing probable direction of water movement.

that as a result of heavy pumping, water is now flowing toward the pumping centers instead of toward streams and lakes as it did previously. However, natural hydraulic gradients have been reversed in only a small part of the metropolitan area.

Bedrock wells in downtown Minneapolis and St. Paul are closely spaced and interfere with one another, especially during the heavy pumping required for air conditioning. Piezometric maps drawn on the Prairie du Chien-Jordan aquifer indicate that the water levels declined about 70 feet in downtown Minneapolis and as much as 90 feet in downtown St. Paul between 1885 and 1965 (fig. IX-6). Water levels within the Mt. Simon-Hinckley aquifer for the same period have been lowered as much as 230 feet in downtown Minneapolis and as much as 170 feet in downtown St. Paul. The greatest declines, in general, coincide with the areas of greatest pumpage. Outward from the well concentrations, the water-level declines have been less. Water levels in the Mt. Simon-Hinckley aquifer generally fluctuate from less than a foot to about 10 feet during the day. Water levels in the Prairie du Chien-Jordan aquifer, however, fluctuate as much as 5 to 10 feet in a day in winter and as much as 5 to 40 feet a day in summer. The water levels generally do not return, in the off-season, to the

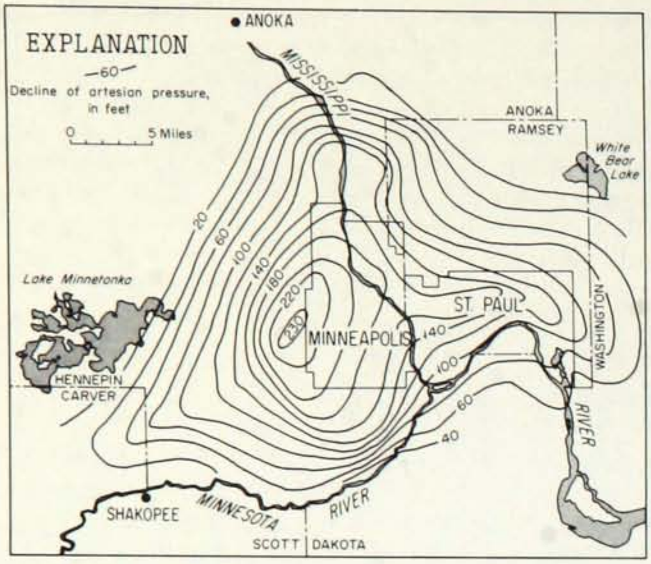


Figure IX-6. Decline of the piezometric surface of Mt. Simon-Hinckley aquifer from 1885 to 1965, in a part of the Twin Cities metropolitan area.

level of the previous year, thus resulting in a general downward trend in water levels. This lowering has caused local problems such as increased pumping costs and pump setting charges. It should be pointed out that the problems associated with water-level declines are mostly economic in nature. Another problem is that most of the possible annual recharge to the bedrock aquifers is being discharged to streams and lakes and evaporated to the atmosphere from land and water surfaces and transpired from plants. Thus, the theoretical yield of the Twin Cities area far exceeds present withdrawals.

The Hollandale Embayment (southern part)

Paleozoic bedrock aquifers provide abundant groundwater supplies within parts of the Hollandale embayment. Important shallow aquifers include sand bodies within the glacial drift and alluvial deposits within the deeply eroded river valleys.

Glacial Drift and Alluvium Aquifers. The bedrock in the Hollandale embayment is covered by 50 to 200 feet of glacial drift, in the same way as in the Twin City basin. An exception is the so-called "driftless area" in the extreme southeastern part of the state, which contains scattered drift but is mostly covered by a thin layer of loess. The drift contains irregularly spaced sand and gravel aquifers of different sizes (fig. IX-4), the most important of which are confined within buried bedrock valleys.

The valley of the Mississippi River is filled with as much as 200 feet of alluvium, which are flood plain deposits as much as 10 miles wide. Wells within the alluvium have yields greater than 1,000 gpm, thus providing adequate water supplies for most industrial and municipal users within the area of the Mississippi River flood plain.

Bedrock Aquifers. The bedrock of the Hollandale embayment comprises a large southward-plunging syncline, the axis of which extends from the Twin Cities to Albert Lea and Austin. With one exception, the Devonian Cedar Valley Formation, water is obtained from the same sandstone formations as in the Twin City basin. In the Austin area, ground water is obtained from the Cedar Valley aquifer. Yields from this aquifer are low to moderate and are generally from small channelways along rock fractures. The underlying Galena, Decorah, and Platteville formations yield small quantities of ground water from fractures and vugs. Except for a silty- or shaly-sandstone confining bed that is 20 to 35 feet above its base, the St. Peter Sandstone generally has a high porosity. The aquifer is a dependable source of small to moderate quantities of water for household and farm use and for a few small municipal and industrial supplies. The Prairie du Chien-Jordan aquifer supplies high yields of moderately hard water. The Mt. Simon-Hinckley aquifer supplies moderate to high yields of relatively soft water, but the Mt. Simon is absent at the northern limits of the southeastern province, and the Hinckley Sandstone is missing in all but the northern prong. Scattered, thin patches of Upper Cretaceous clay, shale, and impure sandstone overlie the Paleozoic surface in about 20 percent of the Hollandale embayment. Yields from these are mostly low—10 to 50 gpm.

East-Central Province

Ground-water resources of the east-central province occur mostly within outwash sand deposits that were left by the glaciers that moved over the area from the northeast. Moderate to high yields are available from the Upper Precambrian Hinckley Sandstone, and very low yields of water of variable quality are available from the fractured crystalline rocks of Early, Middle, and Late Precambrian age.

In this region, the glacial drift ranges in thickness from 50 to 300 feet but generally is less than 100 feet thick. It consists mostly of till that was deposited as moraines of low to moderate relief and as till plains (Wright and Watts, 1969; Leverett, 1932). Small outwash plains occur between the morainal ridges; extensive outwash deposits occur in the western part of the province (fig. IX-4). The red, clayey to silty sands that were deposited in Glacial Lake Nemadji (fig. IX-1) cover much of the northeastern part of Carlton County. These sediments have low ground-water yields. Similar sediments occur in the areas occupied by lakes Aitkin and Upham.

Wells tapping the outwash deposits yield as much as 500 gpm. Large supplies of ground water are available from outwash deposits within the St. Croix River watershed, in the areas east and south of Hinckley, and from the outwash deposits and the river terraces that parallel the Mississippi River from Little Falls downstream to Minneapolis. Probably, recharge of surface waters into these aquifers is rapid. The sandy sediments could provide high-volume, sustained yields of ground water for various uses in industrial plants and in agricultural irrigation.

Moderate to high quantities of water are obtained from thick sand lenses in the glacial drift in the Mesabi district. The direct-shipping iron ore bodies in the district also yield medium to high quantities of ground water. The water from both the iron ore bodies and the sandy drift is similar in quality; it is low in dissolved solids, moderately siliceous, hard to very hard, and high in iron and manganese.

Moderate to high yields of ground water are obtained from the highly fractured rocks of the Trommald Formation and associated Animikian (Middle Precambrian) rocks of the Cuyuna district. The Hinckley aquifer in Pine County and adjoining parts of Kanabec, Isanti, and Sherburne Counties yields moderate—200 gpm to 400 gpm—amounts of hard (121-180 ppm) water having a high content of iron.

Northeastern Province

Ground-water supplies for needs greater than domestic uses are difficult to find in many parts of the northeastern province because of a generally thin drift cover. The Duluth-Superior metropolitan area obtains most of its water from Lake Superior. Water supplies for agriculture, mining, tourism, and lumbering are mostly withdrawn from streams and lakes. The glacial drift and the postglacial alluvium and beach deposits in the province are potential sources of moderate amounts of water. Low yield—1 to 20 gpm—water supplies can be obtained from fracture zones in the bedrock. The fracture zones are difficult to locate, and the water found in them ranges widely in quality.

Western Province

The potential for finding adequate ground-water supplies within the western province varies widely, depending upon the local geology. In the northern part, ground water is available from sand bodies in the drift, and in the southern part it occurs both in sand bodies within the drift and in poorly cemented Cretaceous sandstones.

Glacial Drift

For convenience, the glacial deposits within the western province, which range in thickness from 50 to 500 feet, can be described with respect to three major landscape regions, (1) the Glacial Lake Agassiz plain, in the northwestern and northern parts, (2) the central uplands, and (3) the wet prairie in the southern part. The Glacial Lake Agassiz plain, which was formed by Glacial Lake Agassiz —which inundated parts of northwestern Minnesota and parts of the Canadian provinces of Manitoba, Saskatchewan, and Ontario in late-glacial time—is underlain mainly by clayey to silty sands as much as 150 feet thick. Gravelly sandbars were formed at the sites of former beaches. These are discontinuous and are linear in shape.

The central uplands is an area of medium to high relief that is underlain generally by 200 to 300 feet of glacial drift. An easterly-trending ridge called the Itasca moraine extends from the Lake Agassiz basin through parts of Becker and Hubbard Counties to Leech Lake; to the north, low relief moraines and till plains border the eastward extension of Lake Agassiz. A fishhook-shaped ridge, 25 to 40 miles wide, trends southeasterly from the eastern end of the Itasca moraine to the point where it crosses the Mississippi River valley between Elk River and Clearwater, and then trends northeastward to Grantsburg, Wisconsin. This ridge is called the Alexandria moraine. Outwash plains occupy a much smaller area than the moraines. The outwash and valley train deposits are as much as 250 feet thick.

Within the wet prairie, which extends from the eastern border of the province to the Iowa and South Dakota boundaries and includes the highland ridge known as the Coteau des Prairies, ground moraine or till deposits having low relief form most of the surface deposits. In the same way as the central uplands, the wet prairie has surface deposits of northwestern-provenance materials that range in thickness from 100 to 500 feet. Outwash plains comprise a small part of the near-surface glacial deposits.

Bedrock

The bedrock consists mostly of Lower Precambrian gneisses, schists, granitic rocks, metavolcanic rocks, and metasedimentary rocks (pl. 1). Sioux Quartzite overlies these rocks in many areas on the southwest side of the Minnesota River. In the northwestern part of the state, Ordovician and Jurassic sandstone, shale, and carbonate rocks overlie the basement rocks. Upper Cretaceous kaolinitic residuum, overlying shale, and, locally, carbonate rocks overlie the basement rocks in much of the region.

Northern Aquifers

Ground water can be obtained from many sandy glacial drift aquifers in the northern part of the western province. The yields to wells in the till and the lake plain sediments are generally low, and the water is high in iron and very hard. In parts of the Glacial Lake Agassiz region, where the ground water circulates through the Mesozoic and/or Paleozoic rocks and glacial drift, the water supplies are unsuitable for human consumption because of the high chloride content. Local excessive withdrawals in the Fargo-Moorhead area have resulted in decline in artesian pressure within the glacial aquifers.

The shallowest and most dependable aquifers are the gravelly sands of the abandoned beach bars and the river channel alluvium. Ground-water yields from these deposits generally are low—5 to 100 gpm. The water is very hard (>181 ppm) and high in dissolved solids. Surface and buried outwash deposits yield moderate amounts of ground water—100 to 200 gpm. The water is moderately hard, has high dissolved solids, moderate to high iron, low manganese, and varies in chloride content from 10 to 200 ppm.

Ground-water yields from the bedrock in the northern part of the western province are low in both quantity and quality. The Upper Cretaceous residuum and sandstone have low yields—5 to 50 gpm—are relatively hard, high in iron, and saline. The basement rocks are not suitable sources of water so far as known.

Southern Aquifers

In the southern part of the western province, the principal bedrock aquifers are the Upper Cretaceous residuum and poorly cemented sandstones. The basement rocks are unimportant sources of water. Important glacial aquifers include the buried outwash and the alluvium of the surface and buried stream channels.

Because of its large content of solids, the water from the Upper Cretaceous sandstones is of poor quality, but supplies of better water are not readily available. Intakes for wells tapping the Cretaceous aquifers are 150 to 300 feet deep; yields frequently exceed 150 gpm, although some wells yield only 40 gpm. An alternate source of water in parts of the areas underlain by Cretaceous aquifers is sand bodies within the glacial drift; these have low transmissibilities as well as a high content of solids.

The glacial drift and alluvium within the Minnesota River watershed are major sources of ground water. Commonly, water is obtained from depths of about 100 feet, but locally from depths of less than 20 feet. Buried outwash deposits yield as much as 1,000 gpm to wells as much as 400 feet deep. Alluvium within most of the major tributaries to the Minnesota River, including the Chippewa, the Pomme de Terre, and the Lac qui Parle rivers, yields medium amounts of water. Locally, alluvial deposits exceed 100 feet in thickness and yield up to 1,000 gpm. In the upper part of the Minnesota River Valley the flood-plain alluvial deposits are relatively thin and the basement rock surface is at or near the land surface. In the lower reaches of the river, however, the glacial and postglacial deposits may be major sources of ground water. To date, the yields from the outwash and alluvial aquifers of the Minnesota River watershed have been adequate for municipal and industrial needs.

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