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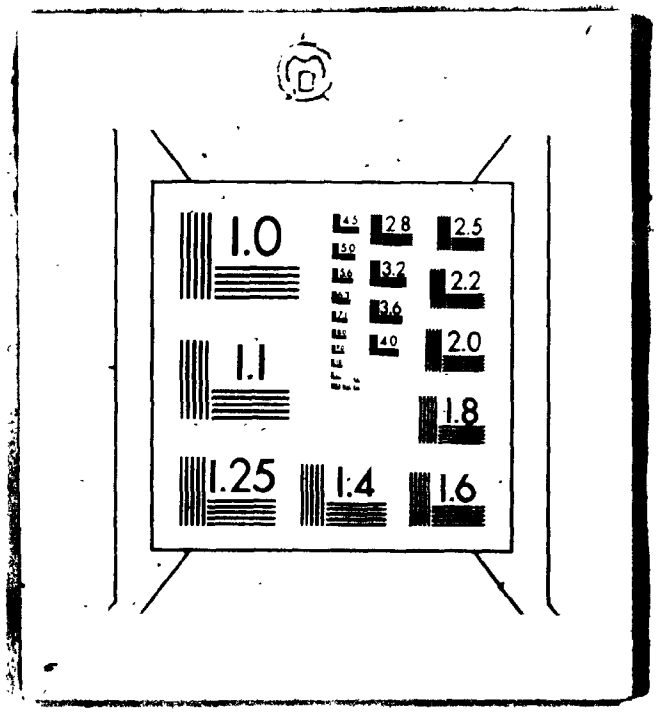

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STRUCTURE AND STRATIGRAPHY
OF THE
NORCAN LAKE AREA,
GRENVILLE PROVINCE, SOUTHEASTERN ONTARIO

by

Frank Adam Karboski, B.Sc.

A thesis submitted to the Faculty of
Graduate Studies in partial fulfilment
of the requirements for the degree of
Master of Science

Department of Geology
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Ottawa, Ontario

Feb., 1980

The undersigned hereby recommend to the Faculty of Graduate Studies acceptance of this thesis, submitted by Frank Adam Karboski, B.Sc., in partial fulfilment of the requirements for the degree of Master of Science.

Thesis Supervisor

Chairman, Department of Geology

FRONTISPIECE

F2 folds in composite amphibolite and tonalite orthogneiss (xenolithic horizon?) of unit 7. The fold pattern is analogous with the Mt. McCreary Antiform and Reddys Lake Synform mapped in the Norcan Lake area. Outcrop surface is nearly horizontal.

Near coaxial folds folding foliation parallel to bedding in marble and paragneiss of the Rossie area, N.W. Adirondacks of New York. The orientation and relative age of these folds are comparable with the F1 through F3 fold sequence recognized in the Norcan Lake area. Outcrop surface is nearly vertical.

ABSTRACT

Rocks of the Norcan Lake area, previously unmapped, lie between the Madawaska Highlands to the northwest and the Kaladar-Dalhousie Trough to the south. Much of the succession comprises granitic gneisses separated by calcitic marble, and includes para-amphibolite, and syenite and tonalite orthogneisses metamorphosed to middle and upper amphibolite facies.

Four deformational events are recognized. The earliest produced a regional axial planar mineral foliation, which was subsequently deformed by three major fold sets. Early recumbent folding and transposition produced a secondary gneissic layering in the granitic gneisses. Major coaxial refolds of the composite foliation, plunging gently to the northeast and southwest with upright to steeply overturned axial surface orientations, dominate the structural pattern.

Granitic gneiss belts, interpreted as metavolcanics, occupy the cores of early phase isoclinal antiforms. The structure and stratigraphy of the area can be correlated with that in the Kaladar-Dalhousie Trough.

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I. INTRODUCTION

I.1. Purpose of Study

1. To provide a geologic map for a structurally critical area of the Grenville Province previously unmapped, located between the Hastings, Haliburton and Madawaska Highlands to the northwest and the Kaladar-Dalhousie Trough to the south.

2. To establish the structural history of the area and propose a stratigraphic succession.

3. To compare the deformational history of the area with previous work in the Kaladar-Dalhousie Trough to the south.

4. To contribute towards the understanding of the origin of the granitic gneisses in the area.

I.2. Location, Access and Physiography

The geologic map (Fig. 3, in pocket) covers an area of about 165 square kilometres located 90 kilometres southwest of Ottawa, Ontario. The map covers the entire township of North Canonto and parts of Blithfield and Brougham townships in the counties of Renfrew and Frontenac, Ontario (see inset map, Fig. 3). The map-area is contained within the Clyde Forks map sheet (No. 31F/2) of the Canadian National Topographic 1:50,000 Series.

The area is accessible by boat and foot from Black Donald and Norcan lakes, and the Madawaska River. Secondary gravel-surfaced roads from highway 508 provide some access into the north-western part. Two powerlines in the area are paralleled by

unsurfaced vehicle trails which provide important access into the area northeast of Norcan Lake and southwest of the Madawaska River.

The maximum relief is 200 metres, with an average of 80-100 metres. The topography is controlled to a considerable extent by the lithology and structure of the underlying rock units. Carbonate-rich rocks weather to form valleys between ridges of more weather-resistant gneisses containing appreciable amounts of quartz and feldspar. Prominent cliffs commonly parallel major lithologic contacts in the area. The trend of the valley and ridge topography generally follows the regional northeast to north-trending foliation. Swamps and small lakes commonly occur within the valleys.

Most of the area is heavily forested, but outcrops are abundant. Only a few, relatively small areas are heavily covered by surficial deposits. Nearly continuous exposure along the north-west-trending shores of Norcan Lake provide an almost complete cross section of much of the map-area.

I.3. Regional Geologic Setting

The study area is located in the Central Metasedimentary Belt, a subdivision of the Grenville Province defined by Wynne-Edwards (1972) (Fig. 1). In a general way, the area is located between the Hastings, Haliburton and Madawaska Highlands to the northwest and the Kaladar-Dalhousie Trough to the south, subdivisions of the Grenville Province in southeastern Ontario described by Hewitt (1956) (Fig. 2). The Hastings, Haliburton and Madawaska Highlands are a broad zone of sub-circular layered granitoid gneiss bodies that are

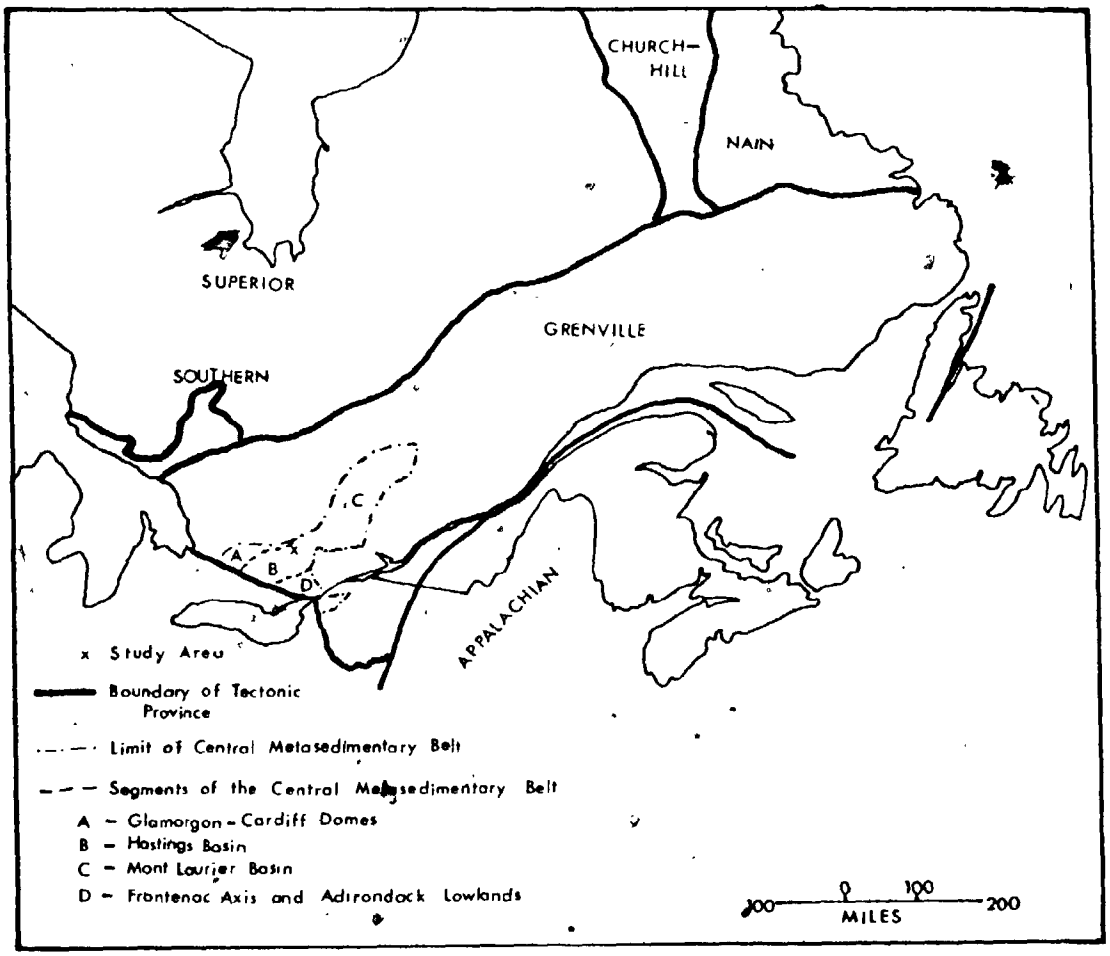

Figure 1. Location of the study area within the Central Metasedimentary Belt of the Grenville Province. Segments of the Central Metasedimentary Belt are shown. After Wynne-Edwards (1972).

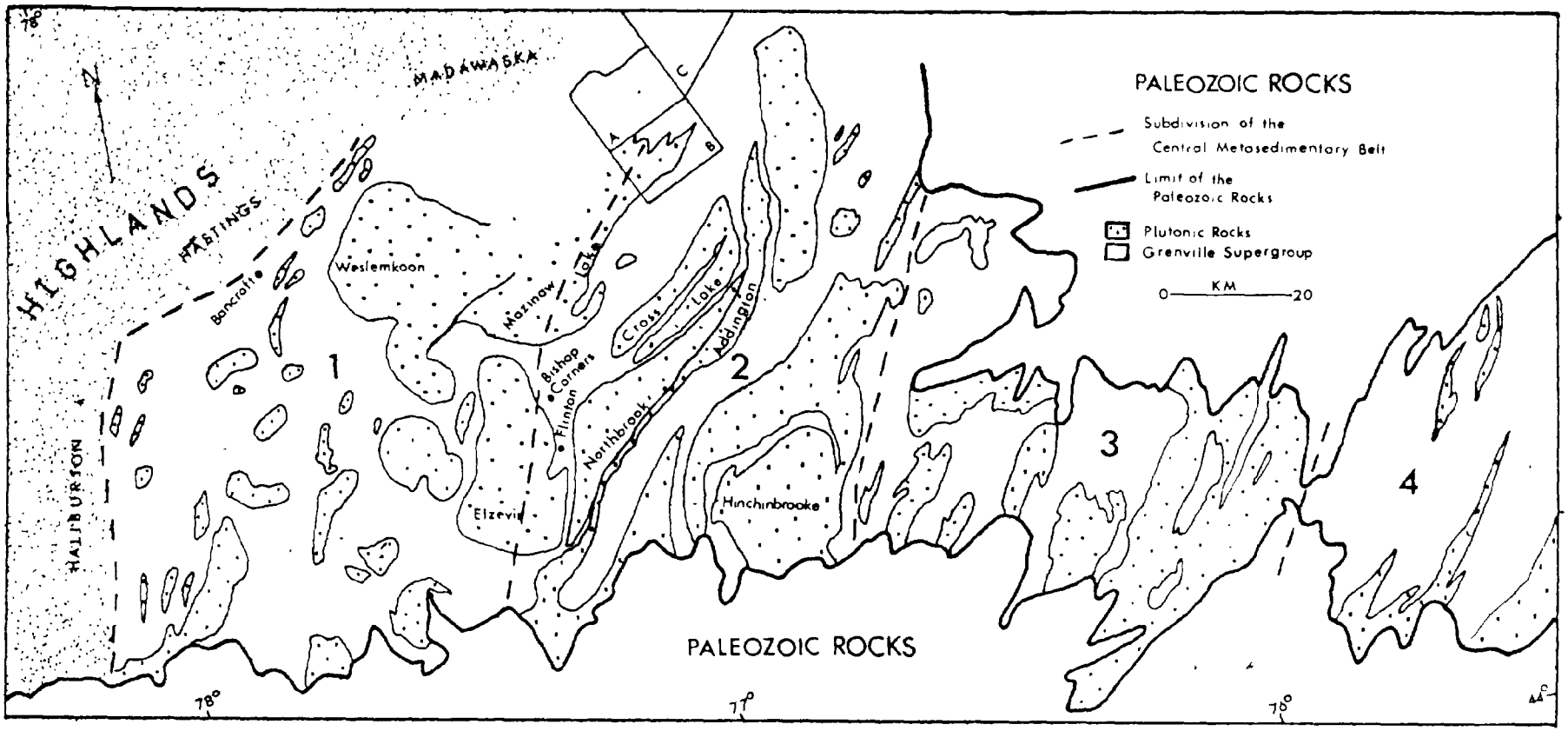

Figure 2. Subdivisions of part of the Central Metasedimentary Belt modified after Wynne-Edwards (1972) and Hewitt (1956): 1) Hastings Basin, 2) Kaladar-Dalhousie Trough, 3) Frontenac Axis and 4) N.W. Adirondacks. North Canonto (A), South Canonto (B) and Blithfield (C) townships are outlined. The study area is largely confined to North Canonto Township. The names of major intrusions discussed in the text are labeled.

separated by thin supracrustal belts of the Grenville Supergroup. This zone was later redefined by Wynne-Edwards (1972) as part of the Central Metasedimentary Belt in transition to the Ontario Gneiss Segment. Within the Kaladar-Dalhousie Trough, rocks of the Grenville Supergroup define long, northeast-trending supracrustal belts that are separated by metaplutonic bodies, largely of granitoid composition. The Elzevir and Weslemkoon batholiths and the Mazinaw Lake Pluton form the northwestern boundary, and the Hinchinbrooke Pluton forms the southeastern boundary. The Hastings Basin and Frontenac Axis flank the Kaladar-Dalhousie Trough to the west and east respectively. These are further subdivisions of the Central Metasedimentary Belt described by Wynne-Edwards (1972).

I.4. General Geology and Previous Work

Within the Hastings Basin and the Kaladar-Dalhousie Trough, rocks of the Grenville Supergroup form a thick succession of volcanic and carbonate-rich sedimentary rocks that have been intruded by numerous large plutons and subsequently deformed and metamorphosed by the Grenvillian Orogeny. Lumbers (1967a,b) has compiled stratigraphic, metamorphic and plutonic data for the Madoc-Bancroft area of the Hastings Basin. He suggested a stratigraphic succession for the Grenville Supergroup and proposed a metamorphic and plutonic history for the region. His proposals have found useful application by geologists working in various parts of southeastern Ontario.

Lumbers subdivided the Grenville Supergroup into the Mayo Group, consisting mainly of marble-rich metasedimentary rocks and the

Hermon Group, comprising predominantly metavolcanic rocks. In the Madoc-Bancroft area the grade of metamorphism is generally low (upper greenschist) and original stratigraphic relationships can be more fully interpreted. Lumbers suggested that pillowed mafic volcanics (Tudor formation) are the oldest exposed rocks in the area, which give way upwards and laterally to more felsic volcanics and contemporaneous metasediments of the Mayo Group. Lumbers considered the major intrusions of the area to have been emplaced during two distinct periods of magmatism; an early biotite diorite series, and a later granitic series. The Elzevir, Weslemkoon, Cross Lake, Hinchinbrooke and Northbrook batholiths or plutons (Fig. 2) are examples of the early biotite diorite series and are predominantly trondhjemitic to granodioritic in composition. The Addington and Lavant granites, sometimes referred to in the literature as gneisses, are considered to be examples of the later granitic series.

Radiometric dating in the area has supported the proposals of Lumbers. Lead isotope ages obtained by Silvers and Lumbers (1966) have been modified by Davidson et. al. (1979) using recent values for the uranium decay constants. The updated results are: 1236±15 Ma. for the oldest exposed volcanics (the Tudor formation of Lumbers, 1967a,b); 1226±25 Ma. for the Elzevir Batholith of the "biotite diorite series"; and 1104±25 Ma. for the Addington Pluton of the "granitic series". Regional metamorphism and intrusive activity continued to about 1050 Ma. (Krogh and Hurley, 1968).

Over the past several years, detailed studies in the Kaladar-Dalhousie Trough have in part supported, and added to, the stratigraphic

proposals of Lumbers (1967a,b). Immediately east of the Elzevir Batholith, the supracrustal belts have a strong north-northeast trend, contain an abundance of metavolcanic rocks, and are bounded by the Elzevir, Northbrook and Hinchinbrooke batholiths (Fig. 2), of Lumbers' "biotite diorite series". Near the village of Bishop Corners (Fig. 2), rocks of the Hermon Group form a succession of tholeiitic and calc-alkaline volcanics approximately 7 kilometres thick (Sethuraman, 1970; Sethuraman and Moore, 1973; Condie and Moore, 1977), which are overlain by marble-rich metasediments (Sethuraman and Moore, 1973). The Elzevir Batholith and probably the Northbrook Pluton intrude the metavolcanics in this area (Moore and Thompson, 1972). An erosional unconformity within the Grenville Supergroup has been documented (Moore, 1967; Sethuraman, 1970; Thompson, 1972; Moore and Thompson, 1972). Rocks above the unconformity are metasedimentary and have been named the Flinton Group; it occurs in elongate and narrow synclines that cross-cut the underlying sediments and Hermon Group volcanics. Deposition of the Flinton Group is considered to post-date all major igneous activity in the area.

The supracrustal belts exhibit a change in regional trend within the Kaladar-Dalhousie Trough. The north-northeast trend in the area east of the Elzevir Batholith changes to a northeast trend, beginning near the town of Bishop Corners and extending up to the vicinity of the study-area (Fig. 2). Marble-rich metasedimentary rocks dominate the exposed succession along this interval and amphibolitic rocks become less abundant. The metamorphic grade increases in this same northeasterly direction (see Fig. 14) and original

textures become increasingly difficult to recognize. For this reason, the origin of much of the amphibolite is uncertain, although a few metavolcanic rocks have been identified (Smith, 1958; Peach, 1958; Rivers, 1976). Compared with the supracrustal belts to the southwest in areas of lower grade, the stratigraphy of this area is not well understood.

The Norcan Lake area was unmapped prior to this study. Faculty and students of Carleton University field camps had mapped in the vicinity of Black Donald Lake (Fig. 3) and these data have been compiled and added to the geologic map in Figure 3. A large area of the northeasterly-trending supracrustal belts was mapped by Smith (1958) and Peach (1958) at a scale of 1:63360 and included seven townships in the counties of Frontenac and Lanark. The mapping of Smith (1958) included the townships of Lavant and South Canonto and this study extended his mapping into the previously unmapped townships of North Canonto and Blithfield. Rivers (1976) studied in more detail the supracrustal rocks mapped by Smith (1958). Geological Survey of Canada Map 1046A, by Quinn (1956), borders to the north of the geologic map in Figure 3. Aeromagnetic data is available for much of the map-area (Norgaard, 1970).

A large part of the Norcan Lake area is underlain by a heterogeneous complex of gneisses, largely of granitic composition. Calcitic marble forms a continuous belt mapped throughout the area and separates areas underlain by these granitic gneisses. A tonalite orthogneiss has been recognized which may correlate with Lumbers' (1967b) biotite diorite series of plutons. The origin of the granitic

gneisses however, has not been established with certainty and these are referred to in this study as heterogeneous gneiss (Fig. 3).

Heterogeneous gneiss has in part, been traced into South Canonto and Lavant townships (Fig. 2) where Smith (1958) correlated similar rocks with the Lavant gneiss. The Lavant gneiss is an elongate mass of granite which, except for a small isolated mass located in Lavant Township, is continuous for 40 kilometres to the southeast where it is referred to as the Addington Pluton (Fig. 2). Smith (1958) however, discusses problems that arise when ascribing a plutonic origin for all the rocks mapped as Lavant gneiss in this area. He notes that in various places the Lavant gneiss grades into paragneiss, appears intergradational with enclosing schistose rocks, and contains abundant units of schist and marble. He writes, "There is some question as to how much of the Lavant gneiss should be termed a 'granite' gneiss and how much should be given a noncommittal name, such as microcline gneiss." (Smith, 1958; pp. 21-22).

The contact between these granitic gneisses and the supra-crustal rocks to the southeast is shown as defining a northwestern boundary to the Kaladar-Dalhousie Trough in Figure 2 and is continuous to the southwest into the area immediately east of the Elzevir Batholith, where the Mazinaw Lake Pluton is recognized. There is little doubt of the plutonic origin of the granite in this area; Ayer (1979) mapped its contact with the metavolcanic rocks to the south and found ample evidence to establish an intrusive origin. One aspect of this study emphasizes the careful consideration that must be given when assigning an origin to granitic rocks in complexly deformed metamorphic terrains.

I.5. Method of Study

The geologic map in Figure 3 includes three map-areas as delineated in the inset map. The Norcan Lake-Madawaska River area was mapped in detail during the summer of 1978; the Closs Lake Mountain area was mapped on a reconnaissance scale during the spring of 1979. The work of Carleton University students participating in an advanced field course in the Black Donald Lake area was compiled by the writer and integrated with the rest of the map. Hereafter, the text pertains only to those areas mapped by the writer.

Mapping was conducted at a scale of four inches to the mile (1:15840). As the area is heavily forested, it was necessary to run pace and compass traverses across strike of the units. Traverses are typically spaced at 0.5 to 1 kilometre intervals in the area mapped in detail, and at 1.5 to 2 kilometre intervals in the area mapped on a reconnaissance scale. The outcrop distribution, lithologic contacts and structural measurements were recorded on air photographs during traverses and compiled on a base map of the same scale (1:15840). Petrographic study was concentrated on the heterogeneous gneiss, but thin sections of all map units were examined.

I.6. Acknowledgements

Drs. J.M. Moore and R.L. Brown provided financial assistance, advice, encouragement, and reviews of the manuscript. The writer is indebted to Dr. Moore for providing the opportunity to do this work.

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Valerie Jackson assisted in field mapping during 1978 and 1979. Her enthusiasm and companionship is appreciated.

I would also like to thank my father, whose advice long ago set me on course for this research.

II. LITHOLOGY

II.1. Introduction

The distribution of lithologic map units in the area is shown in Figure 3. The area has been complexly deformed and subjected to upper amphibolite facies metamorphism; consequently much of the stratigraphy of the area remains poorly understood. Metamorphosed rocks of the Grenville Supergroup (units 3 and 4) and metamorphosed plutonic rocks (units 7 and 8) have been recognized. A large part of the map-area is underlain by rocks of uncertain origin, forming a heterogeneous unit (unit 1) that is composed, to a large extent, of intermixed granitic and quartz dioritic gneisses. Units 5 and 6, which were mapped by Carleton University students beyond the writer's study area, have been omitted from the description.

II.2. Metamorphosed Rocks of Uncertain Origin

II.2.1. Unit 1 is a heterogeneous complex of fine to medium-grained gneisses composed predominantly of two distinct lithologies: pink, biotite-hornblende granitic gneiss and grey, biotite-hornblende quartz dioritic gneiss. These two types are coarsely interlayered in thicknesses which range from less than a metre (Plate 1a,b) up to several tens of metres. During field work an attempt was made to map these two types as separate map units. It was found however, that their distribution, both across and along the strike direction, was too variable to permit contacts to be drawn at the map scale.

Unit 1 is the dominant rock type of the map-area. Its areal

Plate 1

Stratigraphic Relationships Within the
Heterogeneous Gneiss Belts

a) Interlayered granite gneiss (A) and quartz diorite gneiss (B). Axe handle is 65 centimetres long. Green Lake Gneiss Belt, east shore of Norcan Lake.

b) Contacts between feldspathic quartzite (A), granite gneiss (B) and quartz diorite gneiss (C). Green Lake Gneiss Belt, east shore of Norcan Lake.

c) Typical outcrop of uniform, foliated granite gneiss. Axe is parallel to the foliation. Mt. McCreary Gneiss Belt, east shore of Norcan Lake, 1 kilometre northwest of McCreary Lake.

d) Typical foliated quartz diorite gneiss. Note mafic layers in center of photograph and compare with Plate 1e. Green Lake Gneiss Belt, east shore of Norcan Lake.

e) Well layered garnetiferous paragneiss. Outcrop is of a subunit interlayered within the quartz diorite gneiss in Plate 1d. Green Lake Gneiss Belt, east shore of Norcan Lake.

f) Gradational zone between granite gneiss (unit 1) and amphibolitic gneiss (unit 2), referred to as composite gneiss in the text. Amphibolite layers (A) increase in number towards the amphibolitic gneiss (below photograph). Note less distinct mafic layers (B) in granite gneiss and compare with the more typical granite gneiss in Plate 1c. Green Lake Gneiss Belt, northeast shore of Blithfield Long Lake.

Plate 1

a

b

c

d

e

f

distribution can be described in terms of three gneiss belts (Fig. 7). The Green Lake Gneiss Belt is in contact with unit 3, defining a synformal structure in the western part of the map-area; where in the east it is in contact with units 3 and 4, also in a synform. The Mt. McCreary Gneiss Belt is a large antiformal structure in the vicinity of Mt. McCreary. A third gneiss belt, the Closs Lake Belt, is defined in the area mapped on a reconnaissance scale and occurs between units 3 and 7, outlining a large, doubly-plunging antiform.

During field work, colour and total percent mafics provided adequate criteria for distinguishing the granitic and quartz dioritic gneisses. Selected samples from the three gneiss belts were thin-sectioned and mineral modes determined by point counting. The distribution of samples used is shown in the Appendix.

Thin sections were stained for plagioclase and K-feldspar and 1000 points from each slide were counted. A slab of each sample was stained for K-feldspar and provided a means of checking the results. Table 1 contains the results, tabulated in the form of average mineral percentage and range. The data were also plotted on quartz-K-feldspar-plagioclase and quartz-mafics-K-feldspar triangles and are shown in Figure 8.

The data show a bimodal distribution and support field observations that unit 1 consists of two distinct rock types. The classification of Streckeisen (1967) was used to classify the average mode of each group as granite and quartz diorite gneiss.

The granite gneiss has an average grain size of 1 millimetre. The rock is pink on both fresh and weathered surfaces and has an

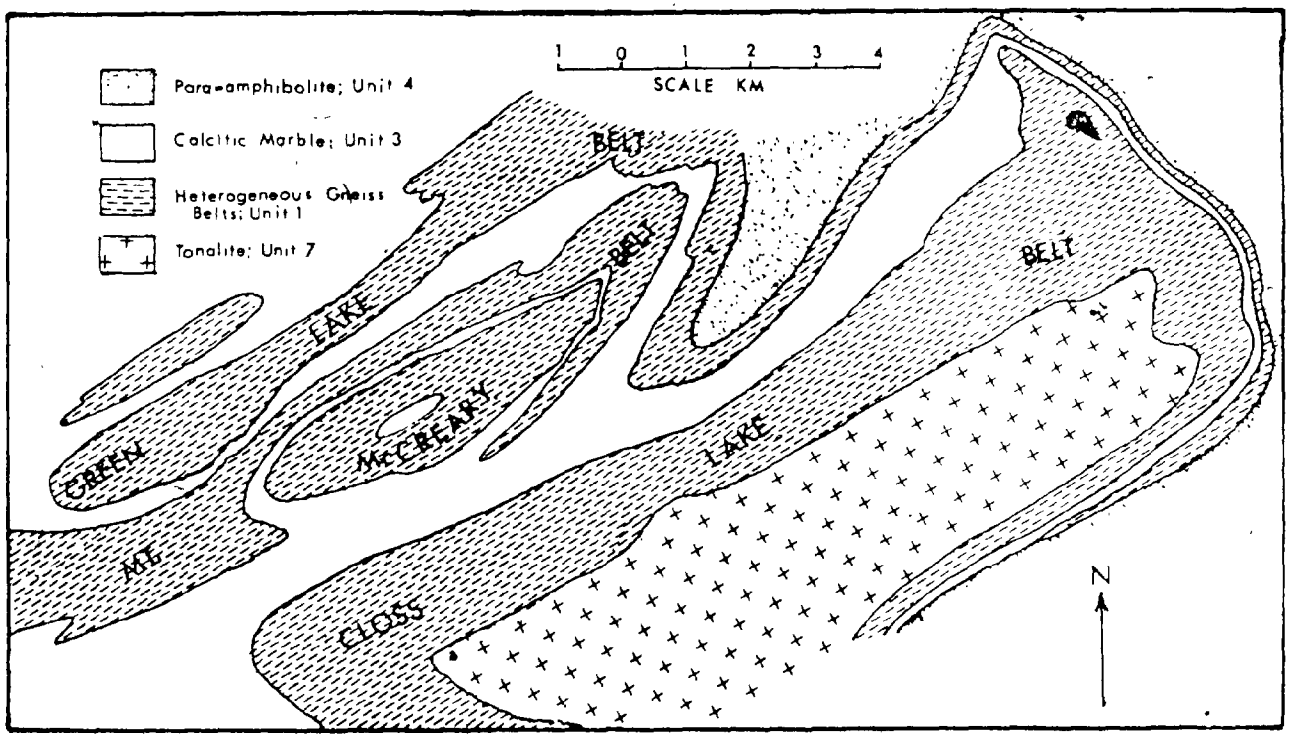


Figure 7. Simplified geologic map of the Norcan Lake area showing the distribution of the heterogeneous gneiss belts discussed in the text. Unit 2 could not be shown at the scale of the figure.

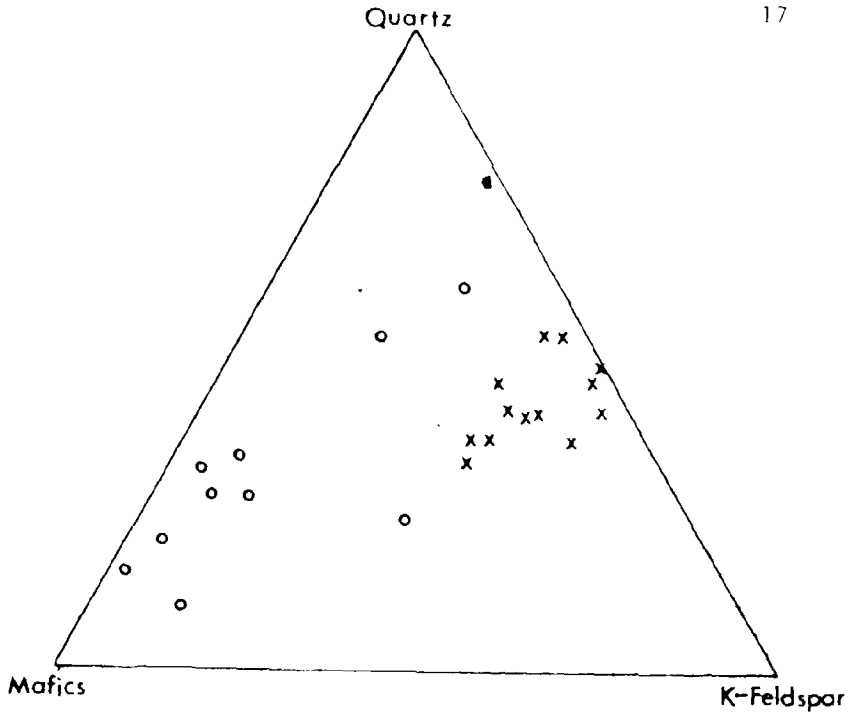
Table 1. Modal Analysis of Unit 1.

QUARTZ DIORITE GNEISS-10 samples

	AVERAGE	RANGE
QUARTZ	14.0	4.8-23.3
PLAGIOCLASE	52.0	47.6-55.9
K-FELDSPAR	4.4	2.0-7.8
BIOTITE	11.3	1.7-26.2
HORNBLENDE	17.0	3.8-34.4
OPAQUES	1.2	0.0-4.0
TOTAL	99.9	

GRANITE GNEISS-13 samples

	AVERAGE	RANGE
QUARTZ	27.5	18.5-40.7
PLAGIOCLASE	35.3	13.5-59.3
K-FELDSPAR	27.2	15.1-40.7
BIOTITE	3.5	2.2-10.0
HORNBLENDE	4.0	0.0-15.5
OPAQUES	1.6	0.0-1.8
TOTAL	99.1	



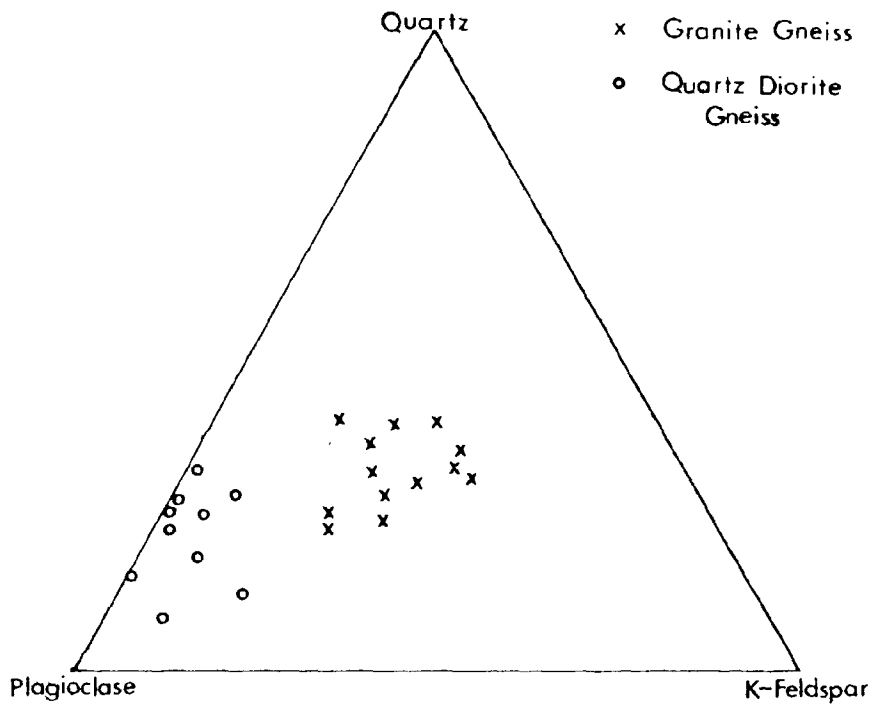


Figure 8. Results of modal analysis of unit 1 (Table 2) plotted on compositional triangles. Mafics apex includes biotite, hornblende and opaques.

average mafic content of 10%. In places, hornblende-rich horizons a few metres thick are encountered and the mafic content reaches 25-30%. Accessory minerals include apatite, sphene and magnetite. Garnet and sillimanite are local accessory minerals. The rock has a compositional layering expressed by the segregation of mafic minerals into thin and diffuse layers less than 1 centimetre in thickness, alternating with leucocratic, quartzo-feldspathic layers. The rock generally contains a planar fabric defined by the preferred orientation of biotite, and commonly a quartz aggregate mineral lineation. Though locally the granite gneiss contains a compositional layering that is probably primary (III.3.1.; Plates 1f; 3a,b), the unit is most typically a uniform foliated gneiss (Plate 1c).

Quartz diorite gneiss weathers grey and is light grey on fresh surfaces. The average grain size of the rock is usually slightly coarser than the granite gneiss. The unit is most commonly a foliated gneiss (Plate 1d) with an average mafic content of about 25%. As in the granite gneiss, the layering is expressed by a diffuse segregation of mafic minerals into thin leucocratic and melanocratic layers. A biotite planar fabric is weakly developed. Elongate prisms of hornblende, occurring in the plane of the foliation, locally define a mineral lineation. Accessory minerals include apatite, magnetite, sphene, garnet and clinopyroxene.

Quartz diorite gneiss is a more heterogeneous rock than the granite gneiss. Figure 8 indicates that the granite gneiss has a more uniform mode than the quartz diorite gneiss, which shows considerable variation in the mafic content. Locally, the quartz

diorite gneiss is gradational, generally through a transition zone a few metres thick, into a garnetiferous, compositionally layered, K-feldspar-hornblende-biotite-quartz-plagioclase gneiss of probable sedimentary origin. The transitional zone is characterized by an increase in the amount of biotite, the appearance of an abundance of garnet and a tendency for biotite and hornblende to form a compositional layering, which in the garnetiferous paragneiss is well expressed (Plate 1e). Rocks similar to the transitional type are also locally present in the quartz diorite gneiss without garnetiferous paragneiss, although here garnet is not as abundant. In these localities it is difficult to define a contact between the transitional type and the more typical quartz diorite gneiss. As in the quartz diorite gneiss, K-feldspar is a minor constituent of these transitional types.

Numerous thin horizons of metasediments were noted in the granite gneiss of the Green Lake Gneiss Belt (Fig. 7). Calcitic marble is the dominant metasediment but feldspathic quartzite, scapolite-calcite para-amphibolite and hornblende-quartz-biotite-plagioclase schist also occur. These types of metasediments were not noted in the quartz diorite gneiss anywhere in the map-area. Apparent thickness of these interlayers ranges from less than a metre to 30-40 metres. Contacts between these metasedimentary units were observed in a few localities, conformable and parallel to the regionally developed foliation (Plate 1b). The thickest of these units could be traced along strike up to three kilometres, demonstrating that these units are laterally persistent.

Unmappable, thin and discontinuous horizons of medium to coarse-grained amphibolite are sometimes encountered in the granite and quartz diorite gneisses. These amphibolites were noted in various localities throughout the map-area and are of uncertain origin.

II.2.2. Unit 2 is an amphibolitic gneiss that persistently occurs as a thin unit between units 1 and 3 in the Mt., McCrearv and Closs Lake gneiss belts. Most typically, it is composed of biotite, quartz and subequal amounts of hornblende and plagioclase. Magnetite is commonly discernible in outcrop. Modal analysis of four representive thin sections (500 counts per slide) give an average of: clinopyroxene: 3%, biotite: 3%, sphene: 4%, magnetite: 4%, quartz: 15%, hornblende: 33%, and plagioclase: 38%. The relative percentages of quartz and biotite were found to be variable. Accessory minerals include microcline, apatite, garnet and scapolite. The unit is fine to medium-grained, with an average grain size of 1 millimetre and is thinly layered, on a scale of less than a centimetre. Hornblende-rich layers alternate with quartz and plagioclase-rich layers that contain clinopyroxene and accessory scapolite. The hornblende-rich layers are distinctly finer-grained than the adjacent layers. Layers, 10-20 centimetres in thickness of calcitic marble are widely spaced throughout the unit parallel to the finely developed layering, and suggest that these lithologic surfaces may represent primary bedding.

A characteristic of the unit however, is the lithologic variation it displays along regional strike. In places, the unit is

a composite gneiss (Dietrich, 1960), composed of alternating layers, generally 10-20 centimetres thick, of amphibolite and leucocratic granite gneiss. In some exposures of this type, it was found that the amphibolite layers decrease in number towards unit 1 and that mafic-rich granite gneiss layers alternate with the leucocratic layers (Plate 1f) gradational into the granite gneiss of unit 1. Also mapped as unit 1 is a fine to medium-grained, magnetite-bearing, homogeneous amphibolite that lacks any indication of the primary layering discussed above. Though the relationship was never observed in the field, the composite gneiss may represent a gradational zone between the homogeneous amphibolite and the granite gneiss of unit 1.

Though unit 2 exhibits considerable variation along regional strike, consistent amphibolitic composition and stratigraphic position justify designating these exposures as a single map unit.

II.3. Metamorphosed Sedimentary Rocks

II.3.1. Unit 3 is composed predominantly of a uniform, medium to coarse-grained calcitic marble. The unit weathers grey or blue-grey, is white on fresh surfaces and massive. Fine to medium-grained flakes of graphite and a light brown mica (phlogopite?) are ubiquitous. Biotite and diopside are common accessory minerals. Tremolite is a rare accessory. Thin and widely spaced layers, generally 5-10 centimetres thick, composed of biotite, quartz and plagioclase are characteristic in many exposures of unit 3.

Unit 3 contains thin units (less than 10 metres in thickness) of amphibolite, biotite-quartz-plagioclase schist and thinly interlayered

marble and paragneiss. These units do not compose an appreciable proportion of unit 3 and could not be mapped at the scale of Figure 3. In one exposure, a coarsening of texture and the appearance of euhedral plagioclase crystals (or pseudomorphs?) were noted away from the contact with the calcitic marble, indicating that some of the amphibolite is derived from mafic sills or dikes.

II.3.2. Unit 4 is a composite unit comprising slightly foliated to massive, fine to medium-grained, garnet amphibolite interlayered with calcitic marble, layered amphibolite gneiss and K-feldspar-quartz-biotite-plagioclase paragneiss with or without clinopyroxene and hornblende. Garnet amphibolite is the dominant type of the unit, but subunits of calcitic marble, amphibolite gneiss and paragneiss comprise a significant proportion of the unit. Thinly interlayered contacts between the garnet amphibolite and subunits of paragneiss and layered amphibolite gneiss are common. All subunits range in thickness from less than a metre to 4 or 5 metres, and are widespread throughout the garnet amphibolite. The close association between the garnet amphibolite and the subunits of metasediments formed the basis for designating the unit as being sedimentary in origin. A volcanic origin for the garnet amphibolite, particularly for relatively thick horizons, should not be excluded however.

II.4. Metamorphosed Intrusive Rocks

II.4.1. Unit 7 is predominantly a uniform, medium to coarse-grained, biotite-quartz-plagioclase gneiss with or without hornblende (Plate 2a).

Plate 2

Igneous Structures and Textures

Plate 2a- Typical uniform appearance of tonalite orthogneiss (unit 7). North shore of the Madawaska River, 2.5 kilometres southwest of Wabun Lake.

Plate 2b- White anorthosite contained within tonalite orthogneiss (unit 7), containing abundant amphibolite fragments (xenoliths?) with reaction rims of hornblende. Pen tip rests on a dark coloured, relict plagioclase crystal. South shore of the Madawaska River, 0.75 kilometres north of Elbow Lake.

It weathers white to light grey and has a "salt and pepper" appearance on fresh surfaces. Four thin sections of typical specimens were point counted (500 counts per slide) and give an average of: microcline: 2%, biotite: 6%, quartz: 33%, and plagioclase: 59%. The thin sections were not stained, but several slabs of the unit were stained for K-feldspar and confirmed the low percentage obtained from the modal analysis. The classification of Streckeisen (1967) was used to classify the average mode as that of a tonalite. From thin sections, a plagioclase composition of andesine was determined.

Unit 7 has two occurrences in the map-area. North of the Madawaska River, tonalite is in contact with units 1, 3 and 4 and is folded about a large northeast-plunging antiform centered in the vicinity of Mt. McCreary (Fig. 3). Many exposures of tonalite along the Madawaska River, east of Wabun Lake, contain angular fragments of fine-grained amphibolite, similar to the massive amphibolite of unit 4. In these exposures the fragments are small, generally less than 20 centimetres in longest dimension and are folded with hinge lines that parallel the major folds in the area. Reaction rims of medium-grained hornblende are well developed around the amphibolite fragments. These fragments are probably xenoliths of unit 4. Locally within the tonalite, discontinuous amphibolite layers commonly 1 or 2 metres in length, define a composite tonalite-amphibolite gneiss several metres in thickness. These horizons may represent xenolith zones in the tonalite (see Frontispiece).

There are other local variations in the tonalite of the Madawaska River area which could not be mapped at the scale of

Figure 3. Horizons several metres thick comprise subequal amounts of clinopyroxene and plagioclase, with minor amounts of hornblende. Aggregates of fine-grained clinopyroxene with reactions rims of hornblende are surrounded by fine to medium-grained plagioclase; this is interpreted as a relict, coarse ophitic texture. Many of the outcrops along the Madawaska River are anorthositic. The rock is medium-grained, white on fresh and weathered surfaces and composed almost entirely of plagioclase. Widely scattered throughout the rock are coarse-grained single crystals of dark blue plagioclase rimmed by finer-grained white plagioclase (Plate 2b). From oil immersion mounts, the index of refraction was determined for these crystals to be that of labradorite. These crystals are probably relict and may have been phenocrysts. The anorthosite also contains, in places, folded amphibolite fragments (Plate 2b).

There is little doubt of the intrusive origin of the tonalite in the Madawaska River area. The uniform structure (Plate 2a), a mode appropriate to an igneous rock and the local preservation of igneous textures support this conclusion.

The second occurrence of tonalite is in the area mapped on a reconnaissance scale. The structure of this area has been interpreted as a large dome (Figs. 4 and 5), with unit 7 in the core extending from the southern end of Norcan Lake northeast to McGonegal Mountain. Mapping in this area revealed no direct evidence of an intrusive origin. On the basis of lithologic similarity however, the unit here has been correlated with intrusive tonalite in the Madawaska River area. This correlation is most reasonable in the

Bartraw Lake Mountain area (Fig. 3), whereas to the south, beginning in the vicinity of Straddlebug Lake and extending to the antiformal closure centered on the southern end of Norcan Lake, the unit is a grey, coarsley layered, composite amphibolite and migmatitic, biotite-hornblende-K-feldspar-quartz-plagioclase gneiss of unknown origin.

II.4.2. Unit 8 has been mapped as a syenite. The rock is dark red on fresh and weathered surfaces, inequigranular and is a uniform augen gneiss composed of biotite, hornblende, plagioclase and microcline, in order of increasing proportions. Quartz is difficult to recognize in hand specimens and occurs in accessory amounts in thin section. Individual augen are composed of medium-grained microcline; the matrix is fine-grained and composed of the essential mineralogy of the rock. Biotite and inequant hornblende crystals define a planar fabric which is deflected around and sharply outlines the augen. Garnet is an accessory mineral.

Unit 8 has only one occurrence in the area and is confined largely to the core of a major synform in the vicinity of Elbow Lake, south of Barrett Chute dam (Fig. 3). The unit was traced to the northeast where it appears concordant between units 1 and 4 at the closure of an adjacent antiform. Within the core of the synform in the vicinity of Elbow Lake however, the unit is in contact with units 1, 4 and 7 and has a discordant map pattern. The igneous intrusive origin of the unit is inferred from the appropriate igneous mineralogy, uniform texture and structure, and its discordant contact relations

with the enclosing units. The microcline augen are interpreted as recrystallized porphyroclasts derived from feldspar phenocrysts.

II.5. Pegmatites and Mafic Dikes

Much pegmatite occurs in the map area. Only the larger masses are shown as unit 9 on Figure 3. There are two types of pegmatite; white, plagioclase-rich pegmatite and pink, granitic pegmatite.

White pegmatite is most commonly found within calcitic marble (unit 3) and tonalite (unit 7). This pegmatite is undeformed and cross-cuts foliation. Some of the white pegmatite in calcitic marble is clearly deformed however, and an intrusive origin was not established.

Pink granitic pegmatite is most commonly found in unit 1, but was also noted in calcitic marble. This pegmatite most commonly appears undeformed and probably arose by partial melting of the granitic rocks in unit 1. Evidence supporting this contention has been noted in the area mapped on a reconnaissance scale (Fig. 3). Here, much of the Closs Lake Gneiss Belt can be described as a migmatite; granitic pegmatite forms the leucosome phase, and biotite and hornblende-rich granitic selvages and variably sized fragments of quartz-biotite-hornblende-plagioclase gneiss form the melanosome phases. The migmatitic structure is variable, ranging from thin and anastomosing pegmatite veins, to a preponderance of pegmatite containing the various melanosome phases rendering stratigraphy meaningless over wide areas in the Closs Lake Gneiss Belt.

Thin mafic dikes have been noted in two localities in the area, one in unit 1 and another in unit 7. These are very fine-grained,

appear undeformed and cross-cut the regional foliation. In thin section, these dikes are composed of hornblende and plagioclase, are equigranular and appear to have been recrystallized.

III. STRUCTURE

III.1. Introduction

Two sets of major folds, with essentially coaxial hinge lines, have been recognized in the Norcan Lake area and their relative ages have been determined. The earlier folds (F_1) are long-limbed isoclines that have been refolded into tight to isoclinal, upright to steeply overturned F_2 folds. Most of the major structures in the area have been interpreted as F_2 folds, but detailed mapping of the Mt. McCreary Gneiss Belt has led to the recognition of major F_1 folds. In this area, F_1 and F_2 folding relationships have been well established and provide a basis for interpreting the structure of near-by areas where folding relationships are less clear. Minor folds associated with these major folding events have been recorded. A third set of folds (F_3) has also been recognized. These are open, gently inclined folds, but are not a prevalent feature of the area.

Other structures include regionally developed foliation, mineral lineations, boudinage and mullion structure. A well developed mineral foliation is folded around the hinges of F_1 and later folds, suggesting that a Pre-F_1 deformational event is recorded in the area. Observed fabric relationships involving minor folds in some outcrops, are interpreted as additional evidence for a Pre-F_1 deformation.

The abbreviation, F_x, refers to a particular episode of folding (x) in a sequence of relative events. All structural elements associated with a particular folding event are described as belonging to F_x. The criteria used to distinguish these folding events include

superimposed fabric relationships and the fold geometry and orientation considered unique to each folding event. The application of these criteria is discussed in following sections. It was not possible to ascertain the absolute time interval between the successive folding events. Table 2 defines the structural nomenclature used.

III.2. Problems of Interpretation

Coaxial refolding poses a number of problems in interpreting the relative ages of minor structures. Individual minor folds are often difficult to assign to a particular folding episode, as F_1 and F_2 can have nearly the same geometry and axial surface orientation; F_2 folds have near-vertical axial planes and could have rotated F_1 axial surfaces into a similar orientation. In the hinge zones of major F_2 folds, F_1 folds are gently inclined to recumbent, suggesting that isoclinal folds with gently dipping axial surfaces are F_1 structures. On Figures 3 and 4, all folds which could not be assigned to a particular folding event with certainty have been given a characteristic symbol.

Mineral lineations, the long axes of boudins and mullion structure are also difficult to assign to a particular episode of folding. These structures parallel local fold axes, and because refolding is coaxial they do not develop interference patterns. On Figures 3 and 4, no attempt is made to distinguish the episode of folding to which these structures belong.

Because F_1 and F_2 folds are tight to isoclinal, minor fold closures, where fabric relationships can best be studied, were only

Table 2. Definition of Structural Nomenclature

SYMBOL	DEFINITION
S_o :	Compositional layering, probably primary in origin.
Pre-F_1:	Earliest recognized folds folding S_o.
Pre-S_1:	Earliest recognized foliation, defined by the preferred orientation of planar minerals, axial planar to Pre-F_1 folds.
F_1-F_3 :	Relative sequence of fold sets which fold Pre-S_1.
S_1 :	Gneissic layering axial planar to F_1 folds.
S_{1-1} :	Composite foliation defined by Pre-S_1 foliation parallel to S_1 gneissic layering.
$S_x X S_y$:	Intersection lineation, involving S_x and S_y planar elements.

occasionally observed. No S_2 axial plane foliation was noted in the area, and Pre-S_1 and S_1 axial planar foliations are discernible only in the hinge zones of related folds. Thus, vergence relationships involving intersecting foliations were of no value in mapping major structures. Also, the sense of asymmetry of minor folds, though locally useful, was of little value in mapping major structures as too few minor folds were observed. The recognition of major structures then, rests heavily on the mapping of lithologic contacts.

III.3. Mesoscopic Structures

III.3.1. Minor Folds and Fabric Elements Developed throughout the area is a penetrative foliation defined by the preferred orientation of planar minerals; inequant biotite and hornblende crystals being most characteristic of the fabric. In some outcrops (Fig. 9; Plate 3a,b), S_o is isoclinally folded and this fabric is noted axial planar to these folds. All other folds in the area, which includes F_1 through F_3 fold sets, bend this fabric, suggesting its development predates these folding events. Therefore, the fabric is designated as a Pre-F_1 structural element and is referred to as the Pre-S_1 foliation (Table 2).

Figure 9 and Plate 3a and b illustrate how Pre-F_1 folding has transposed S_o essentially parallel with Pre-S_1 foliation. As noted in Figure 9, thin and discontinuous quartz-plagioclase segregations have also formed parallel to the Pre-S_1 foliation defined by the planar mineral fabric. Away from hinge zones of Pre-F_1 folds then, S_o is difficult to discern from Pre-S_1 surfaces.

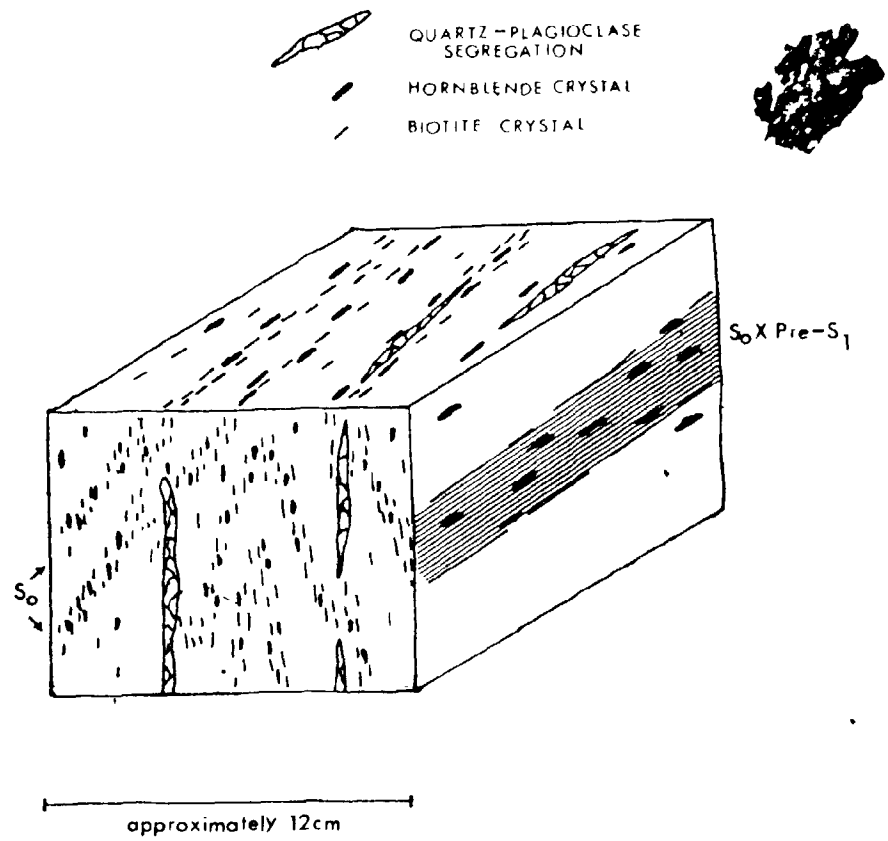


Figure 9. Sketch of a slabbed hand sample illustrating the relationship observed between compositional layering (S_0), Pre-F1 folds and Pre-S1 mineral foliation and incipiently developed gneissoity. Sample is from a paragneiss interlayer in calcitic marble (unit 3).

Plate 3

Minor Structures of the Norcan Lake Area

a) Isoclinal Pre-F1 fold in granite gneiss of unit 1. Compositional layering (S_0) is folded and, though not evident in the photograph, is transected by an axial planar mineral foliation (Pre-S1). Thin layers of calcitic marble (arrows) parallel with the compositional layering are additional evidence that the layering is primary. Scale bar represents 20 centimetres. Green Lake Gneiss Belt, northeast shore of Blithfield Long Lake.

b) Close-up photograph of the core of the Pre-F1 isocline in Plate 3a, illustrating vergence between S_0 and Pre-S1 foliation defined by the planar orientation of biotite and hornblende.

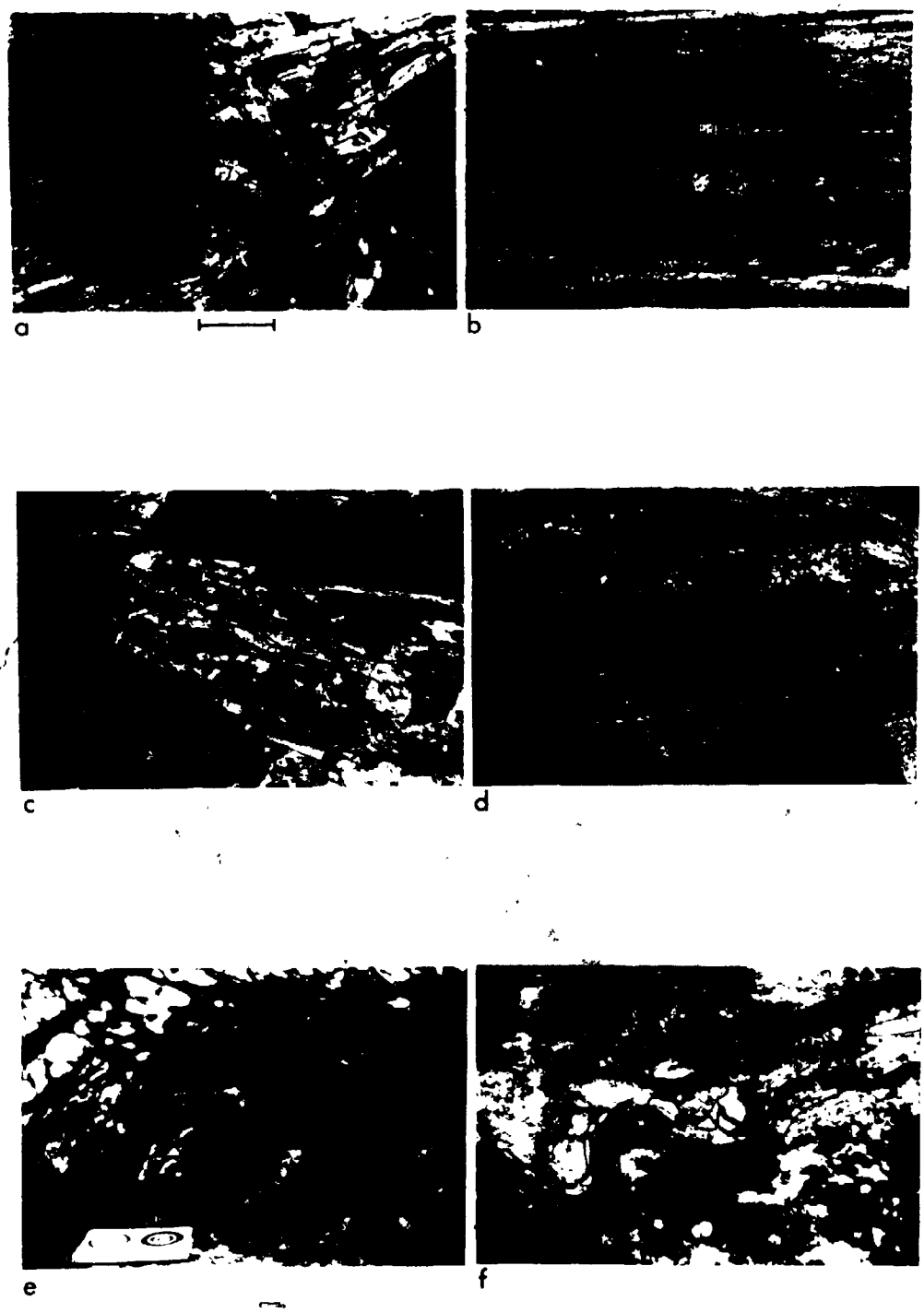
c) Detached F1 fold hinges. Limbs have been strongly attenuated and are now essentially parallel with quartzo-feldspathic segregations (arrows) defining a gneissic layering axial planar to the F1 hinges (S1). Mineral foliation (Pre-S1) is folded about these hinges. Barrett Chute dam exposure of unit 1.

d) F1 folds defined by thin compositional layers. Quartzo-feldspathic segregations of variable thickness (arrows) define an axial planar gneissic layering (S1). Post-tectonic mineral growth, typified by feldspar porphyroblasts (P), has obliterated evidence of Pre-S1 foliation. Barrett Chute dam exposure of unit 1.

e) Type 3 interference pattern developed from coaxial F1 and F2 folding. Folds defined by thin interlayers in calcitic marble (unit 3). Note thickened F1 hinges and the lack of such thickening in the F2 hinges, also, the recumbent F1 orientation little affected by upright F2 folding. East shore of Norcan Lake, 0.6 kilometres southwest of Rock Lake.

f) Photograph of amphibolite layer in calcitic marble (unit 3). Exposure is located at the crest of the Mt. McCreary F2 antiform, east shore of Norcan Lake. Note "parallel" F2 symmetric profiles and the "similar" F1 recumbent profile. The boudinage structure appears to be folded by the F2 folds.

Plate 3


a
b
c
d
e
f

Folds that fold the Pre-S_1 foliation, and have an associated axial plane foliation, have been assigned to F_1. These folds are isoclinal, "similar" in profile, and commonly have attenuated limbs (Plate 3c,d). The axial plane foliation associated with these folds is best described as a "gneissic layering". F_1 fold hinges are bounded by quartzo-feldspathic segregations in an axial planar orientation, and define the gneissic layering (Plate 3c,d). The S_1 gneissic layering tends to be regularly spaced in an outcrop on the scale of a few centimetres (Plate 3c, d), but may vary in thickness (Plate 3d).

Fabric relationships involving Pre-F_1 and F_1 folding are represented best in the gneisses of unit 1. It is only in this unit that Pre-S_1 and S_1 foliations have both been distinguished and their superimposed relationship established. However, F_1 isoclinal folding has transposed the folded surfaces, so that the S_1 gneissic layering has obscured the recognition of Pre-S_1 foliation away from hinge zones of F_1 folds. Therefore it is likely that in outcrops where hinge zones are not exposed, the foliation surface is composite, consisting of Pre-S_1 foliation essentially parallel to S_1. This composite foliation is referred to as S_{1-1} (Table 2).

Well developed foliation is present in all units mapped (excluding unit 9), though post-tectonic recrystallization has partially obscured the fabric, particularly in calcitic marble of unit 3 (IV.3.2.). Pre-S_1 foliation has been distinguished in units 1 and 3 (Fig. 9; Plate 3a,b), but S_1 foliation has been recognized with certainty only in unit 1 (Plate 3c,d). Foliation in units 2,4,7 and 8 is expressed by the preferred orientation of planar minerals

parallel to a variably developed compositional layering, defined by quartzo-feldspathic and mafic-rich segregations. It is assumed that the planar fabric in these latter units represents, for the most part, S_{1-1}.

Minor Pre-F_1 and F_1 folds, distinguished only in units 1 and 3, appear nearly identical in profile. These structures are isoclinal, commonly long-limbed and show relative thickening and thinning along the hinge zones and limbs respectively (Plate 3a,b,c,d; Fig. 9).

F_2 minor structures consist only of folds. These structures are most commonly tight to isoclinal with upright to moderately inclined axial planes. Compared with Pre-F_1 and F_1 folds, little or no thickening is apparent in the hinge zones (Plate 3e,f). Minor F_2 folds have been observed in all units. In calcitic marble (unit 3) open F_2 folds have also been noted. In some exposures of unit 3, both F_1 and F_2 folds with coaxial hinge lines are present and define a Type 3 interference pattern (Plate 3e).

F_2 folds fold S_{1-1}. No axial plane foliation was noted associated with these folds. Isoclinal folding has however, rotated S_{1-1} into an axial planar orientation so that it can be mistaken for a foliation formed by growth of new minerals during F_2 folding. In the Closs Lake Gneiss Belt (Fig. 7), pegmatite believed to have formed by partial melting of unit 1 cross-cuts S_{1-1} and F_2 axial planes.

The only F_3 structures noted in the map-area are open folds with gently inclined axial planes. Mesoscopic F_3 folds are "parallel" in profile. The relative age of these structures has

been determined on the basis of observations made on one exposure of unit 1 near the southwestern synformal closure of the Juniper Creek Basin (structural subarea 1, Fig. 4). Here, an upright F_2 isocline is folded coaxially into an open, gently inclined fold, "parallel" in profile. Other folds with similar orientation and geometry, assigned to F_3 have been noted in units 1 and 3.

III.3.2. Lineations other than hinge lines of minor folds include mineral and mineral aggregate lineations, long axes of boudins and mullion structure developed in competent interlayers in calcitic marble of unit 3.

Mineral lineation is the most abundant of these structures and is noted commonly in units 1,7 and 8. Elongate prisms of hornblende generally form a prominent lineation in the amphibolites found in units 1 and 7, and less commonly in the quartz diorite gneiss of unit 1. Elongate grains and aggregates of quartz define a lineation in the granite gneiss of unit 1, but because of the fine grained nature of the unit the lineation is difficult to measure accurately. Unit 8 has a well developed augen gneiss structure. Individual augen are greatly elongated within the plane of S_{1-1} foliation and define a prominent lineation. A finely spaced ribbing on thin paragneiss interlayers in calcitic marble (unit 3) is here referred to as mullion structure. This structure parallels local hinge lines and is either a finely developed boudinage structure or small amplitude folds confined to the thickness of the individual layers.

F_3 folding has had little effect on the map pattern, and therefore these lineations undoubtedly formed prior to this folding event. The available evidence, discussed below, suggests that lineations formed prior to F_2 folding. As observations of Pre-F_1 structures are limited, it is difficult to assess the significance of Pre-F_1 and F_1 folding events, individually, in regards to the formation of these linear structures.

In contrast to Pre-F_1 and F_1 profiles, F_2 fold profiles consistently have "parallel" geometry on the mesoscopic scale (Plate 3e,f) and for the most part, macroscopic scale (III.6.). This relationship is most evident in the case of unit 1, in which most of the mineral lineations were noted. The "parallel" fold geometry and the lack of an associated axial plane foliation in F_2 folds suggest that much of the penetrative strain on the microscopic scale, which presumably was responsible for the formation of these lineations, was accomplished before F_2 folding. Metamorphic mineral growth during F_2 folding is minimal (IV.3.2.), which supports this conclusion.

In areas where S_{1-1} foliation is gently dipping, upright F_2 folding is considered to have had little effect. In exposures of unit 3 along Norcan Lake, S_{1-1} has 0-25° dips along the crest of the doubly-plunging, Mt. McCreary F_2 antiform (structural subarea 3, Fig. 4). Here, numerous F_1 recumbent isoclines are defined by thin units of paragneiss in calcitic marble. A range of structures from pinch-and-swell to complete boudinage have developed in the competent paragneiss interlayers, the long axes of which tend to parallel F_1

fold hinges, which are commonly detached from their limbs in this area. The prevalence of these structures along the crest of an F_2 structure suggests they developed during an earlier folding event, because such structures are not likely to form at fold closures. F_2 folds in this area do appear to fold these structures (Plate 3f).

These relations indicate that extension has occurred in the axial surfaces of Pre-F_2 folds, with a component of extensional strain perpendicular to the fold axes. Within the plane of the boudinage structure however, the boudins tend to show a variably developed "chocolate tablet" structure, implying a flattening strain is present. The elongate porphyroclasts in unit 8 for example, are interpreted as indicating extension parallel to hinge lines.

III.4. Coaxial Refolding

Evidence supporting the contention that the major F_1 and F_2 fold sets in the area are coaxial exists on the mesoscopic and macroscopic scales. The Type 3 interference pattern traced by stratigraphy in the Mt. McCreary Gneiss Belt (Fig. 7) suggests that early recumbent folds have been refolded by later structures with steeply inclined axial surfaces (F_2), in which the axes of the new folds coincide approximately with the axes of the earlier folds (Ramsay, 1967). In exposures where both F_1 and F_2 minor folds are developed, the same interference pattern exits with parallel hinge lines (Plate 3e).

The attitudes of all hinge lines and lineations are plotted on a stereonet in Figure 10. The points form gently-plunging,

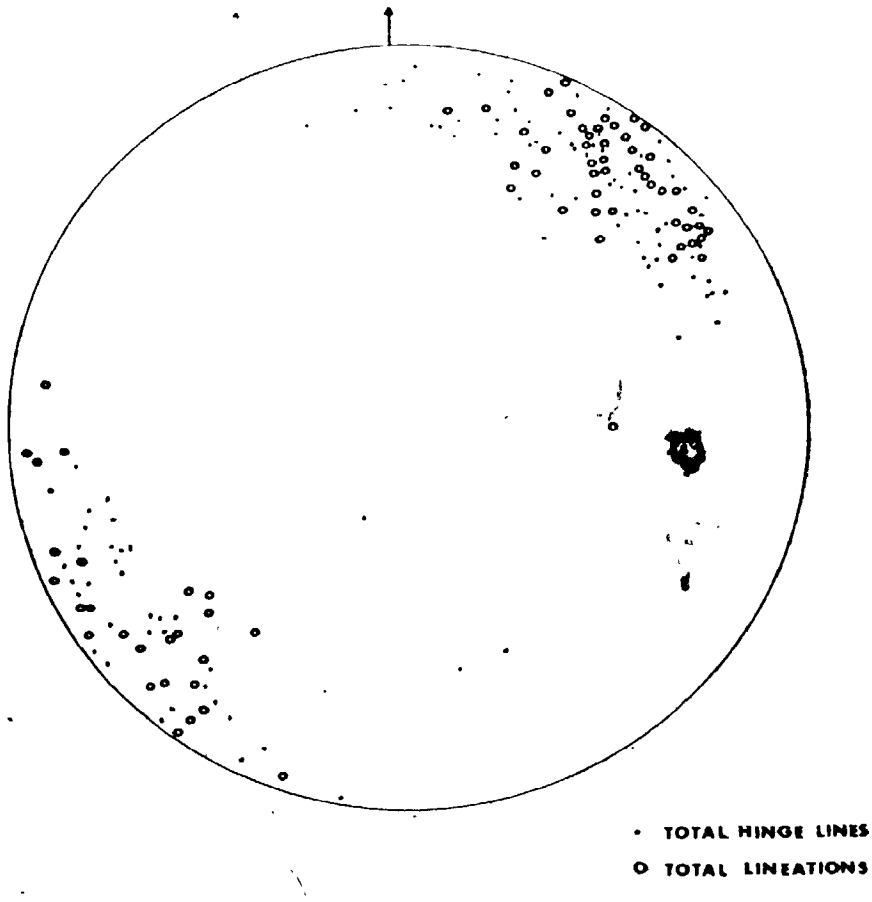


Figure 10. Stereogram containing all hinge lines and lineations measured in the map-area.

northeast and southwest-trending maxima. Both maxima contain F_1 and F_2 structural elements implying near-coaxial deformation. The poles to F_1 minor fold axial planes, and axial planes of folds of uncertain generation, have been plotted on a stereonet in Figure 11. The data are from various outcrops throughout the area and define a great circle, the normal of which plots within the northeast-trending maximum of linear features in Figure 10. Most of these folds have undoubtedly been folded by F_2 folds; the normal in the figure then, can be considered as a regional F_2 axial trend. The figure is thus additional evidence indicating coaxial refolding.

Figure 12 is a stereogram containing F_1 and F_2 hinge lines and poles to F_2 axial planes. The data are from various localities throughout the area and the figure contains only those data for which relative ages could definitely be established. If F_1 and F_2 folding are coaxial, then the maxima defined by the hinge lines should plot on the great circle representing an average regional F_2 axial plane orientation. The average pole to F_2 axial planes in Figure 12 defines a steeply inclined great circle which contains maxima of both F_1 and F_2 hinge lines.

III.5. Descriptions of Structural Subareas

III.5.1. Introduction The structural map (Fig. 4, in pocket) has been divided into five subareas for the purpose of structural analysis. The boundaries for these subareas isolate the major structures in the area. All structural data for each subarea are plotted on individual stereonets in Figure 4. A synoptic diagram (Fig. 13)

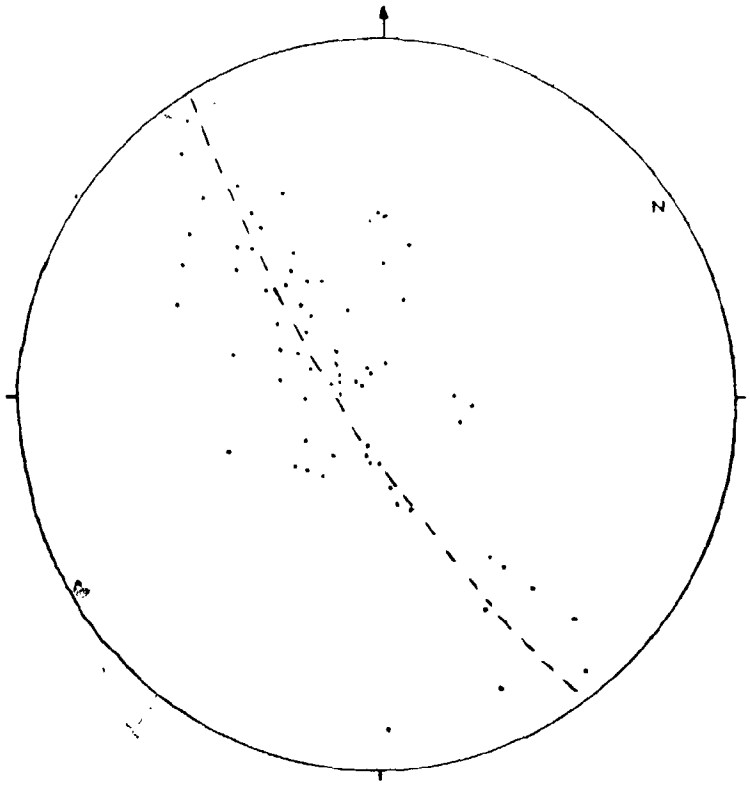


Figure 11. Stereogram containing poles to F_1 axial planes and axial planes of uncertain fold generation. The data are from various localities throughout the map-area. The points define a great circle, the normal of which (N) plots in a position similar to maxima of hinge lines and lineations for the area (see Figs. 10 and 13).

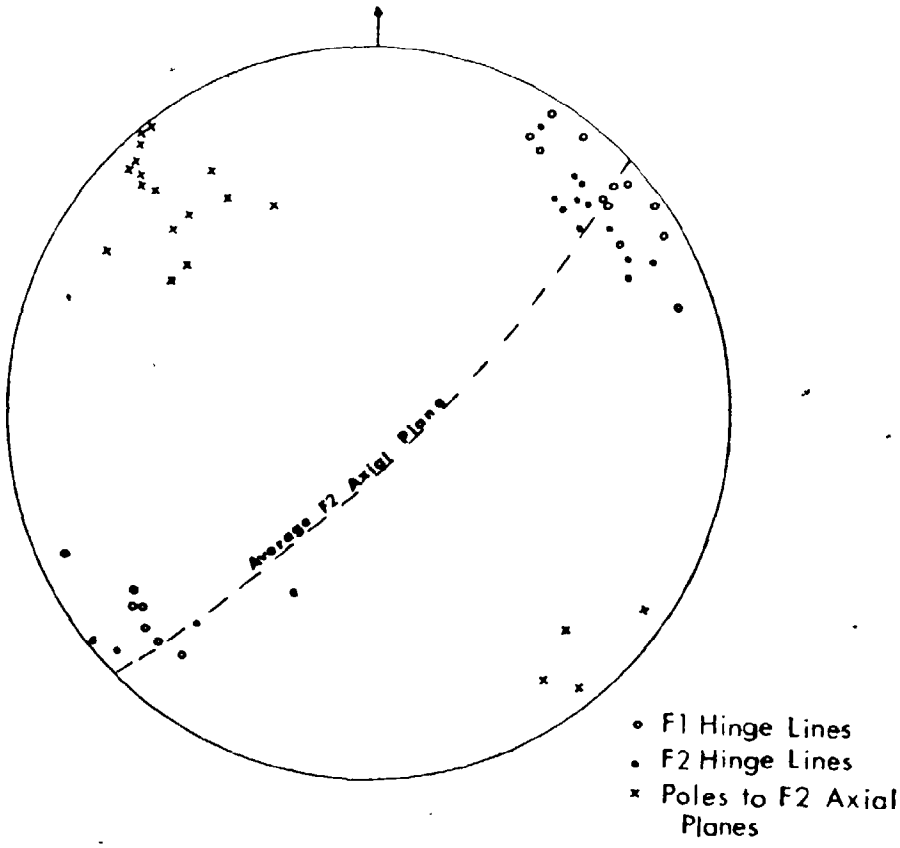


Figure 12. Stereogram containing F_1 and F_2 hinge lines, and poles to F_2 axial planes. The data are from various localities throughout the map-area.

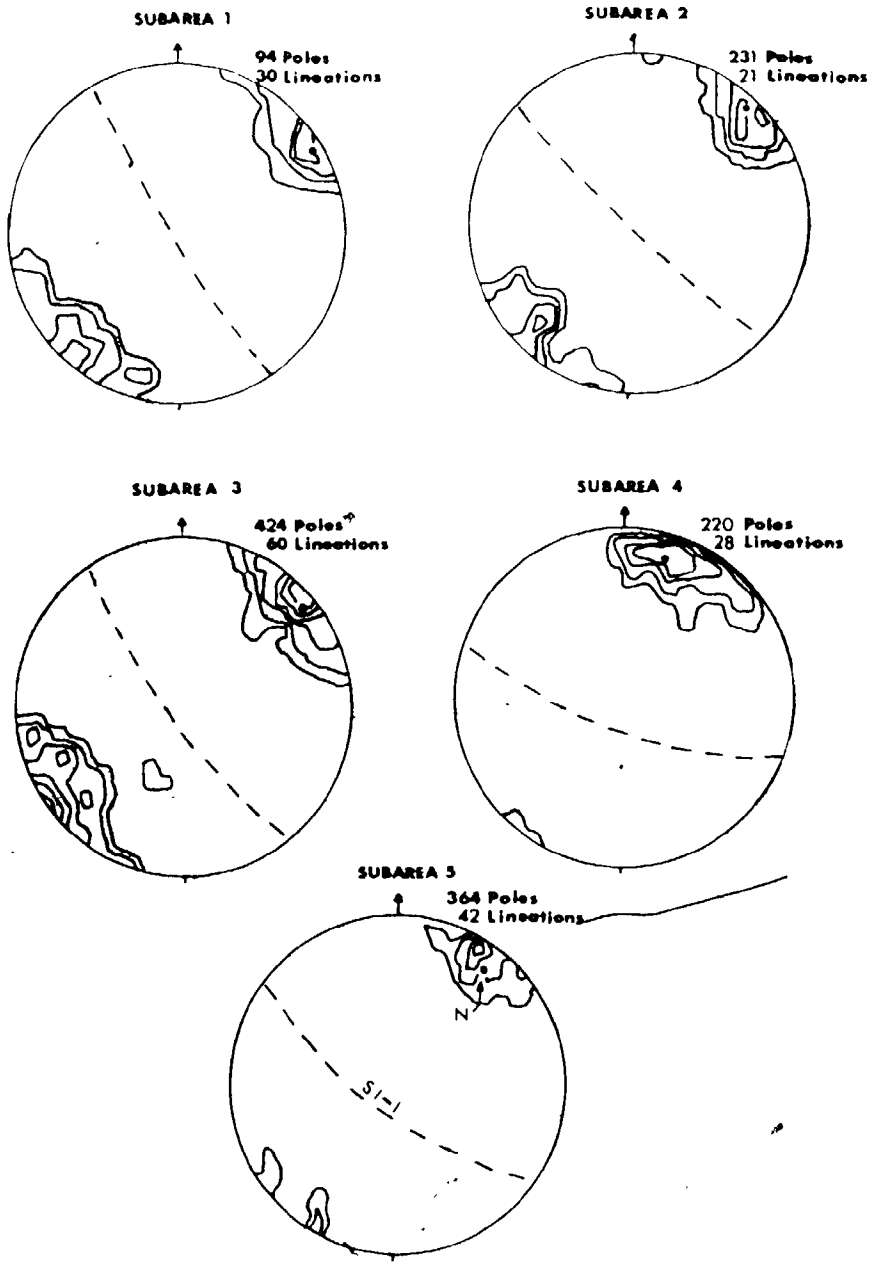


Figure 13. Synoptic diagram for structural subareas 1 through 5, Figure 4. Poles to S1-1 foliations were contoured on other nets; only the average great circle (---) and normal (•) are shown here. Lineations include all measured linear elements for each subarea, and have been contoured here at 0.5, 1.5, 3.5, 5.5, 7.5, and 9.5 percent intervals appropriately.

contains the same data as in Figure 4, contoured and average orientations determined for the various fabric elements. Figure 5 is an axial surface trace map.

III.5.2. Subarea 1; Juniper Creek Basin This subarea contains a doubly-plunging synform defined by the contact between units 1 and 3. The contact is well defined by outcrop control and topography. At the southwestern closure the synform plunges to the northeast and is well exposed. S_{1-1} at the closure has a dip of 23°, probably very close to the plunge of the synform. Several symmetric F_2 folds are present here and have upright axial planes.

The closure at the northeast end of the basin is covered to a large extent by glacial deposits. The presence of a shallow southwest-plunging synform is inferred from the presence of northwest-trending S_{1-1} surfaces that dip gently to the southwest in the area where a closure is suspected from topographic evidence. Also, no exposure of unit 1 exists along Norcan Lake where the unit would be expected to crop out if it continued along strike. Here, calcitic marble (unit 3) is continuously exposed and a southwest-plunging antiform/synform fold pair is defined by the contact of unit 3 with unit 1. The southwest plunge of this fold pair is established by the presence of southwest-dipping S_{1-1} surfaces in the vicinity of the fold closures and by dominant southwest-plunging minor folds.

The stereogram for this subarea (Fig. 13) shows that poles to S_{1-1} define a great circle, with maxima in the northwest and southeast quadrants (Fig. 4). The normal to this great circle plots

within the northeast-trending maximum of hinge lines and lineations. A southwest-trending maximum of linear data is also illustrated. The two maxima of poles to S_{1-1} foliations suggest that the synform is nearly upright in orientation, in agreement with the structural map (Fig. 4). As the structure is traversed across regional strike, the dip direction of S_{1-1} systematically changes from a moderate southeast dip to a moderate northwest dip across the trace of the axial surface.

III.5.3. Subarea 2; Green Lake Synform The dominant structure of subarea 2 is a large synform with the closure located near Green Lake. The structure is defined by the contact between units 1 and 3. S_{1-1} was mapped around the fold closure. Northwest-striking S_{1-1} surfaces dip to the northeast at moderate angles (30-40°). F_2 minor folds and lineations plunge more gently (20°) and are probably more representative of the plunge of the major structure.

A pronounced thickening of unit 1 begins in the vicinity of Green Lake, extending northeast for 2.5 kilometres along strike. The increase of apparent thickness of unit 1 within the synform is probably a result of a plunge reversal of the hinge line. The basinal structure in subarea 1 is well documented. Mineral lineations in the area 2.5 kilometres northeast of Green Lake plunge gently to the southwest.

The stereogram for subarea 2 (Fig. 4) shows two maxima of poles to S_{1-1} foliations in the northwest and southeast quadrants. Hinge lines and lineations form gently-plunging maxima in the

northeast and southwest quadrants. The great circle of poles to S_{1-1} foliations defines a normal which plots within the northeast-trending maximum of linear data (Fig. 13). The two maxima of poles of S_{1-1} foliations reflect a change in the direction of overturning for the structure (Figs. 4 and 5). Southwest of Norcan Lake to Green Lake, S_{1-1} dips consistently and steeply to the northwest. The synform in this area is probably overturned to the southeast. From Norcan Lake northeast to the end of subarea 2, S_{1-1} foliations generally dip steeply to the southwest, suggesting the synform is overturned to the northwest.

III.5.4. Subarea 3; Mt. McCreary Antiform Subarea 3 contains a large, doubly-plunging, F_2 antiform. A Type 3 interference pattern is defined by the stratigraphy and thus, macroscopic F_1 folds are clearly present. The marker horizons for this structure are units 2 and 3. An F_1 fold closure was mapped 2 kilometres southwest of McCreary Lake. Exposure in the immediate vicinity of the closure is poor but it is reasonably established by: the repetition of stratigraphy well mapped in the area; the progressive thinning of unit 1 as the inferred closure is approached; and the continuous exposure of calcitic marble (unit 3) along Norcan Lake which precludes the possibility that unit 1 persists along regional strike.

The F_2 antiformal closure southeast of Mt. McCreary is well exposed. S_{1-1} was mapped around the fold closure where foliations dip to the northeast at 14-20°. Several minor F_2 folds, within units 1 and 2, plunge to the northeast at 14°.

A reversal in the plunge directions of the major structures in subarea 3 exists across a northwest-trending line roughly centered in Norcan Lake. In the area west of Norcan Lake, exposure is rather poor but sufficient to establish major southwest-plunging folds. Foliation in the vicinity of these closures dip to the southwest at 21-25°. Units 1, 2 and 3 are repeated in the domical structure, with calcitic marble (unit 3) as the structurally lowest unit. The axial surface trace of the F_1 structure mapped near McCreary Lake has been interpreted as extending to the southwest, across Norcan Lake, to the synformal closures at Skead Lake and southeast of Sullivan Lake (Figs. 4 and 5). Plunge reversal of this hinge line, along strike of the axial surface trace, has exposed lower structural levels composed of unit 3. Pre-S_1 foliation is folded around these synformal closures and S_1 axial planar gneissosity is developed.

Exposures of unit 3 along Norcan Lake, in the vicinity of the antiformal culmination, reveal evidence of F_1 and F_2 folding. S_{1-1} in calcitic marble generally dips 0-20° and contains numerous F_1 isoclines with recumbent to gently inclined axial surface orientations. F_2 minor folds have markedly steeper axial planes (48-65°) that dip to the southeast and are open to tight structures with little sense of asymmetry (Plate 3f). Though F_2 folds are present, they are uncommon compared to the F_1 structures in this area. F_2 folding has had minimal effect on the orientation of F_1 structures here.

The stereogram for structural subarea 3 shows two maxima of poles to S_{1-1} foliations (Fig. 4). Hinge lines and lineations form two maxima that plunge gently to the northeast and southwest

(Fig. 13). The structural map (Fig. 4) indicates that west of Norcan Lake the structure is overturned to the southeast, while east of Norcan Lake the structure is upright as the limbs dip in opposite directions (Fig. 5).

III.5.5. Subarea 4; Reddys Lake Synform Subarea 4 contains a large synform, the closure of which is located 2.5 kilometres west of Reddys Lake. The subarea also contains a change in regional trend, from northeasterly to northerly in the northeast part of the map-area. The northeastern extent of the Mt. McCreary Antiform has been included in this subarea because of the strong northerly trend present there.

The contact between units 1 and 3 defines the synform. This contact has been continuously traced from the Green Lake Synform northeast into subarea 4 where it is folded about the Mt. McCreary Antiform.

The synformal closure is well depicted on air photographs and was traversed. Exposure of unit 1 at the closure is good. Minor F_2 structures were rarely observed, but S_{1-1} dips to the northeast at 15-18°, which is probably very close to the plunge of the major structure.

Unit 4 is in the core of the synform from just west of Moore Lake northeast to the Madawaska River. Plutonic rocks (units 7 and 8) in the northern part of the subarea have lithologic contacts that appear to be trending at high angles to S_{1-1}. The prominent lineation in unit 8 plunges gently to the north.

The stereogram for the subarea shows a well defined great

circle of poles to S_{1-1} foliations (Fig. 4). The great circle defines a normal that plots within the gently-plunging maximum of lineations and hinge lines. The orientations of the great circle and normal reflect the more northerly trend in the subarea. Poles to S_{1-1} foliations define two maxima on the great circle (Fig. 4). Inspection of the structural map suggests that the maxima represent the two limbs of the synform, that dip in opposite directions indicating the structure is upright.

III.5.6. Subarea 5; Closs Lake Mountain Dome This subarea was mapped on a reconnaissance scale (see inset map, Fig. 3). It is the largest structural subarea and contains fewer data than subareas 1 to 4. As in subarea 4, the change in regional map trend is present.

The structure in the area has been interpreted as a large, doubly-plunging F_2 antiform. The contacts between units 1,2,3 and 7 define the structure. A north-plunging antiform has been mapped south of Barrett Chute dam. S_{1-1} foliations dip 15-25° in the vicinity of the closure. In the southwest part of the subarea near Norcan Lake, a large southwest-plunging antiform has been mapped. The closure is not mapped in detail, but topographic evidence and the presence of northwest-trending S_{1-1} foliations that dip moderately to the southwest in the area of the suspected closure, support the existence of the structure.

Tonalite (unit 7) is mapped in the core of the structure. The contact between unit 1 and the tonalite also defines the antiformal closure south of Barrett Chute dam. In the McGonegal Mountain area (Fig. 4), S_{1-1} has strong westerly trends that dip to the north

at gentle angles and which appear parallel with the lithologic contact. The contact between units 1 and 7 has been traced to the southwest where it is interpreted as being folded by the southwest-plunging antiform southwest of Norcan Lake. The southeast limb of the Closs Lake Mountain Dome remains to be mapped. Unit 3 best defines the structure and additional mapping may reveal the unit is continuous about the dome.

All structural data from subarea 5 have been plotted on steronets shown in Figures 4 and 13. Poles to S_{1-1} define a great circle. Hinge lines and lineations form gently-plunging northeast and southwest-trending maxima.

III.5.7. Summary Two major sets of folds have been recognized in the map-area. Most of the macroscopic structures have been interpreted as tight to isoclinal F_2 folds, that comprise two antiform/synform fold pairs. These structures are well defined by the contact between units 1 and 3. This contact has been traced from the Green Lake Synform (subarea 2), around the Mt. McCreary Antiform (subarea 3), Reddys Lake Synform (subarea 4) and around the northeastern antiform of the Closs Lake Mountain Dome (subarea 5). On the basis of the macroscopic Type 3 interference pattern established, the Mt. McCreary Antiform is clearly an F_2 fold. The adjacent synforms to this structure have been interpreted as complementary F_2 folds. S_{1-1} foliation has been traced around all the major F_2 folds. F_2 minor folds are present at the major F_2 fold closures and have vergence relationships consistent with the macroscopic geometry.

All the macroscopic F_2 folds plunge gently ($14-25^{o}$) to the northeast and southwest. Each subarea, excluding subarea 4, contains a major plunge reversal producing a basin-and-dome map pattern. Hinge lines and lineations form northeast and southwest-plunging maxima in these subareas except for subarea 4, which contains only a northeast-trending maximum. Lineation data illustrate F_1 and F_2 coaxial deformation. Poles to S_{1-1} foliations tend to form a great circle which, for each subarea, defines a normal that plots within the northeast-trending maximum of hinge lines and lineations. This is a reflection of a prevalent northeast plunge direction in the map-area.

The macroscopic F_2 folds are tight to isoclinal, upright to steeply overturned structures. The direction of overturning changes from northwest to southeast in subarea 2, and the Mt. McCreary Antiform (subarea 3) changes from an upright to a southeastwardly overturned structure (Fig. 5). Large F_3 folds, which are apparent on the structural maps (Figs. 4 and 5) and in the down-plunge cross sections (Fig. 6), may be responsible for this variation of macroscopic F_2 axial plane orientations.

III.6. Comparision of F_1, F_2 and F_3 Folding Events

III.6.1. Fold Geometry Subarea 3 contains macroscopic F_1 and F_2 folds and allows a comparision to be made of these fold geometries. Down-plunge cross sections of the area (Fig. 6, in pocket) shows these structures to be very long-limbed isoclines. Exposure of unit 2 at the F_1 closure was not mapped and therefore the amount of

thickening of that unit can not be determined. Considering the "similar" profiles characteristic of F_1 minor fold geometry, thickening of the units in the hinge zones of these long-limbed, major F_1 isoclines, may be considerable.

F_2 folding has refolded the F_1 isoclines to produce a Type 3 interference pattern. The down-plunge section (Fig. 6) is complicated by F_3 folds but it can be seen however, that the Mt. McCreary F_2 antiform approaches a Class 2 fold of similar-type (Ramsay, 1967). However, the thickening of unit 1 is probably more apparent than real in this structure as a constant plunge value was assumed in constructing the sections. The Reddys Lake Synform in subarea 4 appears to approximate a tight Class 1c fold on the down-plunge cross section. Little or no thickening has occurred in the hinge zone of this synform. The Green Lake Synform is an isoclinal F_2 fold, thus both tight and isoclinal F_2 folds have been produced. All F_2 minor folds noted have "parallel" geometry and may be open, tight or isoclinal. F_2 folding probably produced folds more variable in geometry than F_1 folding.

Several macroscopic F_3 folds are apparent on the down-plunge cross sections (Fig. 6) and the structural maps (Figs. 4 and 5) and were also noted in the field. These structures are open, Class 1b folds.

III.6.2. <u>Fabric Elements</u> associated with the early phases of folding in the area are complicated. A penetrative axial planar fabric (Pre-S_1) has been noted in a few exposures to cut the hinges of minor folds (Pre-F_1) that are defined by a compositional layering (S_o),

possibly of primary origin (Plate 3a,b). Pre-S_1 is defined by the preferred orientation of planar minerals. All other folds noted in the area fold such a foliation.

The F_1 fold set represents the first major folding event recognized in the area, and contains folds of macroscopic and mesoscopic scale. The relative age of F_1 folding has been established on the basis of superimposed fabrics. In several outcrops of unit 1, a secondary axial planar fabric (S_1) cuts the hinges of minor folds (F_1) defined by a planar mineral foliation (Pre-S_1). In these exposures, S_1 is best described as a gneissic layering (III.3.1.: Plate 3c,d) which, because of isoclinal F_1 folding, is only distinguishable in the hinge zones of F_1 folds. Nearly all foliation measurements on Figures 3 and 4 are interpreted as representing a composite foliation surface (S_{1-1}), Pre-S_1 being essentially parallel with S_1 away from major F_1 fold closures.

No fabric elements are associated with F_2 or F_3 folds; an axial plane foliation was not observed in these folds, and boudinage structure and mineral lineations are interpreted to have formed before these folding events (III.3.4.).

III.6.3. <u>Fold Orientation</u> F_1 folds are considered to have been recumbent before F_2 folding. The Type 3 interference pattern in the area implies that F_1 axial surfaces were at large angles to the steeply inclined to upright F_2 axial surfaces. Also, in the hinge zones of major F_2 structures where F_1 orientations are least expected to have changed during F_2 folding, F_1 minor folds are gently inclined to recumbent.

Coaxial F_2 folding refolded F_1 folds and their associated foliations into tight to isoclinal, upright to steeply overturned, macroscopic F_2 folds prevalent in the area. The major caxial hinge lines experience plunge reversals through the horizontal, producing gently-plunging, northeast and southwest-trending basins and domes. This structural feature is not a result of cross folding as implied by Ramsay's (1967) Type 1 interference pattern, as no suitable fold set, orientated at large angles to F_1 and F_2 coaxial hinge lines, has been discerned in the map-area. Two possibilities are most likely for explaining this interference pattern: F_1 folds may have been noncylindrical and nonplanar before F_2 folding; or strain may have been inhomogeneous during F_2 folding, resulting in variable rotation of the coaxial hinge lines.

Because of the prevalent F_2 folding in the area, F_1 fold orientation is variable. The stereogram in Figure 11 illustrates how poles to F_1 axial surfaces define a great circle, the normal of which plots in a position similar to normals defined by S_{1-1} foliations folded by major F_2 folds (Fig. 13). Macroscopic F_1 folds defined in structural subarea 3 (Fig. 4) have steep axial surface orientations along the limbs of the Mt. McCreary antiformal refold, but are gently dipping over a significantly large area along the crest of this structure.

III.6.4. Significance of Pre-F_1 Folding Unfortunately, data concerning the Pre-F_1 folding event are limited. It is necessary to postulate the event however, because all the major folding events,

F_1 through F_3, fold a foliation defined by the preferred orientation of planar minerals. The existence in the area of minor folds which do not fold such a foliation, but fold a compositional layering which is probably primary in origin, is additional evidence of a Pre-F_1 deformational event.

The conclusions concerning the Pre-F_1 folding event led to a number of problems when attempting to distinguish Pre-F_1 from later structural elements, particularly those of F_1:

1. Pre-F_1 and F_1 minor folds exhibit the same geometry; both are "similar" in profile and are isoclinal.

2. An axial planar foliation is associated with both Pre-F_1 and F_1 folds. Observations made in this study however, indicate that the two fabrics are distinct; S_1 characterized by a "gneissic layering" distinct from the planar mineral fabric associated with Pre-F_1 folds.

3. Fabric relationships are discernible only in the hinge zones of folds. Isoclinal F_1 folding has transposed the early foliation so that the Pre-F_1 and F_1 axial surfaces are expected, for the most part, to be parallel, even on a regional scale.

IV. METAMORPHISM AND DEFORMATION

IV.1. Introduction

It is beyond the scope of this work to study in detail the metamorphic history of the area. Some observations and interpretations have been made concerning the metamorphism and its relationship to deformation.

IV.2. Metamorphic Grade

Most of the map-area is believed to have been subjected to high grade metamorphism as defined by Winkler (1976), or the upper amphibolite facies of Turner (1968). The Appendix contains a table of mineral assemblages determined in thin section: amphibolitic rocks are composed essentially of green pleochroic hornblende and plagioclase, with minor clinopyroxene and garnet; siliceous calcitic marble is characterized by clinopyroxene (diopside?), with minor tremolite; quartzo-feldspathic rocks locally contain sillimanite and granitic pegmatite believed to have formed by partial melting.

Metamorphic studies by Moore and Thompson (1972), Rivers (1976) and Chappell (1978), have resulted in defining metamorphic index mineral zones for the Kaladar-Dalhousie Trough (Fig. 14). From the lowest grade area (upper greenschist) in the vicinity of Bishop Corners, metamorphic grade increases to the northeast and southeast. Rivers (1976) identified a sillimanite-K-feldspar zone in the vicinity of Clyde Forks (Fig. 14) and confirmed the decrease in grade towards the southwest, as established by Moore and Thompson (1972).

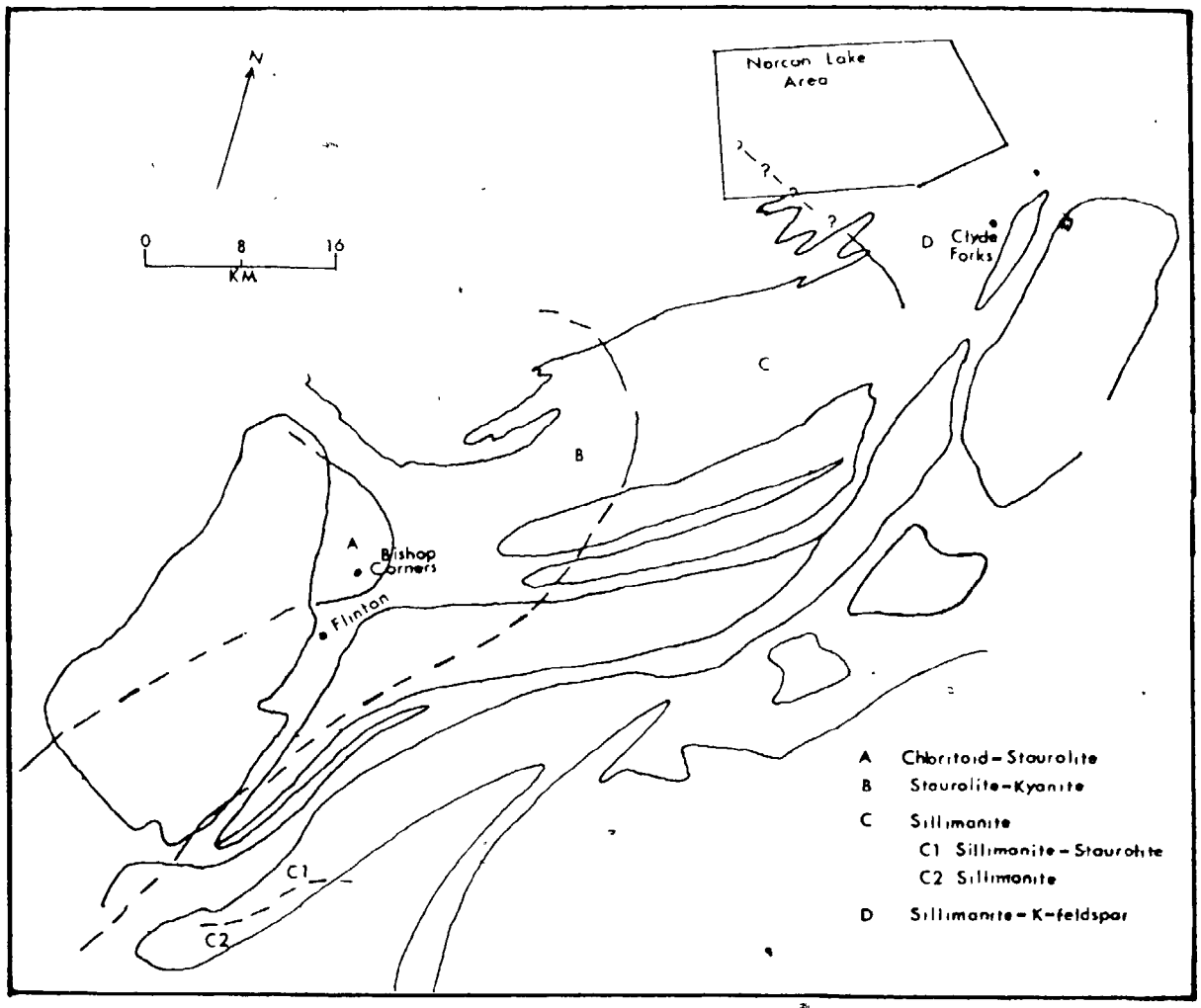


Figure 14. Mineral zones of the Kaladar-Dalhousie Trough. Zonal boundaries are broken where conjectural. After Moore and Thompson (1972), River (1976) and Chappell (1978). The sillimanite-K-feldspar zonal boundary identified by Rivers (1976) is suggested to extend into the Norcan Lake area.

There exists some evidence that an isograd is present in the map-area, corresponding to Rivers' muscovite breakdown reaction in the presence of quartz to produce sillimanite and K-feldspar. In the vicinity of Green Lake (Fig. 3), granite gneiss of the Green Lake Gneiss Belt (Fig. 7) contains the assemblage muscovite-quartz. Four kilometres to the east, granite gneiss of the Closs Lake Gneiss Belt contains sillimanite and K-feldspar, and muscovite is absent. Sampling was inadequate to better define this isograd. Granite gneiss contains sillimanite at several localities in the map-area and apparently it is only in the Green Lake area that the lower grade assemblage occurs (see Appendix). On the basis of these observations, the muscovite breakdown isograd defined by Rivers in the Clyde Forks area, has been suggested to extend into the Norcan Lake area (Fig. 14).

IV.3. Relationship Between Metamorphism and Deformation

IV.3.1. Syntectonic Mineral Growth The composite S_{1-4} foliation (Table 2) is the most pervasive fabric noted in thin section, hand samples and outcrops. On the basis of observations made at fold closures, S_1 is defined as a gneissic layering (III.3.1.; Plate 3c,d) which overprints, particularly away from F_1 hinge zones, a penetrative foliation (Pre-S_1) defined by the preferred orientation of planar minerals (Fig. 9; Plate 3a,b).

The Pre-S_1 foliation is defined by the preferred planar orientation of biotite sheaves and inequant hornblende crystals, and less commonly, tabular quartz and feldspar. Biotite and hornblende crystals are commonly optically orientated; such crystals

defining the foliation go extinct under the same stage position and crossed nicols. As the Pre-S_1 foliation is axial planar to Pre-F_1 folds, the growth of these minerals is considered to have been syntectonic with Pre-F_1 folding. In the granite gneiss of unit 1, sillimanite forms fibrous pods that lie within Pre-S_1, or sillimanite prisms define trains that parallel Pre-S_1. The sillimanite however, was not observed in the axial planar orientation to S_o, as was hornblende and biotite, and thus its syntectonic relationship with respect to Pre-F_1 is questionable. Alternatively, the sillimanite foliation present within these sillimanite-bearing samples may represent S_{1-1}.

The development of quartzo-feldspathic segregations axial planar to F_1 folds (S_1) indicates syntectonic mineral growth during F_1 folding. These segregations are coarser grained than the host rock and are granitic in composition, being composed predominantly of K-feldspar and lesser amounts of quartz and plagioclase. Biotite and hornblende commonly form mafic selvages around the segregations. The composition and texture suggest an origin by partial melting for the segregations (Hyndman, 1972; pp. 293-294).

IV.3.2. <u>Post-S_{1-1} Mineral Growth</u> In nearly all thin sections there is evidence of post-S_{1-1} mineral growth. Crossed biotite flakes, with random planar orientations, lie at high angles to S_{1-1}. Porphyroblasts of hornblende and feldspar with equant grain shapes overgrow S_{1-1} (Plate 3d). Quartz and feldspar occur as equant grains within the felsic segregations defining S_1 (Plate 3d), and equant

hornblende crystals commonly define the mafic segregations. These textures are interpreted as representing post-tectonic mineral growth (Spry, 1969; pp. 257-259).

The extent of this mineral growth is variable. Many amphibolitic rocks show little such evidence, with few crossed biotite crystals and hornblende porphyroblasts; S_{1-1} is readily discernible. In other thin sections however, post-tectonic mineral growth is extensive and S_{1-1} is recognized only by a diffuse gneissic layering. The gneisses which compose unit 1 show a more uniform degree of recrystallization. Gneissic layering is generally evident in thin section, but quartz, feldspar and hornblende have equant grain shapes resulting in a near granoblastic texture. Biotite planar fabric is generally discernable, though the development of much crossed biotite obscures its recognition. Calcitic marble has a granoblastic polygonal texture. Recrystallization has probably obliterated any evidence of foliation, and indeed it is not uncommon to find a lack of foliation in more uniform exposures of calcitic marble.

IV.3.3. <u>Timing of Metamorphism and Deformation</u> If the Pre-S_1 foliation is used as a relative time marker, the Pre-F_1 fold system developed synmetamorphically. The grade of metamorphism attained during this folding event is not known with certainty, though the presence of biotite and hornblende within the Pre-S_1 foliation suggests conditions of the amphibolite facies. The Pre-S_1 foliation is folded by F_1 folds. The granitic segregations representative of partial melting and defining the S_1 foliation, indicate that F_1

folding was also synmetamorphic, and that peak, high grade metamorphism (Winkler, 1976) existed. It is most reasonable to suggest that sillimanite-K-feldspar assemblages formed at this time.

The S_{1-1} foliation is folded by F_2 folds. No axial planar foliation was found associated with this folding event, suggesting that significant new mineral growth did not occur during this time. The post-S_{1-1} mineral growth discussed above therefore, may have taken place between the F_1 and F_2 folding events. The different geometries of these fold sets and the pervasive development of fabric elements during Pre-F_1 and F_1 folding, not associated with F_2 folding (III.3.2.), suggest different rheologic rock properties during these successive events. This difference might be attributed to a decrease in metamorphic grade and is an interpretation consistent with the above conclusion.

It is certainly possible however that an S_2 foliation was developed, but which was subsequently made obscure by the post-tectonic mineral growth. In this case, S_2 would have probably been only weakly developed, as no evidence of its existence was noted. Indeed, it is the lack of an S_2 foliation which makes it difficult to assign a relative age to the post-S_{1-1} mineral growth.

There is evidence that metamorphic grade remained high during F_2 folding. Much of the Closs Lake Gneiss Belt (Fig. 7) is described as a migmatite (II.4.3.). Granitic pegmatite forms the leucosome phase, and variably sized fragments of quartz-biotite-hornblende-plagioclase gneiss and hornblende and biotite-rich granitic selvages from the melanosome phases. The position of the sillimanite-K-feldspar

zone assemblages (see Appendix) appears to correlate with the extensive migmatitic structures observed in the Closs Lake Gneiss Belt. Unit 1 in the Green Lake and Mt. McCreary gneiss belts (Fig. 7) is typically finer grained than the gneisses in the Closs Lake area and evidence of partial melting is observed only locally. The leucosome phase of the migmatite cross-cuts S_{1-1} and F_2 folds and appears undeformed. This relationship is noted on the mesoscopic scale. Also, a large mass of similar intrusive pegmatite was mapped in the hinge zone of a large north-plunging F_2 antiform on the Closs Lake Mountain Dome (structural subarea 5, Fig. 4), suggesting that similar relations exist on the macroscopic scale.

The grade of metamorphism during F_3 folding is unknown. The style of mesoscopic F3 folding is not markedly different from that of F2; F_3 folds are of a "parallel" type and lack associated foliation parallel to thier axial surfaces. This is suggestive of similar rheologic rock properties during F_2 and F_3 folding. The grade of metamorphism during F_3 then, may have not been significantly lower than F_2.

In summary then, it is suggested that amphibolite facies metamorphism was attained during Pre-F_1 folding. Partial melting is evident during F_1 folding and therefore peak metamorphic conditions are interpreted to have existed during this time. Evidence of partial melting during Pre-F_1 folding is lacking, and therefore sillimanite-K-feldspar assemblages may be more indicative of metamorphic conditions during F_1 folding. Metamorphism outlasted F_1 folding and remained at high grade through F_2 folding, as granitic pegmatite, derived by partial melting of unit 1, cross-cuts F_2 folds. New mineral growth during F_2 folding however, was minimal.

V. STRATIGRAPHY

V.1. Structural Succession

No top determinations were possible in the map-area. Stratigraphy in this area must thus be presented in terms of a structural succession, which may not represent the original stratigraphic position of units.

Two major sets of coaxial folds are recognized. Recumbent first phase folds have been refolded into more upright second phase folds, producing a Type 3 interference pattern characteristic of the area. Post-F_2 structures have had little influence on the map pattern and structural sequence of events. No tectonic slides or faults have been recognized, but intense deformation and metamorphism may have obscured such structures. Furthermore, the significance of the Pre-F_1 deformational event in regard to the stratigraphy in the area is unknown.

Figure 15 illustrates the relationship between the inferred structure and structural succession. The central part of the map-area can be divided into two successions. The contacts between units 1,2 and 3 define a Type 3 interference pattern within the Mt. McCreary Antiform. The Mt. McCreary Gneiss Belt (Fig. 7) occupies the cores of refolded F_1 isoclines. Here the structural succession is established, with unit 1 being successively overlain by units 2 and 3 (Fig. 15).

A second succession of units is defined by the contact between units 1 and 3, which outlines the major F_2 structures in the area.

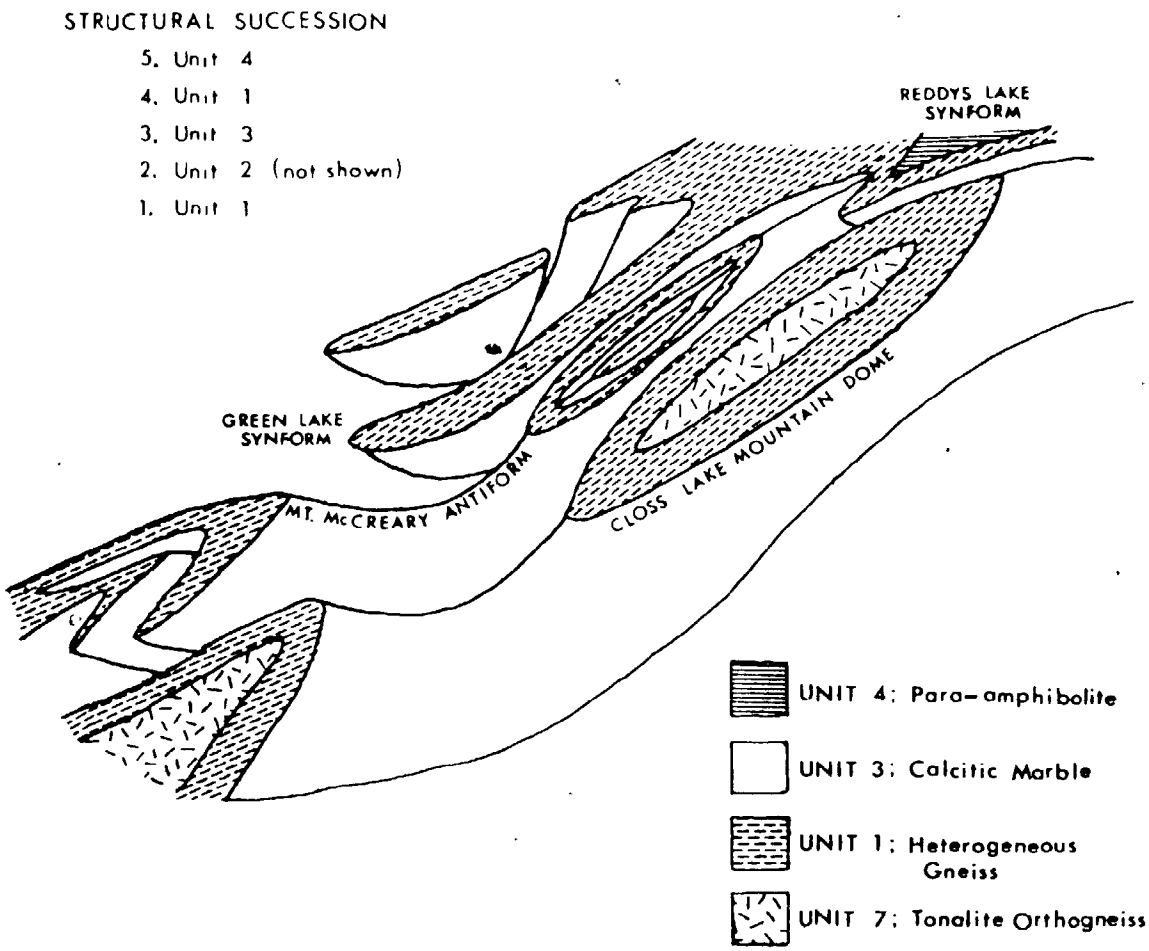


Figure 15. Schematic drawing illustrating the inferred relationship between structure and the structural succession proposed for the Norcan Lake area.

Unit 4 is in the core of the Reddys Lake F_2 synform. The structural succession in the vicinity of this synform begins with unit 3 as the structurally lowest unit overlain by units 1 and 4 (Fig. 15).

The relationship between these two structural successions could not be established with certainty. There are two possibilities:

1. Unit 1 everywhere in the map-area represents the same stratigraphic horizon which, because of refolding, is structurally repeated. As the Reddys Lake Synform is an F_2 structure, the occurrence of unit 1 there must be the result of an earlier fold which is not evident within the boundaries of the map-area.

2. Unit 1 in the map-area occupies two separate stratigraphic horizons. This alternative would imply that the structural succession established within the Mt. McCreary Antiform is immediately overlain by the structurally higher units in the complementary Reddys Lake Synform.

Within the confines of the map-area there is no evidence that the occurrence of unit 1 in the Reddys Lake Synform is the result of earlier folding. The available evidence favors the second possibility:

1. The Green Lake Gneiss Belt differs from the Mt. McCreary Gneiss Belt in that numerous horizons of metasediments are interlayered in the former and not the latter.

2. Unit 2 is a persistent marker between unit 3 and the Mt. McCreary Gneiss Belt, but is absent from the succession within the Reddys Lake Synform.

3. Unit 4, which is lithologically distinct from all other map units in the area, is in contact with the Green Lake Gneiss

Belt in the Reddys Lake Synform. This unit is absent from the succession established within the Mt. McCreary Antiform.

The structural succession interpreted for the area then begins (Fig. 15) with the Mt. McCreary Gneiss Belt (unit 1) in the cores of F_1 isoclines successively overlain by units 2, 3 and the Green Lake Gneiss Belt (unit 1). Unit 4 is the structurally highest unit in the core of the Reddys Lake Synform.

This succession is repeated across the north-plunging antiformal closure on the Closs Lake Mountain Dome (Figs. 4 and 15); the Closs Lake Gneiss Belt then, may represent the same structural level as the Mt. McCreary Gneiss Belt. Here, it can be seen that the succession is relatively thin, beginning from the bottom of unit 3 upwards through the Green Lake Gneiss Belt and including unit 4. An estimate of 750 metres was obtained for the combined thickness of these three units from Figure 4. The thickness of the succession is not obvious in the general vicinity of Norcan Lake; however, the type of tight to isoclinal refolding with subhorizontal hinge lines that exhibit reversals in plunge, would tend to create long and narrow outcrop patterns out of relatively thin stratigraphic units.

V.2. Intrusive Activity

The intrusive origin of units 7 and 8 has been discussed (II.4.). The rocks contain a pervasive foliation assumed to be S_{1-1} (Table 2). F_1 structural elements have been discerned within these units and the rocks have been folded by F_2, so their emplacement preceeds the major deformations recognized in the area, or perhaps

is contemporaneous with F_1 folding. Pre-F_1 fabrics have not been distinguished with certainty in these units, and so the relative timing of emplacement with respect to this event is not known.

The tonalite in the vicinity of the Madawaska River is in contact with the Green Lake Gneiss Belt, and tonalite is also in contact with the Closs Lake Gneiss Belt in the Closs Lake Mountain Dome. Assuming these contacts are intrusive, tonalite has been emplaced into different structural levels. The map pattern of the syenite suggests that it may have been emplaced largely between unit 4 and the Green Lake Gneiss Belt as a sill-like intrusive sheet.

Undeformed pegmatite has intruded all map units. Much of the granitic pegmatite has formed by partial melting of the gneiss belts. Some of the white pegmatite in the area is deformed, and was found most commonly in the calcitic marble and associated with the tonalite. These may represent late stage, potassium deficient pegmatitic phases of the tonalite. Undeformed white pegmatite has also been noted associated with unit 7, and these may represent partial melts of the tonalite.

V.3. Stratigraphy of the Gneiss Belts

The origin of the gneiss belts remains more speculative than in the case of the rest of the map units. The available evidence concerning the unit can be differently interpreted. However, unit 1 is a layered rock closely associated with metasediments and its origin is thus more readily interpreted as a layered complex of sedimentary and volcanic, rather than plutonic rocks. The principal

points of evidence are:

1. Unit 1 is a coarsely interlayered complex of fine to medium-grained, granite and quartz diorite gneiss. The thicknesses of these interlayers are variable, but generally on the order of several tens of metres. Most typically, the granite gneiss is a uniform foliated rock (Plate 1c) that varies little in modal mineralogy throughout the map-area (Fig. 8; see Appendix). Locally however, the granite gneiss contains a compositional layering that predates the development of the Pre-S_1 foliation (Plates 1f; 3a,b), and is probably primary in origin. In contrast, the quartz diorite gneiss is a more variable lithology and is locally gradational into garnetiferous paragneiss (Plate 1e).

2. In the Mt. McCreary Gneiss Belt, unit 2 occurs as a thin stratigraphic horizon for over 15 kilometres along strike. On the northwest limb of the Closs Lake Mountain Dome, unit 2 is at a similar stratigraphic position for nearly 10 kilometres along strike.

3. Unit 1 forms two separate stratigraphic horizons in the structural succession. The Green Lake Gneiss Belt is interpreted to be structurally higher than the Mt. McCreary Gneiss Belt; interlayers of metasedimentary rocks have been noted in the former but not in the latter of these belts (II.2.1; Plate 1b).

A plutonic origin is considered least probable. The known plutonic rocks of the area are medium-grained, inequigranular, uniform gneisses with locally preserved relict textures. They are in marked contrast with the more heterogeneous and finer grained gneisses of unit 1, as well as the stratigraphic evidence discussed above.

Alternatively however, unit 1 may represent shallow.conformable intrusive sheets. Such an origin would account for the stratigraphic relationships discussed above for the unit, except perhaps for its coarsely layered nature. Some mechanism, such as multiple injections from a differentiating magma chamber or a process of differentiation within the intrusive sheet itself, is needed to account for the coarse layering in the unit. It is possible that the quartz diorite gneiss is a metasediment, as it locally grades into garnetiferous paragneiss. In this case, granite gneiss is the only igneous rock contained in unit 1 and the interlayers of this type may represent sills and/or dikes in the quartz diorite gneiss.

A sedimentary origin for the complex is also considered unlikely, particularly for the granite gneiss, though such an origin would account for the stratigraphic relationships discussed. It seems unlikely that thick and laterally extensive sedimentary deposits would occur that, when recrystallized during metamorphism, would produce the fairly uniform granitic modes observed. Significant variation would be expected in the quartz to feldspar ratios. Also, when the granite and quartz diorite gneisses are considered individually, these units most typically are uniform foliated gneisses that lack compositional layering (except for that which can be explained by metamorphic differentiation).

The writer believes that unit 1 represents an interlayered complex of volcanic and sedimentary rocks. All the stratigraphic relationships are easily reconciled with this interpretation. For the most part, unit 1 is suggested to represent ash flows that

approach rhyolitic and dacitic compositions. If the unit was composed entirely of acidic flows, one would expect to find coarse breccia textures frequently associated with such deposits, particularly rhyolites. The possibility of intense deformation and metamorphism obscuring such textures can not be excluded however. Relatively fine-grained ash flows commonly form thick, laterally extensive deposits. Cyclic volcanism can explain the coarsely layered nature of the unit. Sedimentary horizons within unit 1 would represent periods of volcanic quiescence. Shallow intrusive and flow equivalents of the ash flows are probably represented.

Acidic volcanics have previously been recognized in the Kaladar-Dalhousie Trough. The upper part of the volcanic succession studied by Sethuraman (1970) and Sethuraman and Moore (1973), in the vicinity of Bishop Corners, contains andesites and felsic pyroclastics. Ayer (1979) mapped a volcanic complex in the vicinity of Mazinaw Lake where an appreciable portion of the succession is composed of interlayered rhyolites and andesites. Relict textures suggest the rhyolites were deposited as ash flows. Intrusive equivalents to the complex were also recognized. In these areas, it has been demonstrated that the Flinton Group metasediments unconformably overlie the Hermon Group metavolcanics, and that their deposition postdates all major igneous activity (Moore and Thompson, 1972). As the rocks of the Norcan Lake area are interpreted as comprising a succession of metavolcanic and marble-rich metasediments, it is suggested that they correlate with Hermon and Mayo Group rocks and thus are pre-Flinton Group in age.

Carl and VanDiver (1975) have reinterpreted the granitic bodies in the N.W. Adirondacks, originally interpreted as igneous phacoliths by Buddington (1929), as ash flow deposits. These rocks are coarsely layered and have been suggested to represent the structurally lowest unit in the region, being overlain by a thick succession of marble and paragneiss (Lewis, 1969; Foose, 1974; Foose and Carl, 1977).

V.4. Type Localities

The northwest-trending shores of Norcan Lake provide an excellent cross section of the map-area; exposure is nearly continuous and the line of section transects most of the significant major structures and all of the heterogeneous gneiss belts (Fig. 7). Excellent exposures of units 1 and 3 exist in the core of the Mt. McCreary F_2 antiform (structural subarea 3; Fig. 4) and here, F_1 and F_2 folding relationships are well displayed.

A second type locality is suggested in the vicinity of Blithfield Long Lake, southeast of Barrett Chute dam (Fig. 3). The structural succession interpreted for the area is repeated across the northeast limb and fold hinge of the Closs Lake Mountain Dome (structural subarea 5, Fig. 4). All the map units are represented here, excluding the tonalite orthogneiss (unit 7). Key relationships pertinent to the origin of unit 1 are typified by the contact relations between the Green Lake Gneiss Belt and enclosing units. Exposure is good, particularly along the shores of Blithfield, Mud and Gling lakes. With a minimal amount of foot-work, the structural succession proposed for the area can be examined. A more detailed map of this type locality is provided in the Appendix.

VI. REGIONAL COMPARISONS

VI.1. Introduction

The results of this study will here by compared primarily with the work of Venkitasurbramanyan (1969), Thompson (1972), Rivers (1976) and Chappell (1978), who have studied the deformational history of parts of the Kaladar-Dalhousie Trough (Fig. 16). The results of Rivers, immediately to the south of the map-area, are an obvious source of comparison with those from this study. The ability to correlate structures along the strike length of the Kaladar-Dalhousie Trough will be emphasized. Table 3 summarizes the deformational history interpreted for the Norcan Lake area; major structures in the Kaladar-Dalhousie Trough are outlined in Figure 16; and the result from previous studies concerning the relative timing of metamorphism and deformation are compared with those from the Norcan Lake area in Figure 17.

VI.2. Structural Correlation Within the Kaladar-Dalhousie Trough

VI.2.1. Pre-F_1 Structures Thompson (1972) and Chappell (1978) in areas B and C respectively (Fig. 16), discuss evidence of a deformational event which precedes their F_1 folding events. This deformation is marked by a mineral foliation defined by the preferred planar orientation of chlorite, biotite and amphibole, and was noted in the Hermon Group metavolcanics. Thompson (1972) and Chappell (1978) report evidence establishing this deformational event as predating deposition of the Flinton Group metasediments. Locally, older rocks beneath the

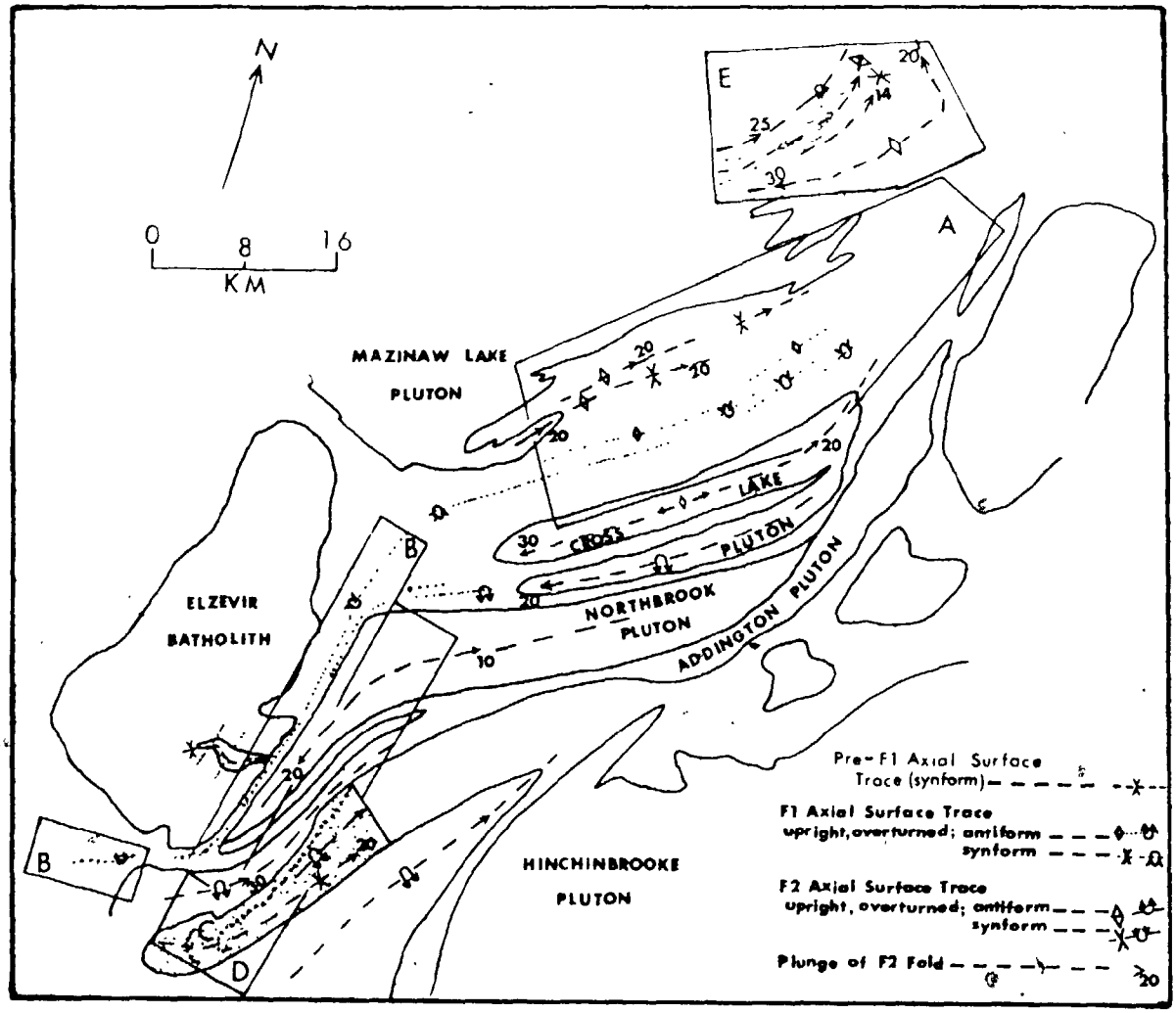


Figure 16. Correlation of major structures within the Kaladar-Dalhousie Trough. Compiled from McDonald (1970), Moore and Thompson (1972) and from study areas A (Rivers, 1976), B (Thompson, 1972), C (Chappell, 1978), D (Venkitasubramanyan, 1969) and E, this study. For clarity, area C has been stippled.

FOLD SYSTEM	OBSERVED ELEMENTS	MINOR STRUCTURES			MAJOR FOLDS			METAMORPHIC GRADE
		Profile	Axial Surface Orientation	Plunge of Fold Axes; Lineation	Profile	Axial Surface Orientation	Plunge	
Pre-F_1	Axial foliation (Pre-S1); Fold axial planes and hinge lines; Extension lineation(?); SoXPre-S_1.	Long-limbed isoclines; "similar" type.	Variable	Northeasterly trends, shallow plunges (few data).				Medium Grade(?)
F_1	Axial foliation (S1); Fold axial planes and hinge lines; Extension lineation; Pre-S1XS1.	Long-limbed isoclines; Detached hinges; "similar" type.	Variable	000-60°-trend, 0-30°-plunge to the NE/SW.	Long-limbed isoclines.	Variable, generally steeply overturned to the NW/SE.	30-60° trend, 0-30° plunge to the NE/SW.	High Grade
F_2	Fold axial planes and hinge lines.	Open to isoclinal; "Parallel" type.	000-50°-strike, moderate to steep dips NW/SE.	000-50°-trend, 0-30°-plunge to the NE/SW.	Tight to isoclinal; Upright to overturned; "Parallel" type.	000-50°-strike steep NW/SE dips.	000-50° trend, 0-30° plunge to the NE/SW.	High Grade
F_3	Fold axial planes and hinge lines.	Open; "Parallel" type.	Gently inclined	000-60°-trend, 0-30°-plunge to the NE and SW(?).	Open; "Parallel" type.	Gently inclined	Shallow, NE and SW plunges (?).	?

Table 3: Summary of the structural history inferred for the Norcan Lake area.

unconformity contain the early foliation parallel with the F_1 Flinton Syncline, whereas above the unconformity, F_1 folds developed in the Flinton Group fold only primary bedding.

The earliest deformational event in the Norcan Lake area is marked by a penetrative mineral foliation axial planar to minor folds defined by compositional layering. No major folds associated with this event have been recognized in the area. Chappell (1978; pp. 76-78) suggests that an isolated body of metavolcanic rocks in the Elzevir Batholith defines a Pre-F_1 synform; the Pre-Flinton foliation contained within these rocks has an east-west trend which becomes progressively rotated into parallelism with S_1 foliation as the Flinton Syncline is approached (Fig. 16). Chappell suggested that the early foliation may be related to the emplacement of the Elzevir Batholith.

VI.2.2. <u>F_1 Structures</u> There is general agreement among the interpretations concerning the first phase of major deformation recognized in the region. In areas A to D (Fig. 16), this deformation resulted in the development of a pervasive and penetrative mineral foliation that has transposed primary compositional layering. Large scale, tight to isoclinal folds were produced at this time and have been recognized in all these areas. Before the later folding events, these folds and associated S_1 foliations are thought to have been gently inclined to recumbent in all areas except for B, where Thompson (1972) considers that S_1 was steep in this area and that later folding events have had little effect on its orientation. It

has been suggested that the close proximity of the Elzevir and Northbrook plutons have influenced the development of structures in the area (Rivers, 1976; Chappell, 1978).

The major F_1 structures in the Norcan Lake area are also considered to have been recumbent prior to later folding events. In contrast with F_1 folds in areas A to D (Fig. 16), the early Pre-S_1 foliation is folded by these structures. Chappell (1978) suggested that the open warps of the isolated volcanic body in the Elzevir Batholith, which may define a Pre-F_1 synform, is a result of F_1 folding (Fig. 16).

In all areas except for B, F_1 hinge lines are considered to have shallow plunges. In Flinton Group conglomerate, Thompson (1972) found a prominent L_1 elongation lineation to be commonly steep, and that F_1 hinge lines have variable orientations. He concluded that heterogeneous strain during folding caused a variable rotation of F_1 hinge lines. Rivers (1976) concluded that a prominent elongation lineation formed during F_1 folding. From strain analysis, he found that F_1 profiles are considerably flattened and concluded that the elongation lineation formed parallel to gently plunging F_1 hinge lines. Conclusions compatible with those of Rivers (1976) have been suggested in this study. Chappell (1978) did not find an L_1 elongation lineation in area C (Fig. 16).

Detailed mapping of the Clare River Structure by Chappell (1978) (Area C, Fig. 16), led to the recognition of a major low angle fault associated with F_1 folding in the area. Psutka (1976) also suggested that a thrust fault accompanied F_1 folding in an area immediately north of the Clare River Structure.

VI.2.3. F_2 Structures Major folds of the second generation in the Norcan Lake area are coaxial with the earlier major structures and are upright to steeply inclined. This observation is in agreement with those of Rivers (1976) and Chappell (1978) in areas A and C respectively (Fig. 16). F_1 and F_2 coaxial folding produced a Type 3 interference pattern, of which large scale examples have been documented in the Norcan Lake area and Clare River Structure. In these two areas, F_1 axial surface traces are clearly folded by second phase structures. The lack of such documentation in other areas is not unexpected; because the folded surfaces have been transposed parallel to the regional S_1 foliation, F_1 hinge zones are difficult to define in the field.

Venkitasurbramanyan (1969) also records major upright F_2 folds in area D (Fig. 16), but concludes that F_1 and F_2 folding was not coaxial; F_1 fold are considered to have had northwest trends whereas F_2 folds trend northeast. In area B, F_2 folds are restricted to the mesoscopic scale and have variable orientations, probably a result of inhomogeneous strain during F_2 folding and a range of orientations that S_1 surfaces had prior to this event (Thompson, 1972).

Fabrics related to F_2 folding are subordinate compared with those produced during F_1 folding. A weak and locally developed S_2 foliation is noted in areas A to C (Fig. 16). Rivers (1976) finds that an L_2 intersection lineation (S_1XS_2) is only weakly developed in area A and that F_2 fold profiles have parallel geometry, showing considerably less flattening than F_1 folds. These conclusions are

in good agreement with the present findings in the Norcan Lake area. A strong L_2 mullion lineation is noted in area C (Fig. 16). The mullion lineation in the Norcan Lake area could also be an F_2 structure.

Macroscopic F_2 folds are common in the Kaladar-Dalhousie Trough (Fig. 16). Plutonic rocks occupy the cores of major F_2 antiforms. The Clare River Structure (Area C, Fig. 16) is a major F_2 synform developed in the supracrustal rocks, and several major F_2 antiform/synform fold pairs are noted in area A. All of these structures are generally tight to isoclinal with upright to steeply overturned axial surfaces, that plunge gently to the northeast and southwest. A similar map pattern is present in the Norcan Lake area; doubly-plunging F_2 folds prevail and tonalite orthogneiss occupies the core of the Closs Lake Mountain Dome. Chappell (1978) noted that the major plutons of the region do not appear to have been significantly incorporated into the supracrustal rocks by isoclinal F_1 folding. Yet there is little doubt as to their involvement in the tight to isoclinal, upright F_2 folding. The recognition of significant decollement surfaces associated with recumbent F_1 folding (Psutka, 1976; Chappell, 1978) may be important in this respect. It is also possible however, that all of the major F_1 structures have not been recognized.

VI.2.4. <u>Later Structures</u> Shallow-plunging, northeast-trending folds with gently inclined axial surfaces have been noted as F_3 folds by Rivers (1976) in area A (Fig. 16). These structures are parallel in profile, show no fabric development and have not greatly influenced the map pattern. This is in accordance with the results

from this study, though macroscopic examples of these structures may have influenced the direction of overturning of upright F_2 folds.

Thompson (1972) records a similar event in area B (Fig. 16) but is designated as F_4 in his sequence. He also records northwest-trending, vertical F_3 folds and suggested that these may be related to the major northwest-trending, open folds which bend the major F_2 axial surface traces in the area. Chappell (1978) may have similar structures in area B (Fig. 16). It is possible that the change in regional trend in the Norcan Lake area is a result of northwest-trending folds, but such minor folds of this orientation have not been observed.

VI.2.5. Conclusions As a result of the detailed studies cited, much has emerged concerning the structural history of the Kaladar-Dalhousie Trough.

Two major episodes of deformation are widely evident in this region. First phase recumbent folds were formed; a penetrative regional foliation was formed at this time and tectonic slides have been recognized associated with this folding. These structures were refolded into gently-plunging, northeast and southwest-trending, tight to isoclinal, more upright major folds. In area B and D however (Fig. 16), the narrow outcrop width of the supracrustal belts between the Elzevir and Northbrook plutons has probably influenced structural development; only first phase structures are recognized in area B, which have steep orientations believed to have been little effected by later folding events (Thompson, 1972). F_1 and F_2 folding may not

have been coaxial throughout area D, though both major folding events are recognized (Venkitasubramanyan, 1969). Divi and Fyson (1973), in the Hastings Basin near Bancroft (Fig. 2), describe how fold geometry is considerably influenced by the near-presence of a granitic pluton.

The results of the work in the Norcan Lake area support the contention that regionally developed recumbent structures have been refolded into prevalent, more upright folds. There is little doubt that F_1 and F_2 folds defined in the Norcan Lake area correlate with those of Venkitasubramanyan (1969), Rivers (1976) and Chappell (1978) in the supracrustal belts to the south. In the Norcan Lake area however, Pre-F_2 folding, for the most part, is more complicated than in these study areas. Major F_1 folds bend a planar mineral foliation that is the axial surface of earlier folds defined by compositional layering. No macroscopic Pre-F_1 folds have been recognized within the map-area, but the prevalence of the Pre-S_1 foliation in the area is evidence that such structures may exist.

Pre-F_1 structural elements, including a mineral foliation and possibly a major synform (Fig. 16), have been locally noted by Thompson (1972) and Chappell (1978). They record sufficient evidence establishing this event as predating deposition of the Flinton Group metasediments. It is suggested that the Pre-F_1 event recognized in the Norcan Lake area may correlate with the pre-Flinton Group deformation discussed by Thompson and Chappell. The relative timing of these two events as an early deformation, that precedes a well established structural sequence involving F_1 and F_2 folding in various parts of the Kaladar-Dalhousie Trough, lends support to this correlation.

The effects of Pre-F_1 deformation is not everywhere recognized in rocks older than the Flinton Group. F_1 folds developed in both Flinton and Hermon Group rocks of the Clare River Structure (Area C, Fig. 16) fold only primary compositional layering (Chappell, 1978). Rivers (1976) did not find evidence for a Pre-F_1 event in area A (Fig. 16), though many of his detailed observations were restricted to rocks of the Fernleigh Belt, which may correlate with the Flinton Group (Rivers, 1976). The available evidence suggests that the effects of Pre-F_1 deformation may be variable in the region.

VI.2. Comparisons With Other Segments of the Central Metasedimentary Belt Early recumbent structures have also been recognized in the supracrustal rocks of the Hastings Basin (Fig. 2). Divi (1972) and Carmichael (1968), in areas near Bancroft and 40 kilometres south of Bancroft respectively, consider that F_1 folds were gently inclined prior to F_2 folding. In these areas, F_2 folds are steeply inclined and are considered coaxial with F_1 folds. Best (1966), also in the vicinity of Bancroft, reached similar conclusions concerning F_1 and F_2 fold orientations, but the axial directions are not considered coaxial.

Wynne-Edwards (1967) in the Westport map-area of the Frontenac Axis (Fig. 2), has also mapped coaxial refolds in the supracrustal rocks. The axial surfaces of these refolds are steeply dipping, thus analogous with the F_2 folds previously discussed.

In the N.W. Adirondacks (Fig. 2), structural studies by Lewis (1969), Brocoum (1971) and Foose (1974), all document evidence for refolded structures. Two major sets of folds are noted by these

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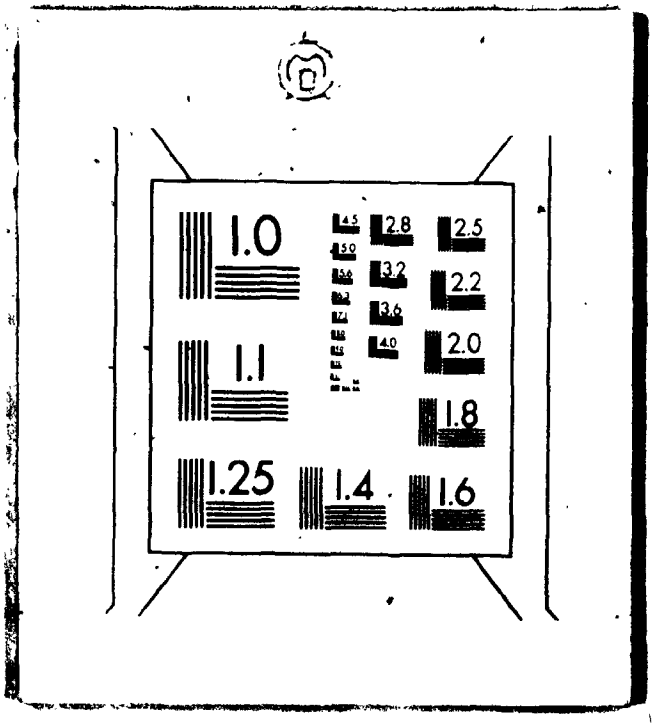
OF / DE



workers; Lewis (1969) considers the refolds to be coaxial, whereas Brocoum (1971) and Foose (1974) do not.

VI.3. Metamorphism and Deformation

Metamorphism in the supracrustal belts of the Kaladar-Dalhousie Trough has been studied by Hounslow and Moore (1967), Moore (1967), Sethuraman (1970), Psutka (1976), Thompson (1972), Rivers (1976) and Chappell (1978). Though there is local evidence of a metamorphism associated with the deformational event which predated the Flinton Group, all workers cite evidence of a single major, regional metamorphic event. Figure 17 is an illustration showing the relative timing between these metamorphic events and the major phases of deformation recognized in the Kaladar-Dalhousie Trough.

There is general agreement among the interpretations concerning the relationship between deformation and metamorphism. Folding during F_1 was accompanied by extensive metamorphic mineral growth which produced the most pervasive foliation in the region, mapped as S_1. Metamorphism outlasted F_1 folding, though the timing between the culmination and the F_1 and F_2 events seems to be variable in the region (Fig. 17); with the culmination having been attained during F_1, or between the F_1 and F_2 folding events. In those areas where the culmination is considered to be post-F_1, the index minerals are found to overprint the S_1 foliation.

Thompson (1972), Rivers (1976) and Chappell (1978) all note that no penetrative S_2 foliation is associated with F_2 folds, and

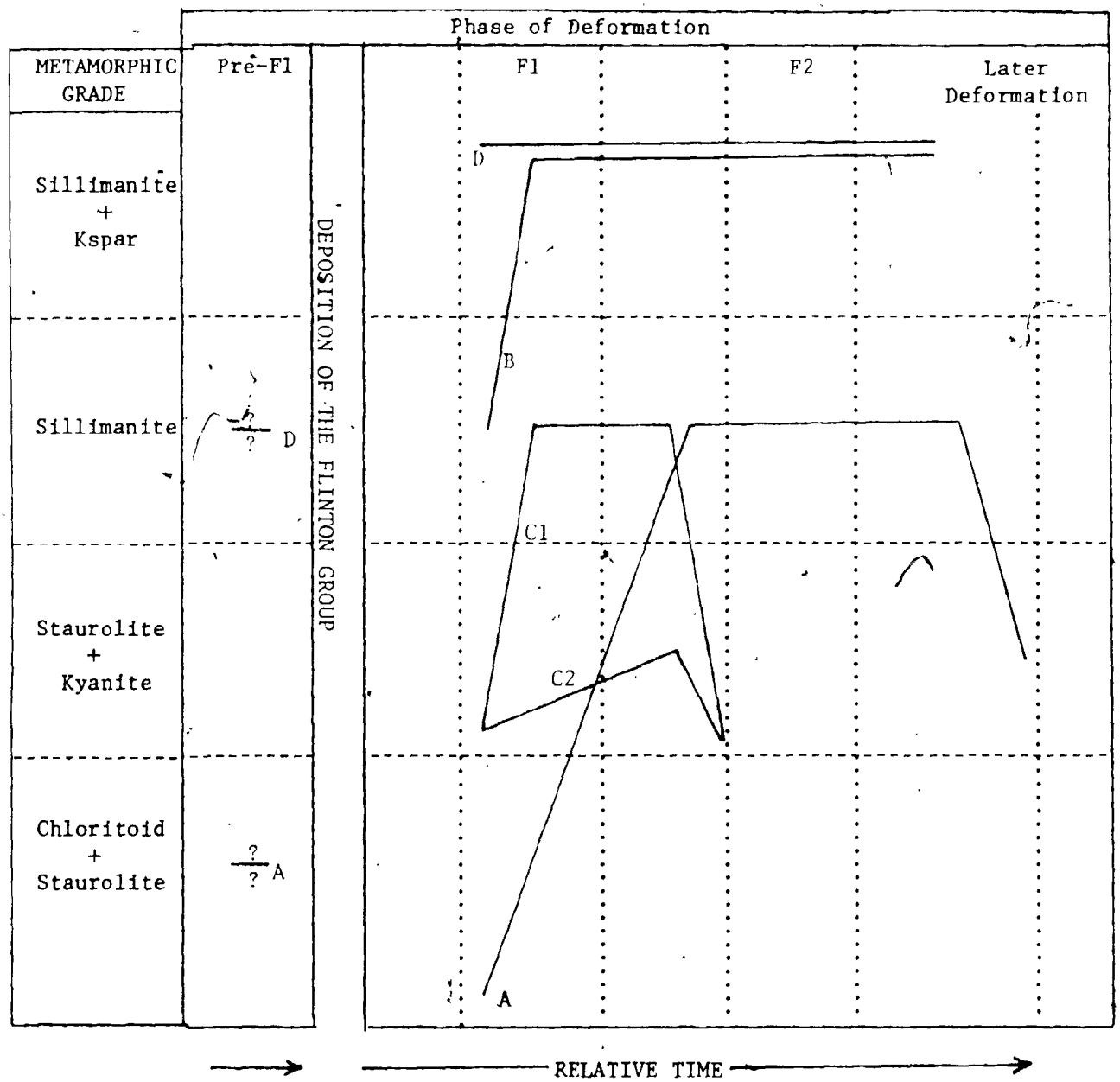


Figure 17. The relative timing between metamorphism and the deformational events recognized in the Kaladar-Dalhousie Trough. The curves represent the results of: Thompson (1972)-A; Rivers (1976)-B; Chappell (1978)-C; and this study-D. Chappell (1978) recognizes a difference in the relative timing of the metamorphic peak and the F1 event in the sillimanite (C1) and staurolite-Kyanite index zone (C2) of the Clare River Structure. The peak metamorphic grade attained during Pre-F1 is not known with certainty. It has been suggested that the Pre-F1 deformational event recognized in the Norcan Lake area may correlate with the pre-Flinton Group deformation recognized by Thompson (1972). Referenced study areas are located in Figure 16, and the distribution of index zones are shown in Figure 14.

conclude that new metamorphic mineral growth was minimal. Chappell concludes that temperatures were relatively low, as metamorphic minerals exhibit undulose extinction around F_2 hinges. Further to the north however, Thompson and Rivers both cite evidence that metamorphic grade remained high during F_2 folding as pegmatite, derived by partial melting, locally cross-cuts F_2 folds.

The regional distribution of the metamorphic index mineral zone boundaries (Fig. 14) has been cited as evidence that the metamorphic culmination was reached during or after F_2 folding, as they do not appear to be folded by these structures (Thompson, 1972). In the Norcan Lake area, it has been suggested that the culmination had been reached during F_1 folding and, in agreement with Rivers and Thompson, remained high during F_2 folding though metamorphic mineral growth was minimal.

It has been suggested that the Pre-F_1 event in the Norcan Lake area may correlate with the pre-Flinton Group deformation discussed by Thompson (1972) and Chappell (1978) (VI.2.5.). This conclusion implies that the Norcan Lake area has undergone polymetamorphism (Fig. 17). Though evidence other than structural/stratigraphic arguments is lacking in support of this implication, it has been concluded that high grade metamorphism during F_1 folding produced a gneissic layering (S_1) that overprints an early mineral foliation (Pre-S_1) (III.6.2.; Plate 3a,c). This type of metamorphic mineral growth exemplified by gneissic layering is not unexpected, as the penetrative Pre-S_1 mineral foliation has been transposed parallel to the axial surface direction of F_1 folds.

VII. SUMMARY OF CONCLUSIONS

Two major sets of folds have been recognized in the Norcan Lake area. The relative age, geometry and orientation of each fold set has been determined on the basis of minor structures and detailed mapping of the major structures in the area. Minor structures indicate that a yet earlier deformational event is also represented in the area, but no major structures associated with this event have been recognized.

Minor isoclinal folds have been noted that fold only compositional layering, probably of primary origin, and which have an axial planar foliation defined by the preferred orientation of inequant minerals (Pre-S_1). These structures have been designated as Pre-F_1 elements. F_1 minor folds are long-limbed isoclines that fold a planar mineral foliation. These folds are "similar" in profile and have an associated penetrative axial planar foliation (S_1) best described as a gneissic layering. Minor structures designated as F_2 consist only of open, and more commonly, tight to isoclinal folds which exhibit "parallel" geometry. On the mesoscopic scale, F_1 and F_2 fold sets are coaxial and have produced Ramsay's Type 3 interference pattern. Hinge lines and lineations (predominantly mineral lineations) consistently plunge gently to the northeast and southwest. F_2 axial planes are steeply inclined to upright, whereas F_1 axial surfaces are more variable in orientation in the area, and can be found recumbent to upright.

Macroscopic examples of both F_1 and F_2 folds have been recognized, though the latter structures dominate the map pattern.

Reversal in plunge of the coaxial hinge lines produces a basin-and-dome interference pattern characteristic of the Norcan Lake area. F_1 and F_2 folding relationships are well established at the crestal culmination of the Mt. McCreary F_2 antiform (structural subarea 3, Fig. 4). Here, mesoscopic F_1 isoclines and associated axial surfaces are prevalent and are recumbent to gently inclined; the orientation of these structures may not have been significantly altered during F_2 folding. The major F_1 axial surface traces are clearly folded by the upright to steeply overturned Mt. McCreary Antiform (Fig. 5). Isoclinal major F_1 structures fold Pre-S_1 foliation and transposed compositional layering and thus the pervasive foliation in the area represents, for the most part, a composite foliation surface (S_{1-1}). Mapping foliation around major fold closures then, is not singularly a criterion for distinguishing F_1 from F_2 folds. Apart from the F_1 structures in the Mt. McCreary Antiform, all the major folds of the Norcan Lake area can be designated as F_2 structures. At the fold closures, the composite foliation of the area traces the major structure and no secondary foliation is developed. F_2 minor folds are present and have symmetries consistent with the major structures. The orientation of all these F_2 folds is upright to steeply overturned.

An additional set of structures, F_3, has been recognized. These are open, gently inclined, "parallel" folds coaxial with F_1 and F_2 structures. No fabrics were found associated with these folds, which are not a prominent structural element in the area. Macroscopic examples of these folds however, may be responsible for locally changing the direction of overturning of F_2 folds.

The structural succession proposed for the Norcan Lake area, from bottom to top, consists of the following sequence of lithologic map units:

1. Heterogeneous gneiss of the Mt. McCreary and Closs gneiss belts (Fig. 7); unit 1
2. Amphibolitic gneiss; unit 2
3. Calcitic marble; unit 3
4. Heterogeneous gneiss of the Green Lake Gneiss Belt (Fig. 7); unit 1
5. Para-amphibolite; unit 4

The upper occurrence of heterogeneous gneiss in the succession is interpreted as a separate stratigraphic horizon, but the possibility of structural repetition can not be excluded. In the Closs Lake Mountain Dome (Fig. 7) tonalite orthogneiss underlies the heterogeneous gneiss and therefore the base of the succession is unknown. Tonalite orthogneiss also exists higher in the proposed succession in the Madawaska River area and is in contact with units 3, 4 and the Green Lake Gneiss Belt. A syenite orthogneiss (unit 8) may have been emplaced as a sill-like body between the Green Lake Gneiss Belt and para-amphibolite (unit 4).

The succession above the lower heterogeneous gneiss (3 through 5 above) is relatively thin, approximately 750 metres, across the north-plunging F_2 antiform of the Closs Lake Mountain Dome (structural subarea 5; Fig. 4). The style of tight to isoclinal refolding (Type 3 of Ramsay, 1967) and subhorizontal, doubly-plunging hinge lines, has produced elongate and narrow outcrop belts for relatively

thin stratigraphic units. This may be a common structural feature in areas adjacent to the Norcan Lake area.

Examined in detail, the heterogeneous gneiss belts, which are composed to a large extent of pink granite and grey quartz diorite gneiss, are not uniform masses of rock which can readily be assigned a plutonic origin. The coarsely interlayered nature of the unit, the local presence of primary compositional layering and gradations with paragneissic rocks, and the lateral persistence of closely associated rocks along regional strike, all suggest that sedimentary and volcanic processes are responsible for the origin of the heterogeneous gneiss.

Most of the Norcan Lake area is believed to have been subjected to high grade metamorphism. In the Closs Lake Mountain Dome, quartzo-feldspathic rocks have undergone partial melting producing extensive migmatitic structures. This migmatite terrain is believed to lie above a suggested quartz-muscovite-sillimanite-K-feldspar isograd. The major deformation in the area is considered to be synmetamorphic: biotite, hornblende, and possibly sillimanite define the Pre-S_1 foliation; K-feldspar-rich segregations define S_1 gneissic layering; and granitic melts cross-cut F_2 folds in the migmatite terrain.

Structural correlations have been suggested with previous studies in the Kaladar-Dalhousie Trough to the South. Synmetamorphic deformation producing early phase recumbent structures (F_1) and coaxial, more upright refolds (F_2), though not prevalent, is a common structural sequence in various parts of the region (Venkitasubramanyan,

1969; Rivers, 1976; Chappell, 1978). The findings from this study support the conclusions of Thompson (1972) and Chappell (1978), who cite evidence for a Pre-F_1 deformational event which, in the Flinton area, predates deposition of the Flinton Group metasediments.

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APPENDIX

I. Distribution of samples used for the modal analysis of unit 1, and of quartz-muscovite-sillimanite-K-feldspar assemblages.

II. Table of mineral assemblages for the Norcan Lake area.

III. Detailed geologic map of the structural succession in the vicinity of Blithfield Long Lake.

Appendix (continued)

List of Abbreviations

	Abbreviation
Quartz	QZ
K-feldspar	KF
Plagioclase	PL
Biotite	BI
Hornblende	HN
Garnet	GN
Clinopyroxene	CP
Calcite	CC
Scapolite	SC
Sphene	SP
Magnetite	MG
Tremolite	TM
Phlogopite	PH
Sillimanite	SI
Muscovite	MU

List of Symbols

- Granite gneiss sample used in modal analysis of unit 1

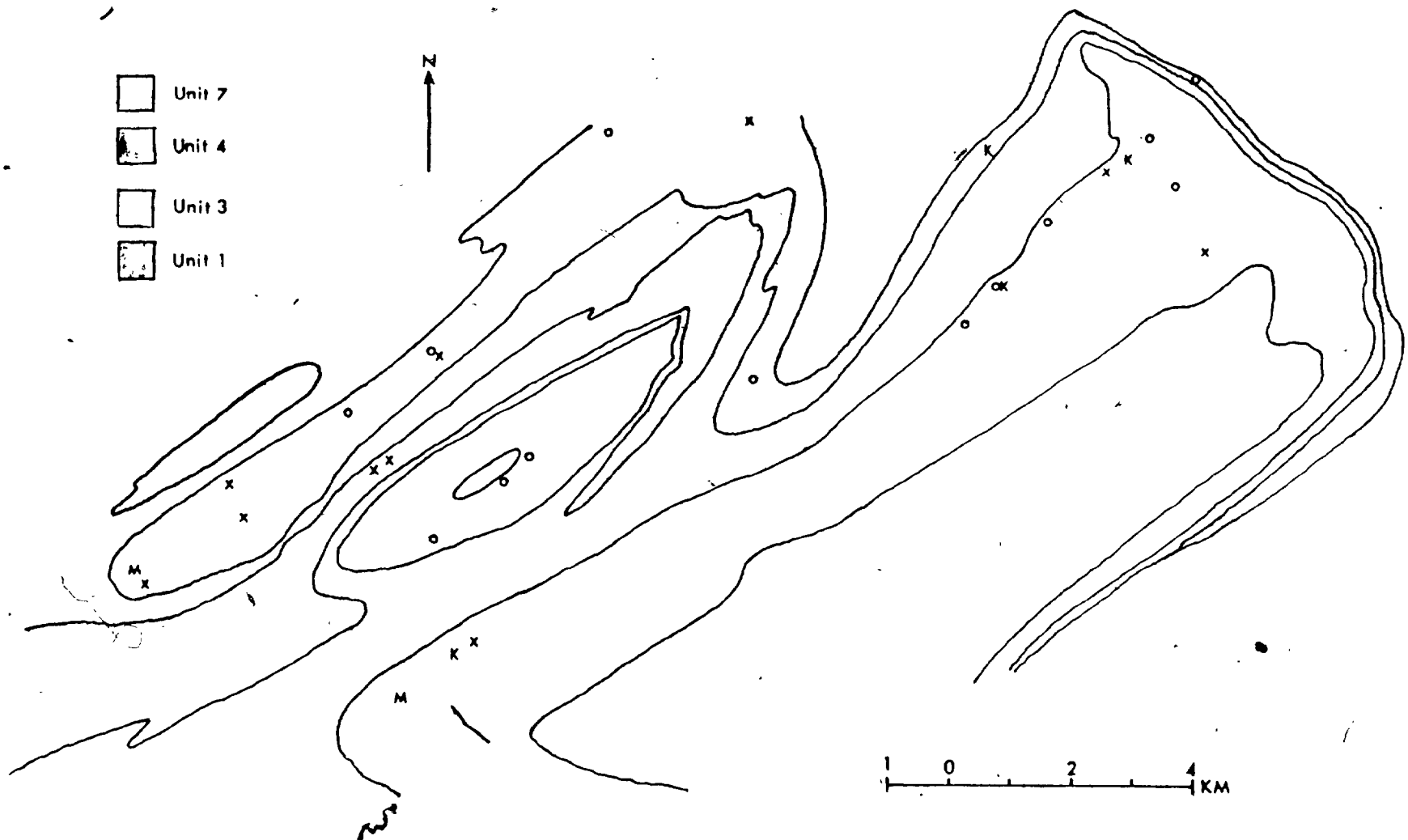
- Quartz diorite gneiss sample used in modal analysis of unit 1

- Muscovite-quartz-sillimanite assemblage

- Quartz-sillimanite-K-feldspar assemblage

x - Major rock constituent, greater than 5%

- - Minor rock constituent, less than 5%



I. Distribution of samples used for the modal analysis of unit 1, and the location of MU-QZ-SI-KF assemblages.

Appendix (continued)

II. Table of Mineral Assemblages for the Norcan Lake Area

UNIT	LITHOLOGY	QZ	KF	PL	BI	HN	GN	CP	CC	SC	SP	MG	TM	PH
1	Granite gneiss	x	x	x	x	x								
1	"	x	x	x	x		-							
1	Quartz diorite gneiss	x	x	x	x	x								
1	"	x		x	x		-							
1	Quartzite	x											x	x
1	"	x	x					x						
1	Amphibolite			x		x		x						
2	Amphibolitic gneiss	x		x	x	x								
2	"	x		x		x		-						
2	"			x		x				-				
3	Calcitic marble							x	x	x				
3	"					x		x	x					
3	"							x	x				-	
3	"							x	x					x
3	impure marble		x						x					x
4	Para-amphibolite			x		x	-	-						
7	Tonalite	x	-	x	x	x								
8	Syenite		-	x	x	x	-							

UNIT	LITHOLOGY	MU	QZ	SI	KF
	Muscovite-Sillimanite-Quartz Zone				
1	Granite gneiss	-	x		
	"	-	x	-	
	Quartz-Sillimanite-K-Feldspar Zone				
1	Granite gneiss		x	-	x
	"		x	-	x
	"		x	-	x
	"		x	-	x

Appendix (continued)

III. Detailed Geologic Map of the Structural Succession in the Vicinity of Blithfield Long Lake

LEGEND

Trend and plunge of F1 folds.....

Long axis of boudin.....

Pre-F1 mullion lineation.....

Composite S0-1 foliation: compositional layering parallel to Pre-S1 foliation.....

Composite S1-1 foliation: Pre-S1 foliation parallel to S1 gneissic layering.....

Assumed S1-1 foliation.....

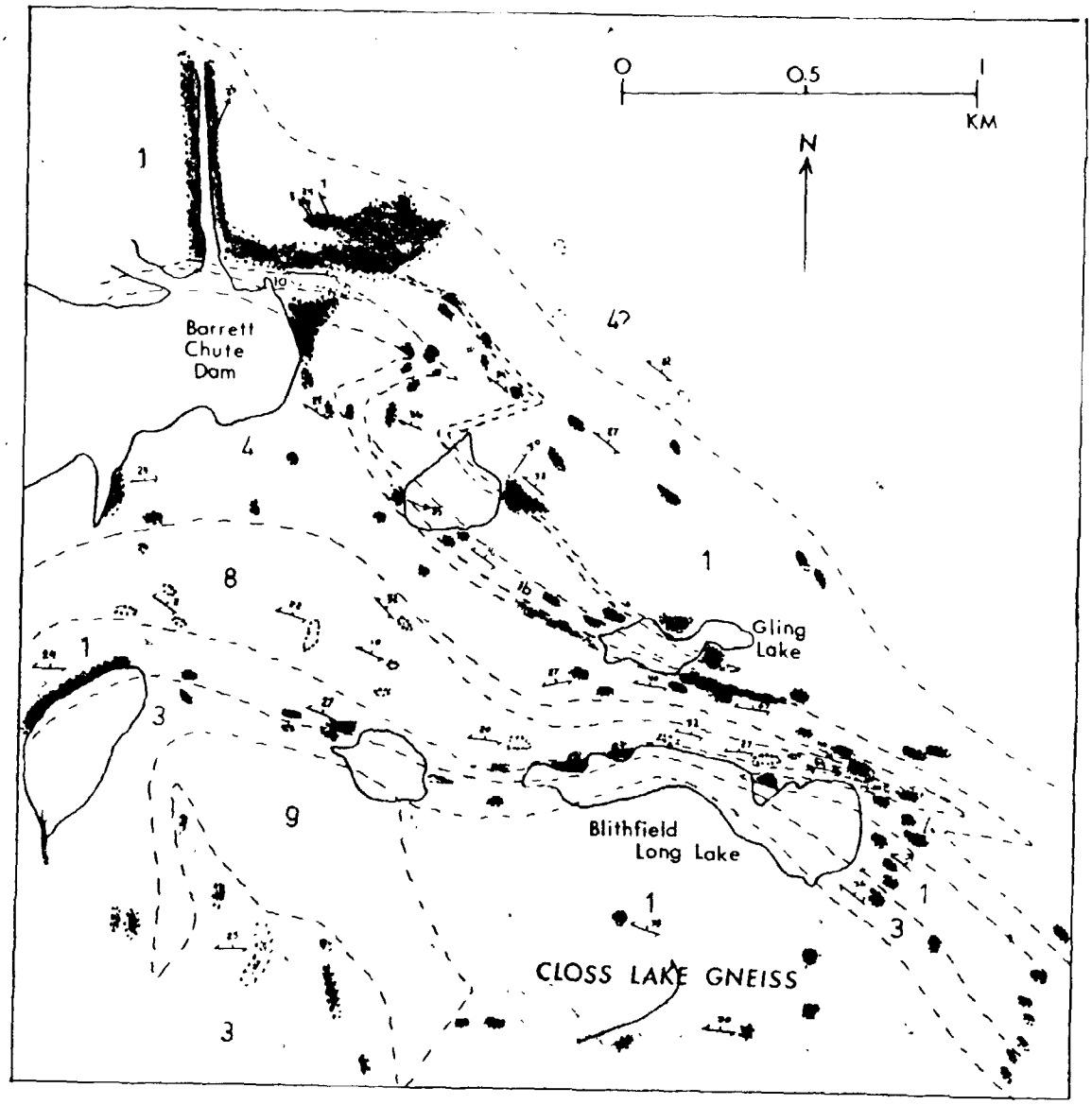
Notes:

Units designated as on Figure 3. Additional units include feldspathic quartzite (1a) and quartz-K-feldspar-biotite schist (1b).

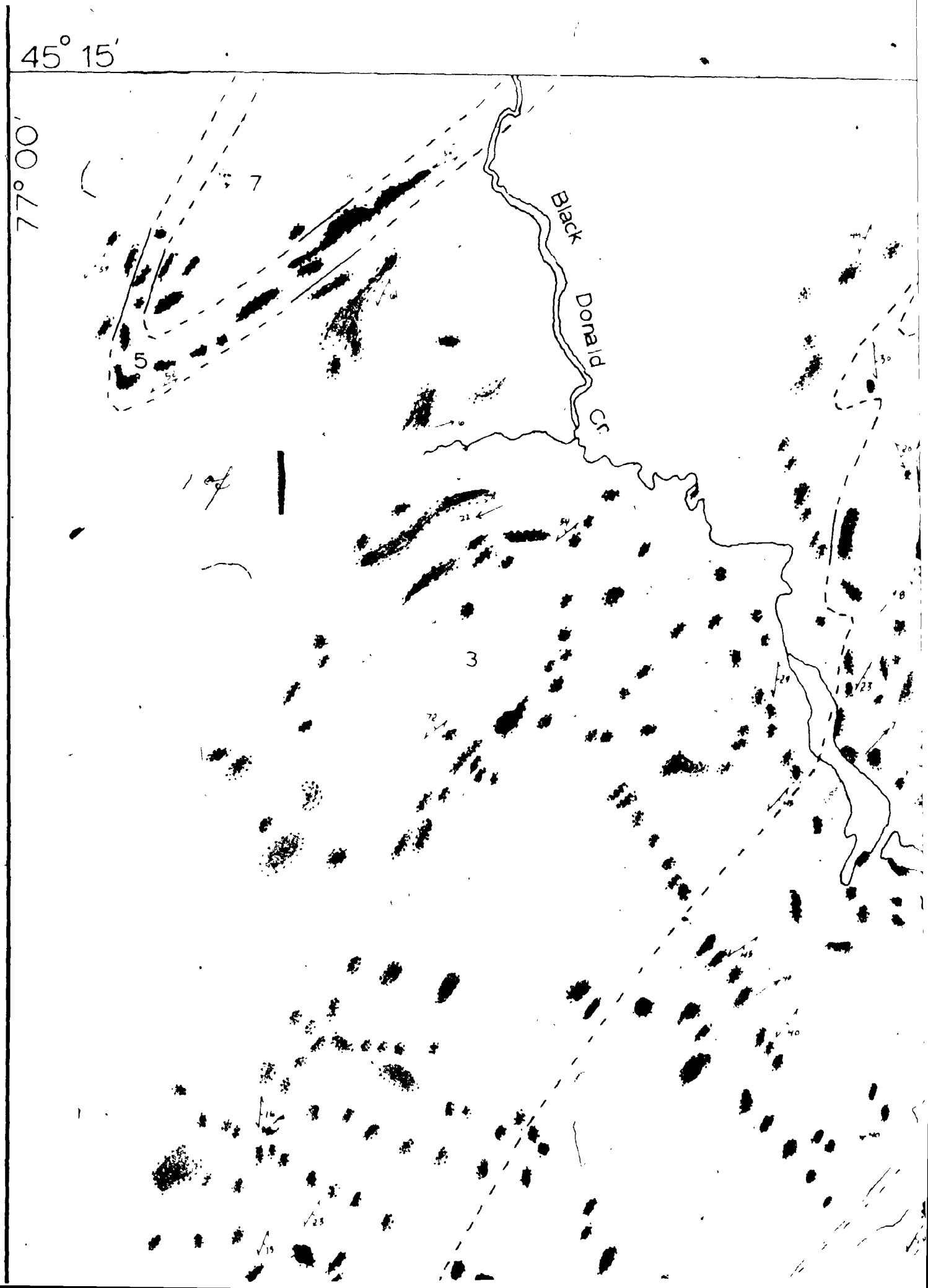
The area was reexamined during the fall of 1979 and contacts consequently modified.

The map covers an area along the northern limb and fold closure of the Closs Lake Mountain Dome (Figs. 4 and 5), interpreted as an F2 antiformal structure. The folds in the vicinity of Barrett Chute and Blithfield Long Lake are interpreted as shallow-plunging F1 folds. The structural succession for the Norcan Lake area begins with the Closs Lake Gneiss (unit 1) overlain by calcitic marble (unit 3) (unit 2, amphibolitic gneiss, was not observed in this immediate vicinity, but may be present), Green Lake Gneiss (unit 1), and para-amphibolite (unit 4). Syenite orthogneiss (unit 8) also exists in the succession. The higher occurrences of heterogeneous gneiss (unit 1 at Barrett Chute) and para-amphibolite may be the result of F1 structural repetition.

Appendix (continued)



III. Detailed Geologic Map of the Structural Succession in the Vicinity of Blithfield Long Lake.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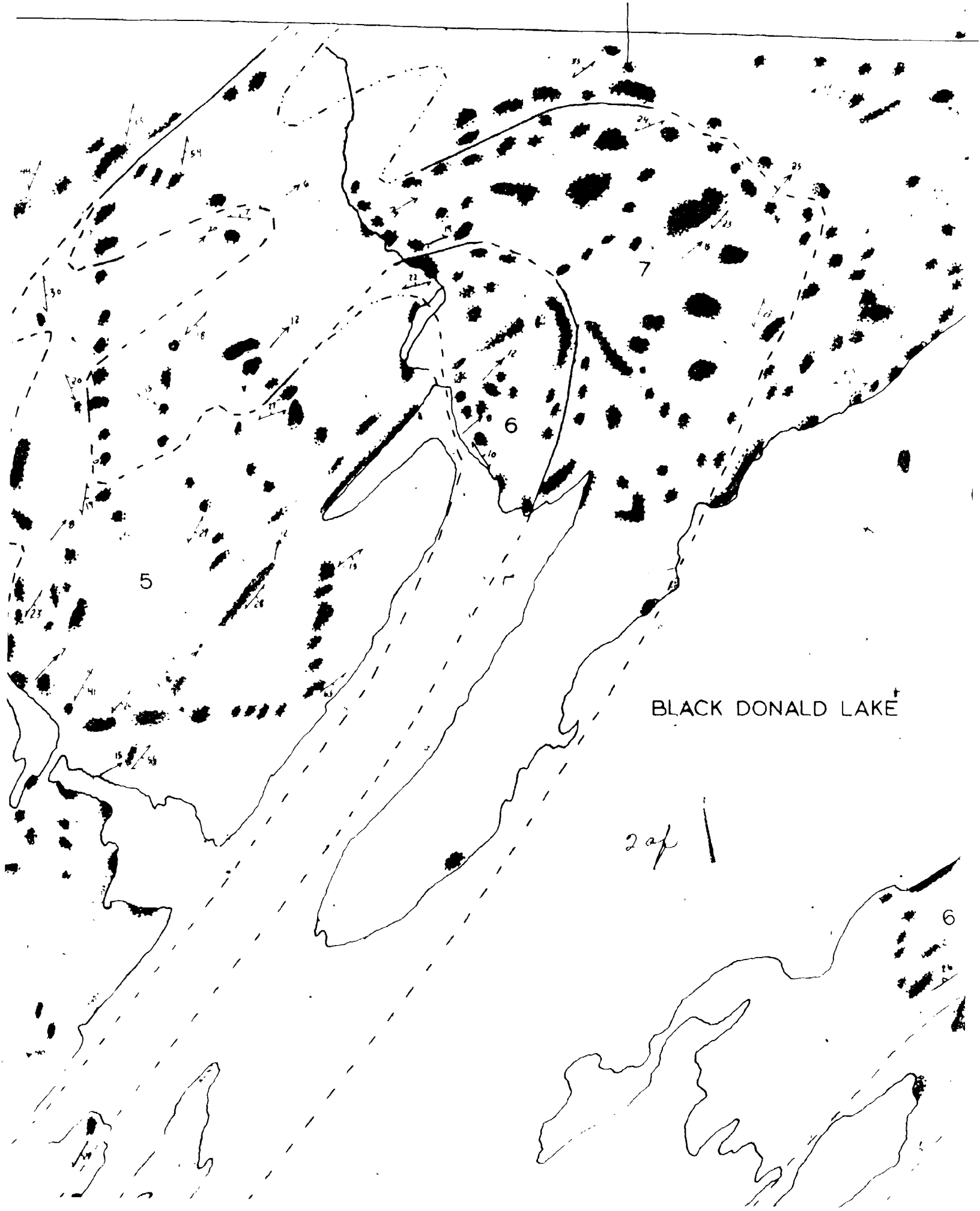
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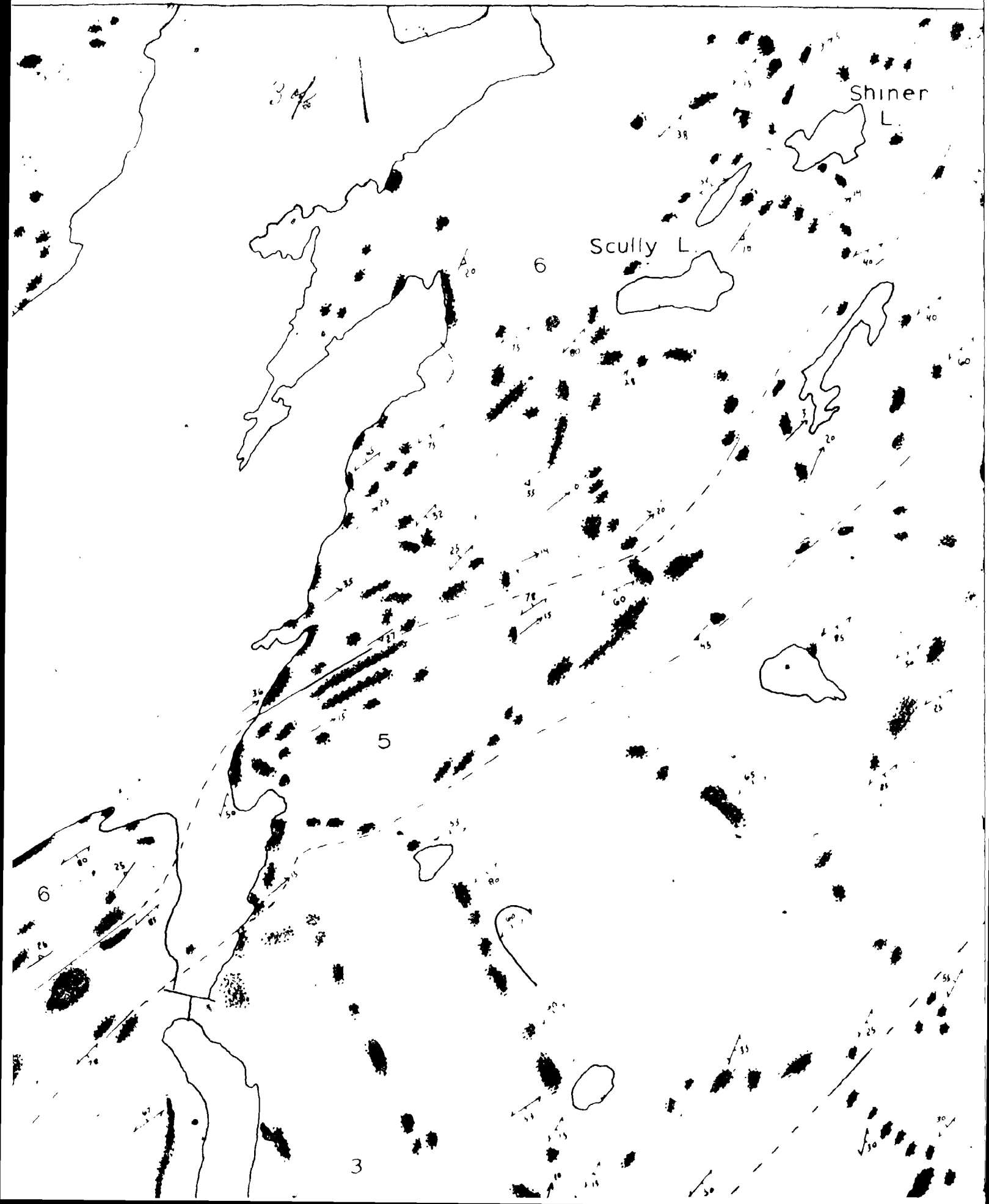
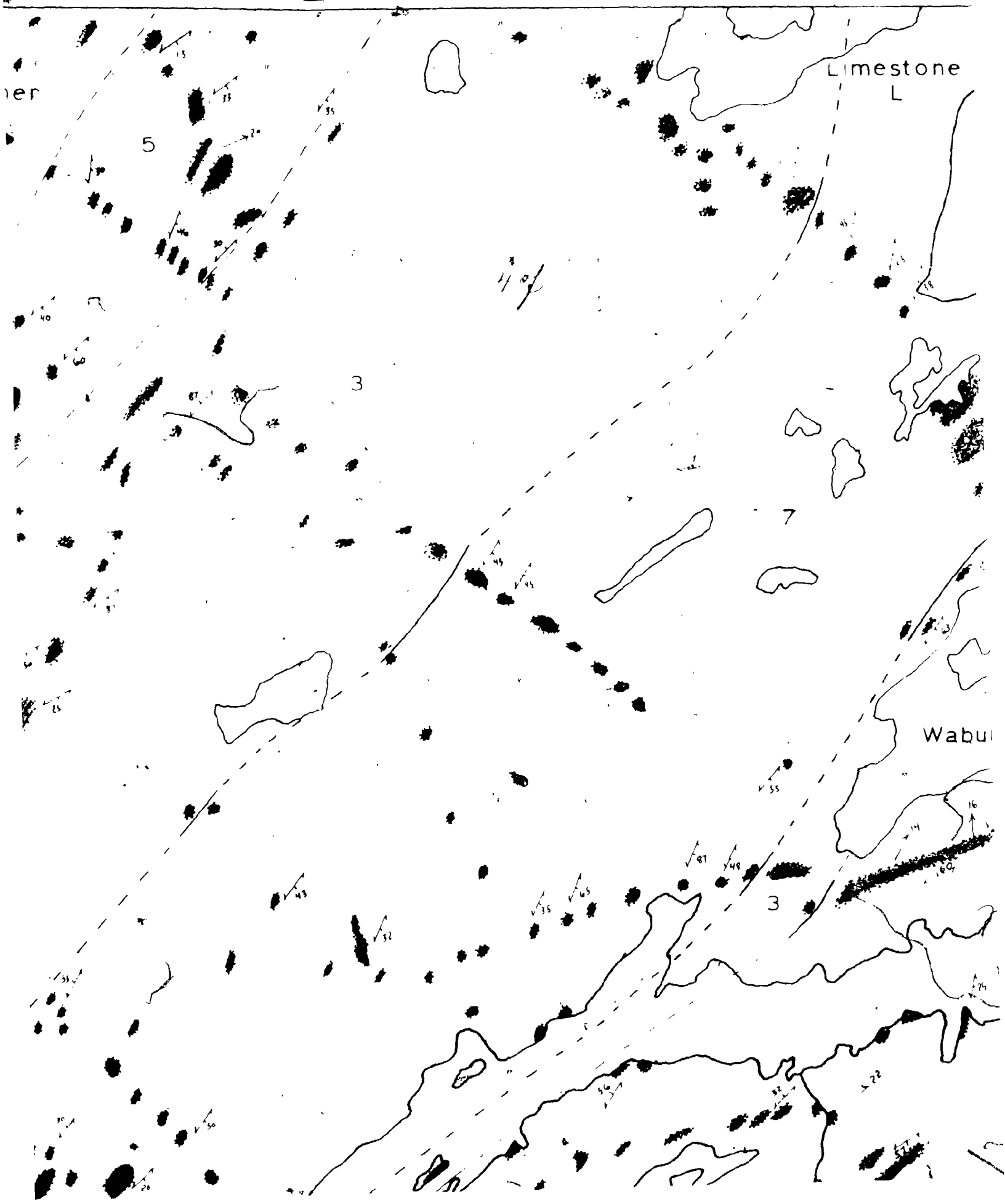
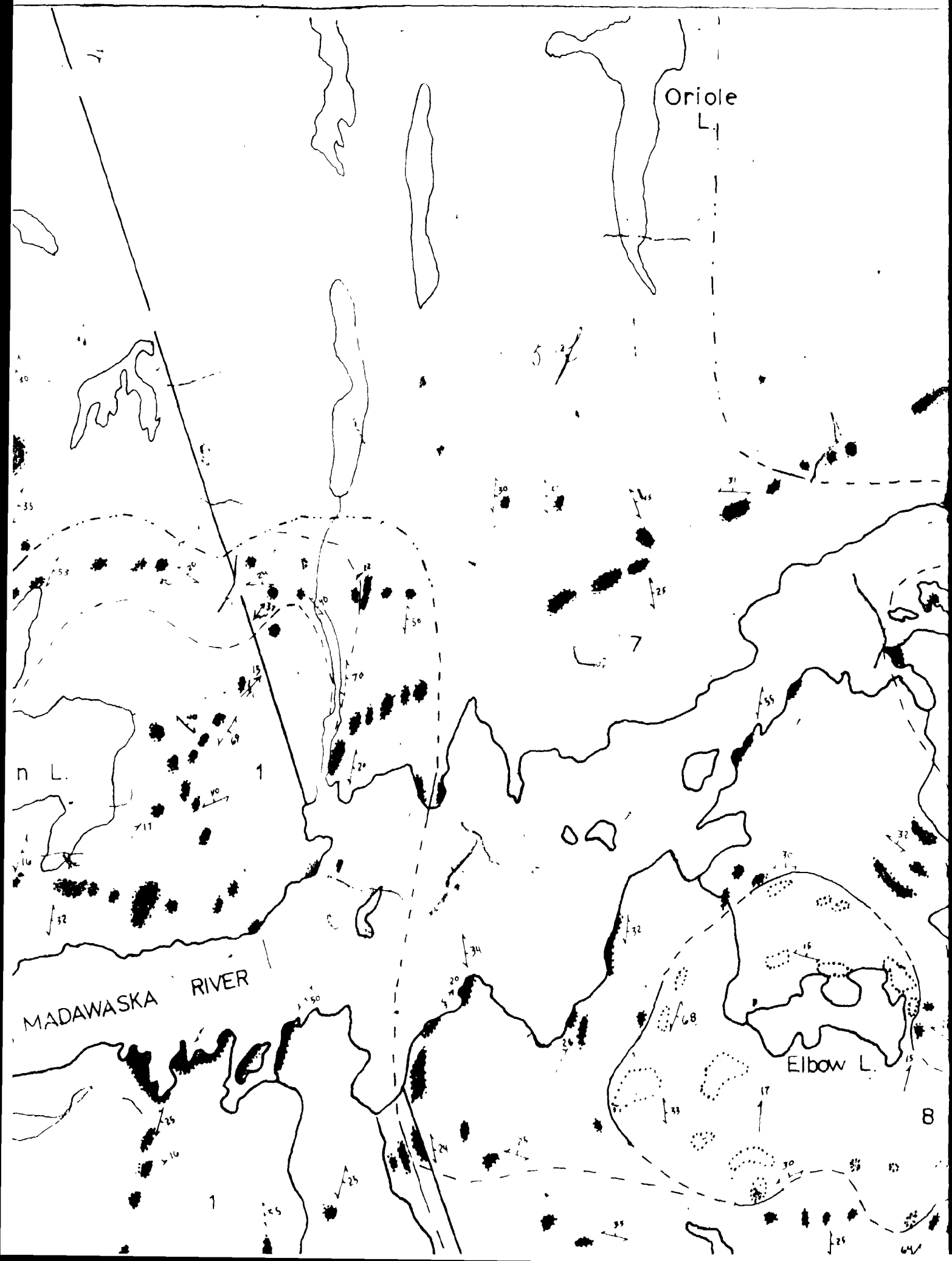
2 of 1
6

GEOLO



LOGY OF THE NORCA

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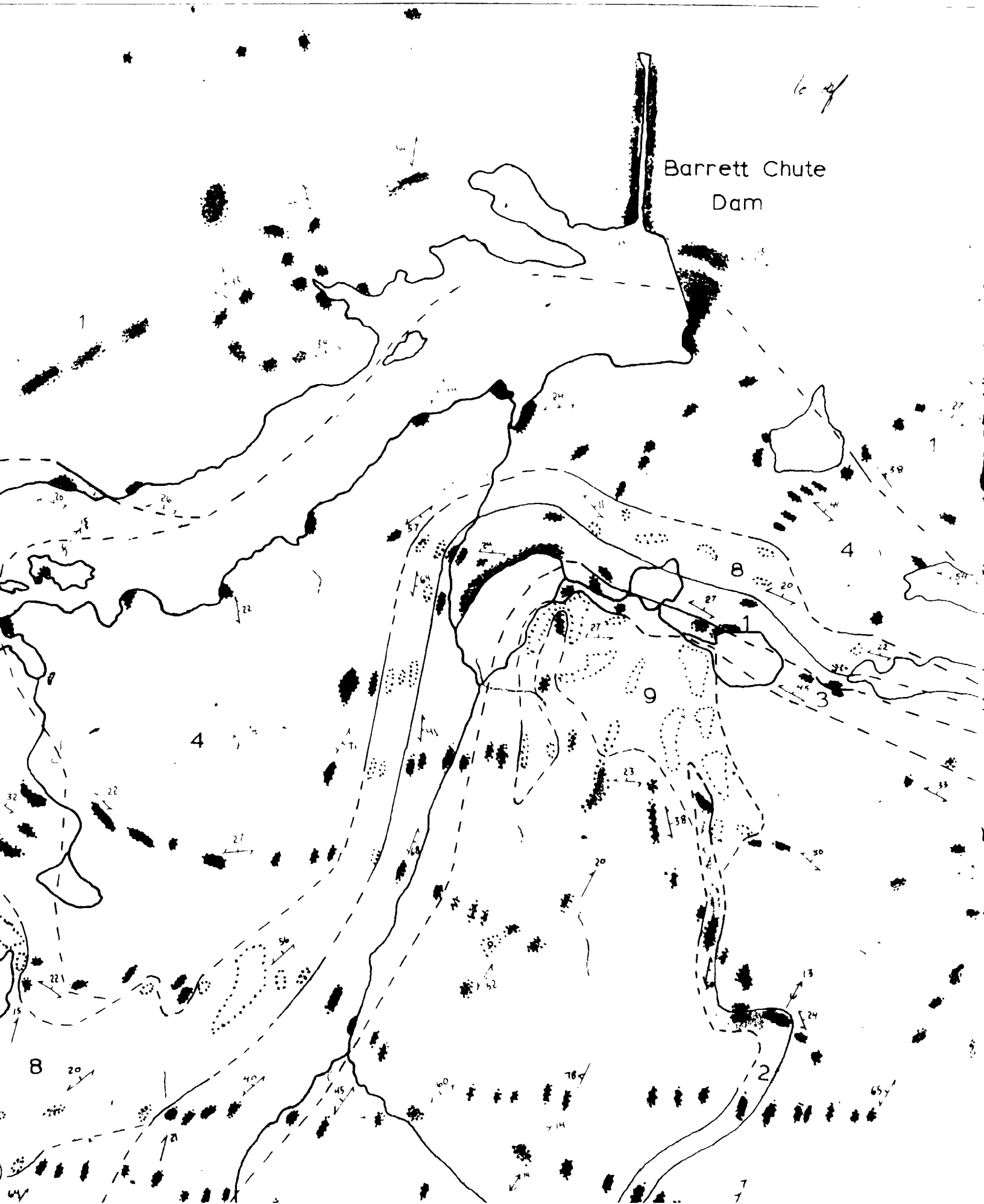
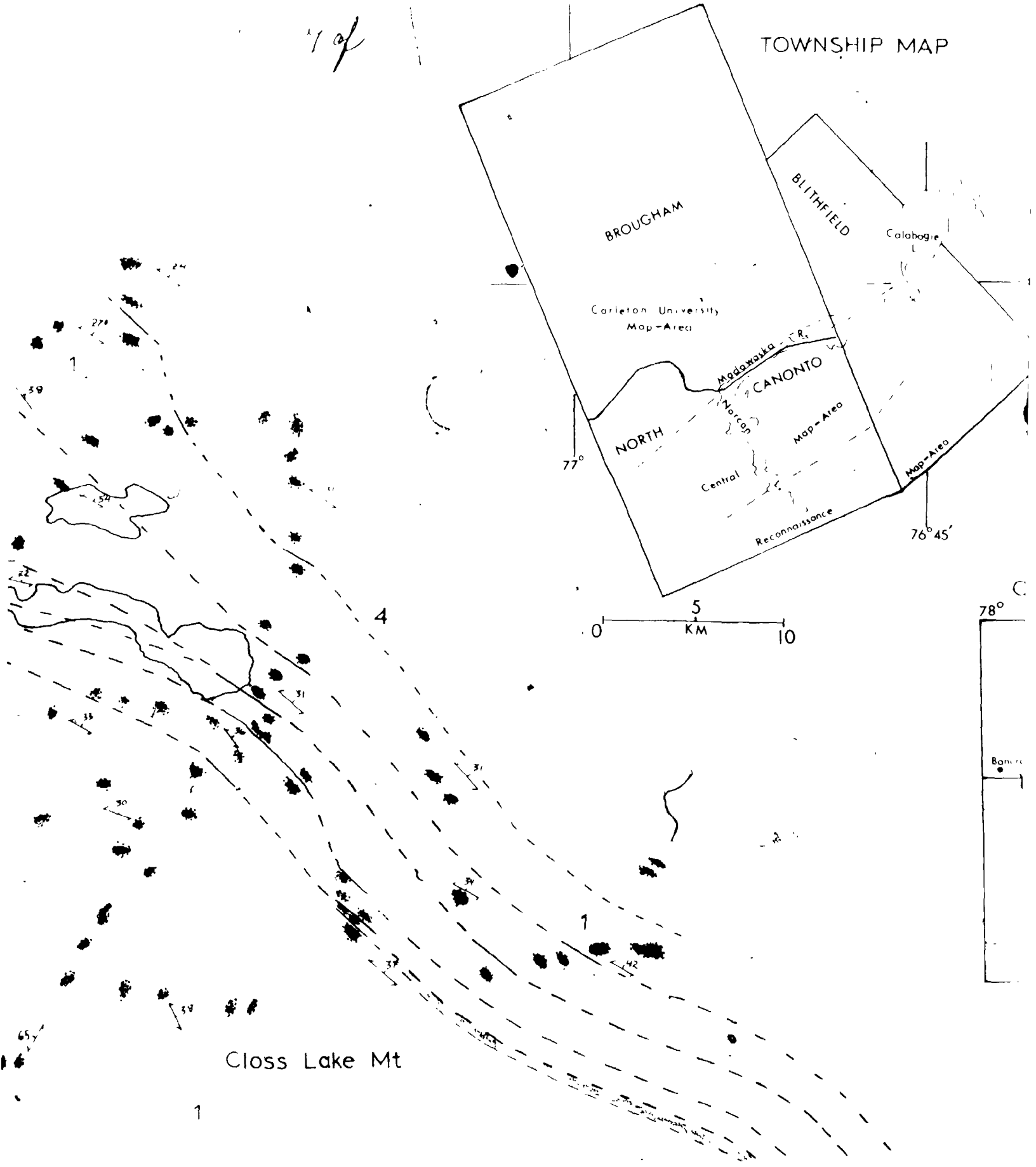
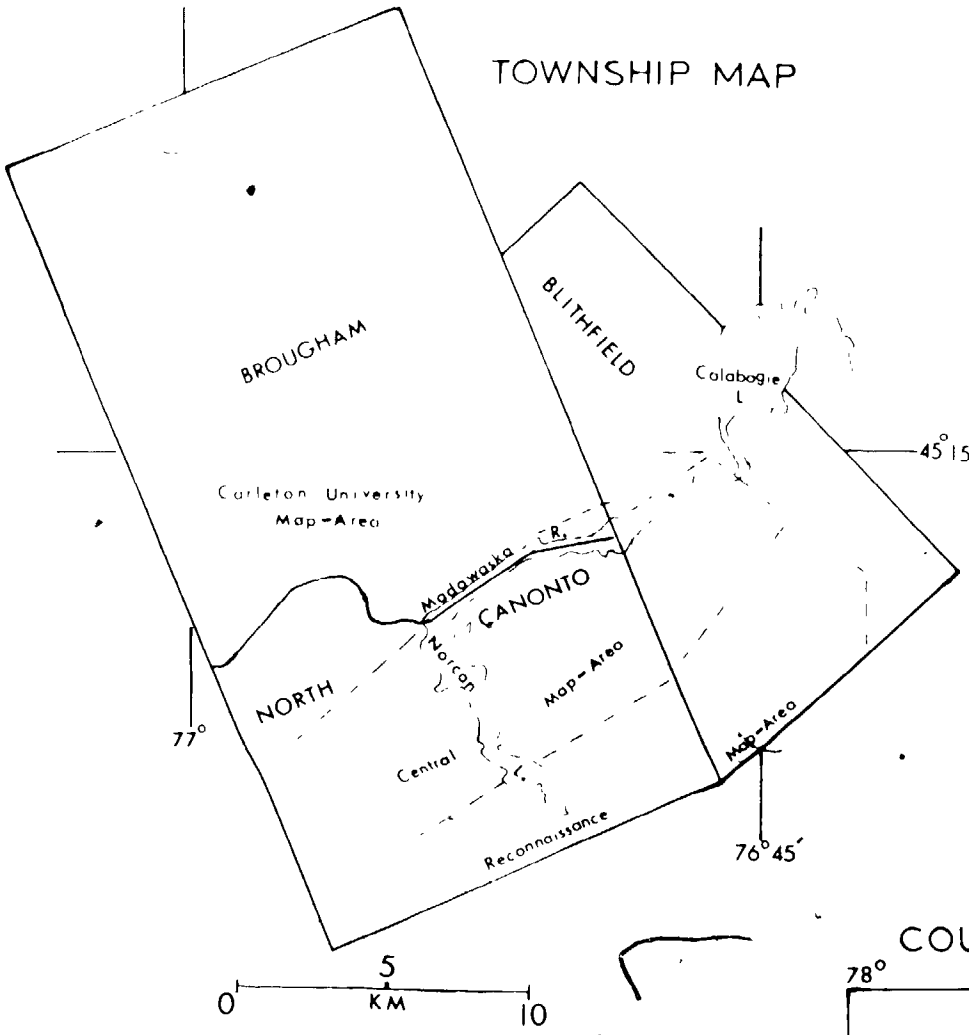

Barrett Chute
Dam

LOCATION MAPS

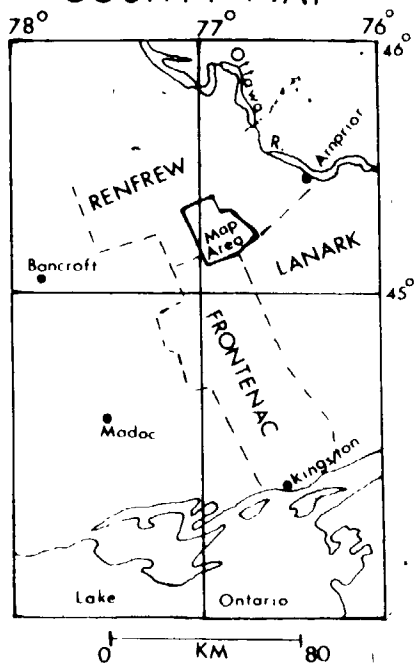


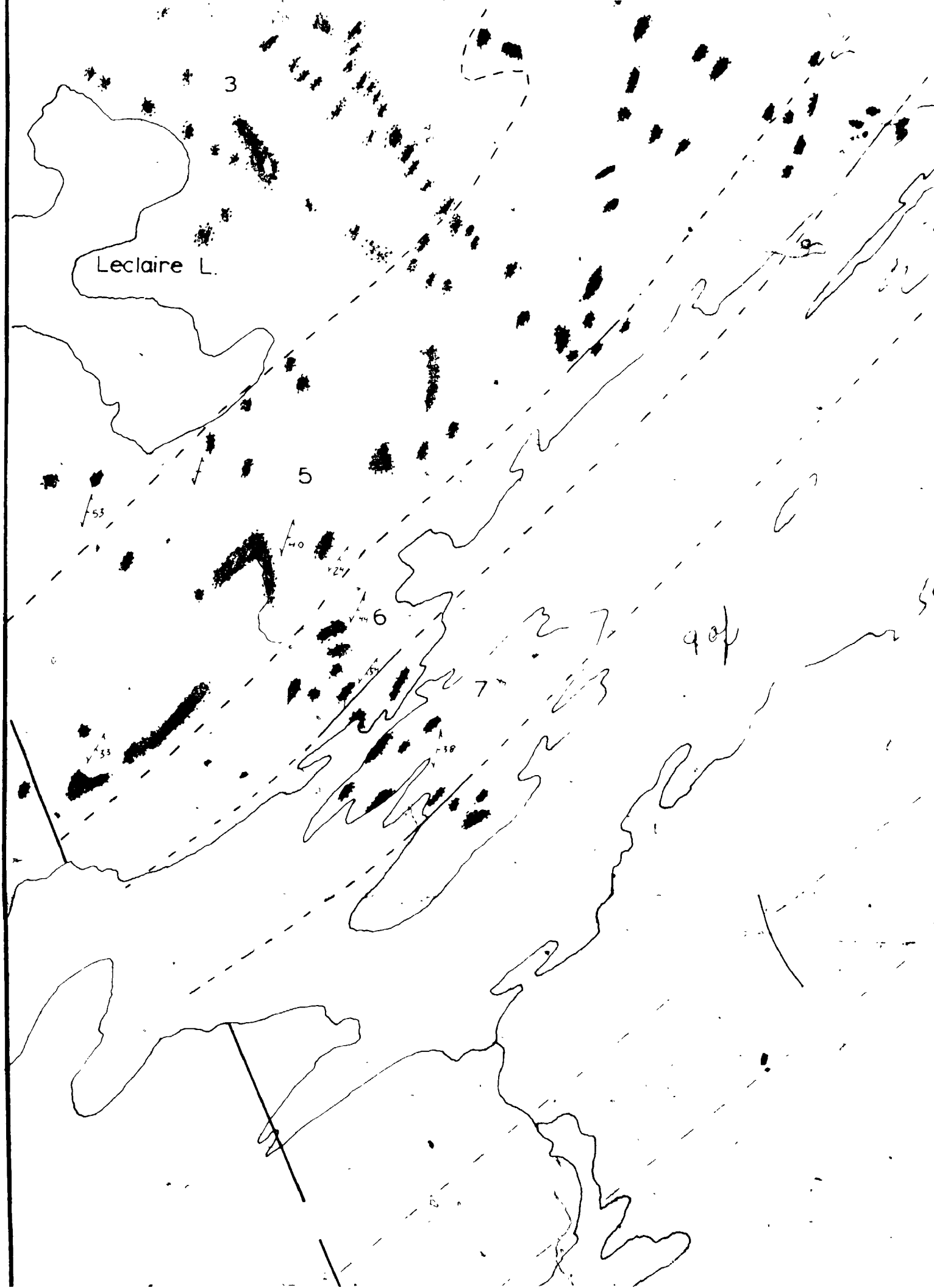
LOCATION MAPS

TOWNSHIP MAP



COUNTY MAP

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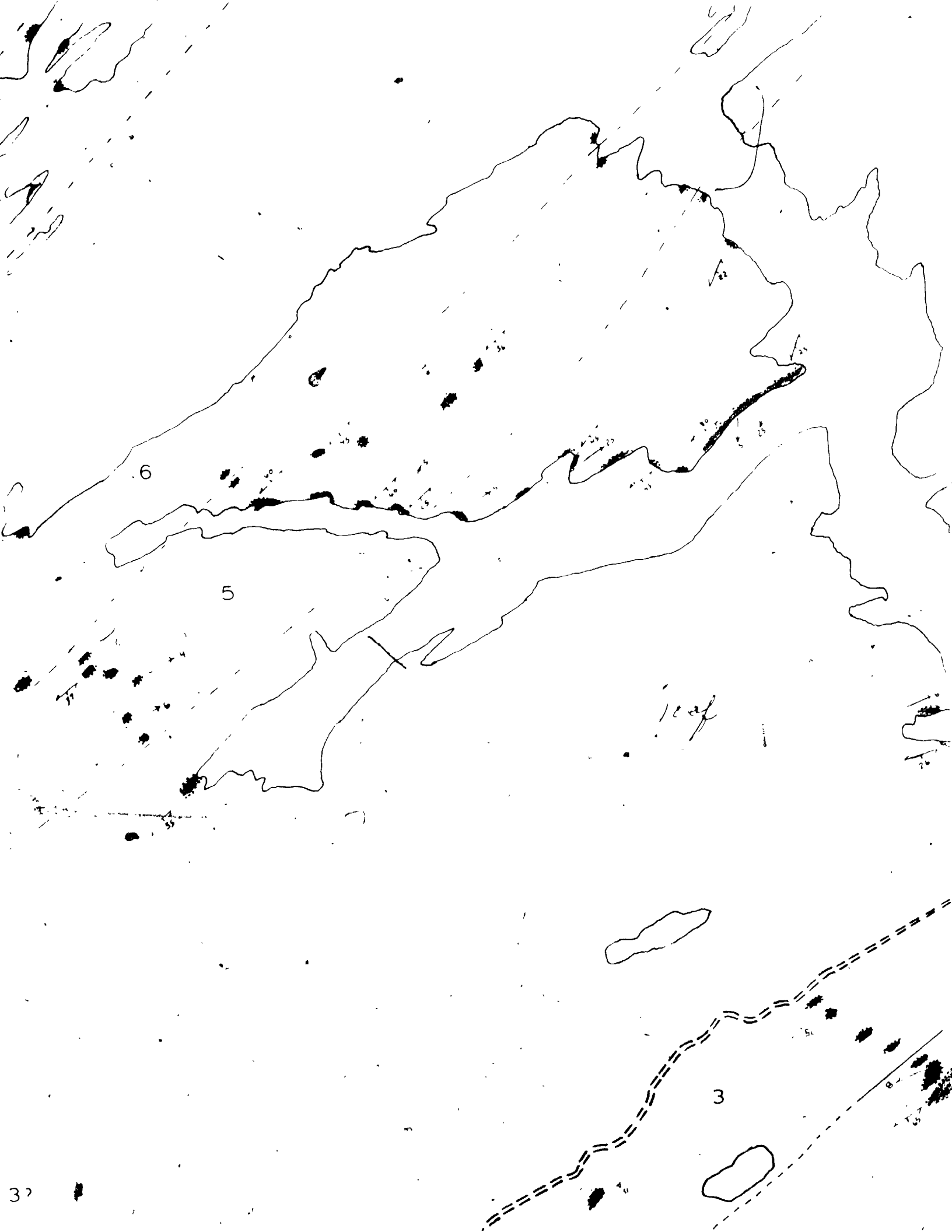
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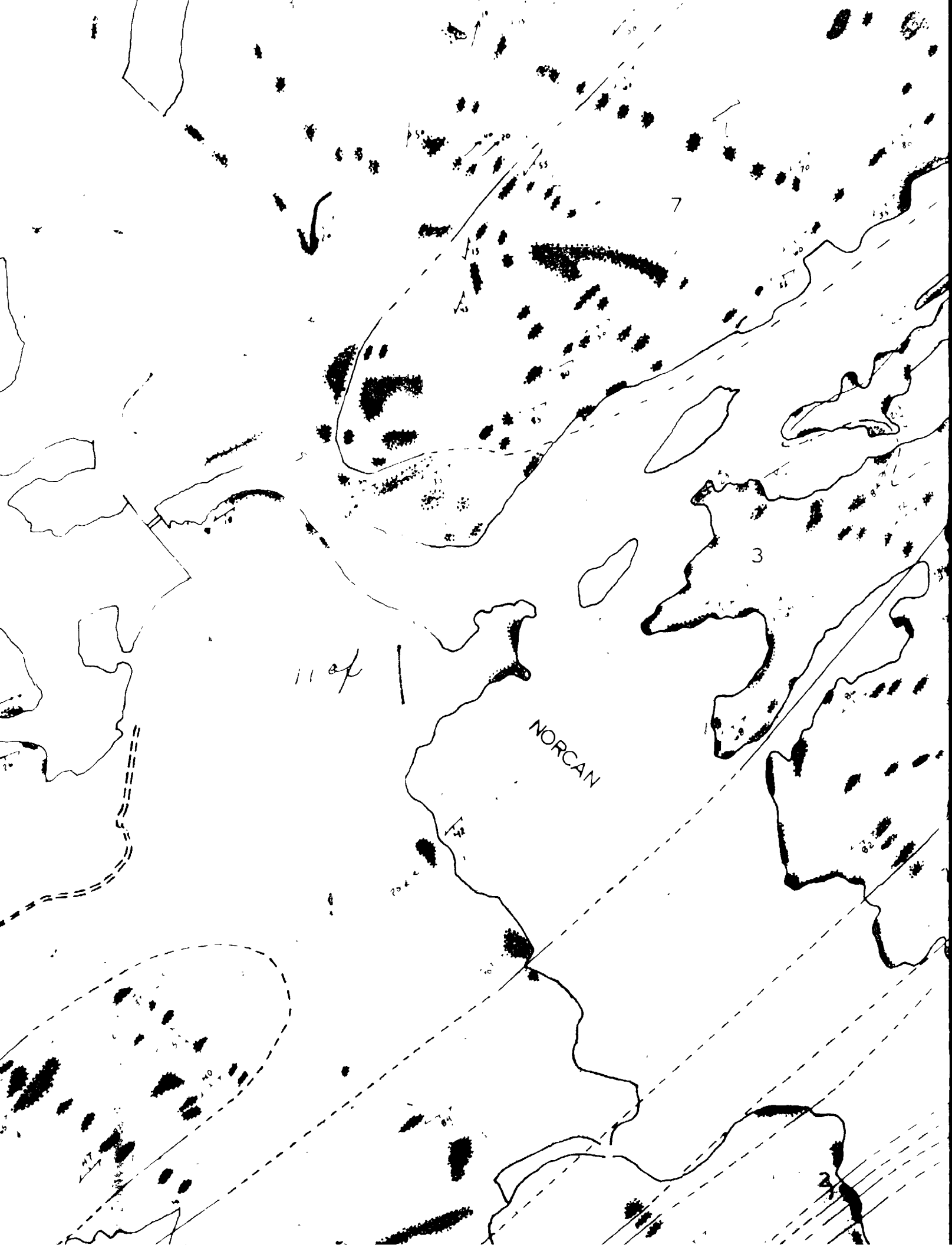
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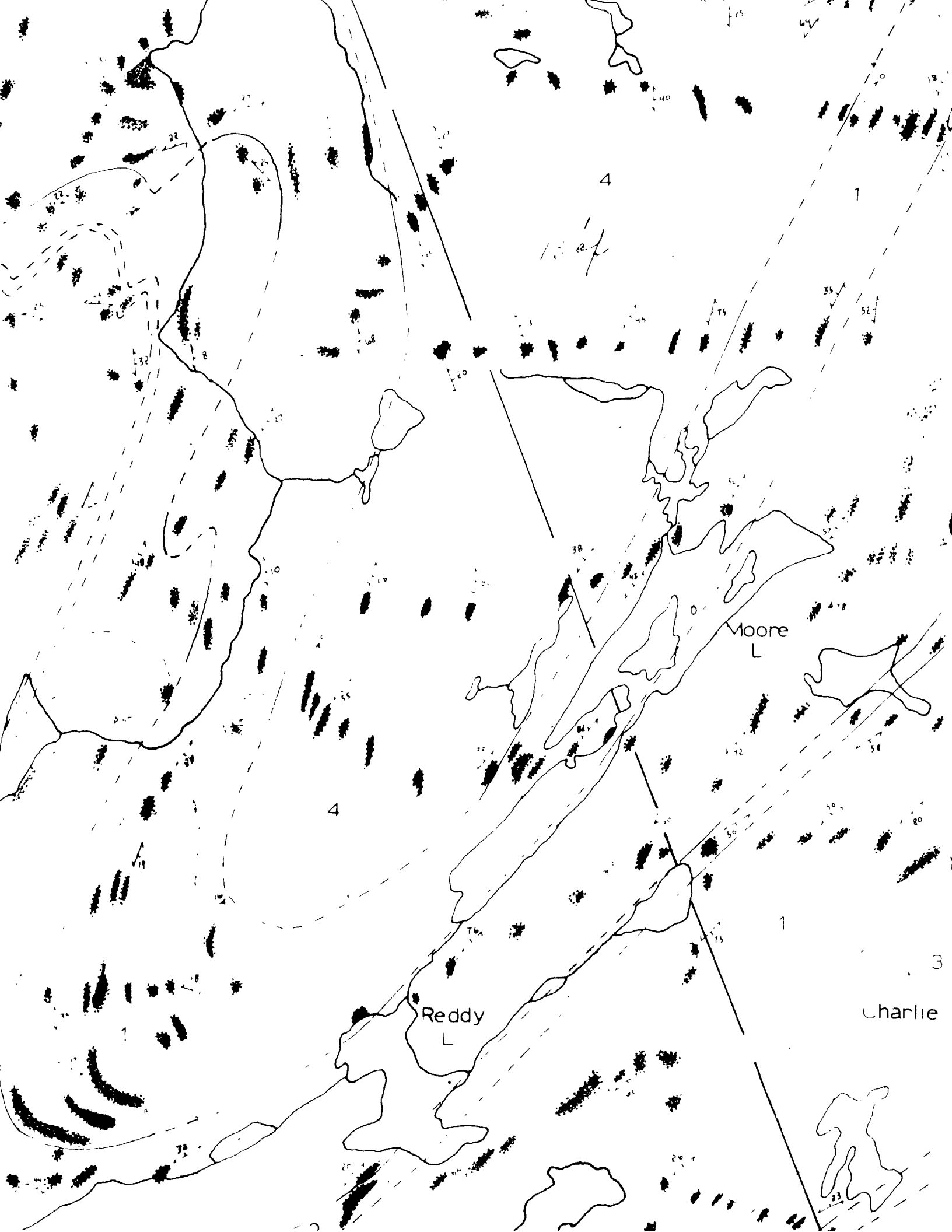
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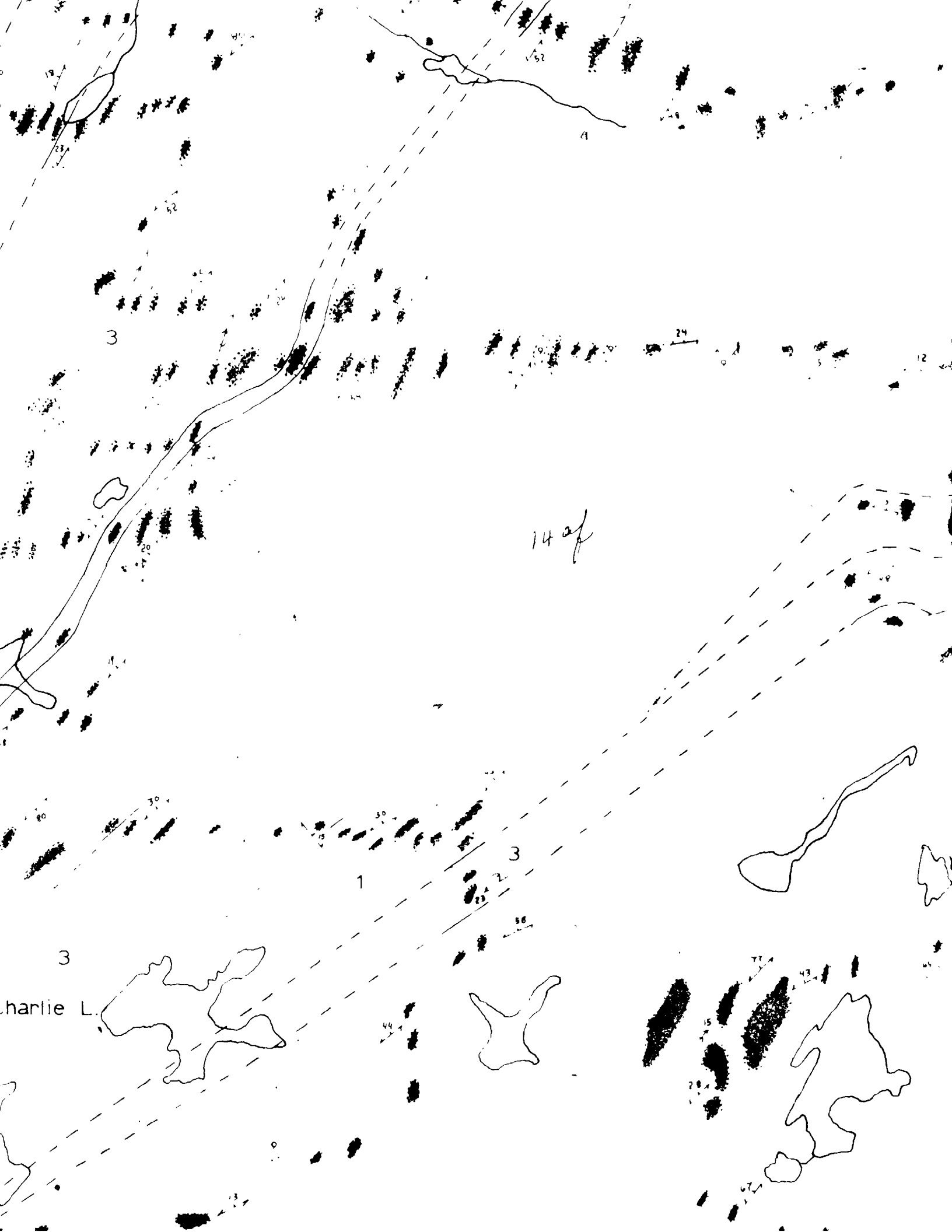
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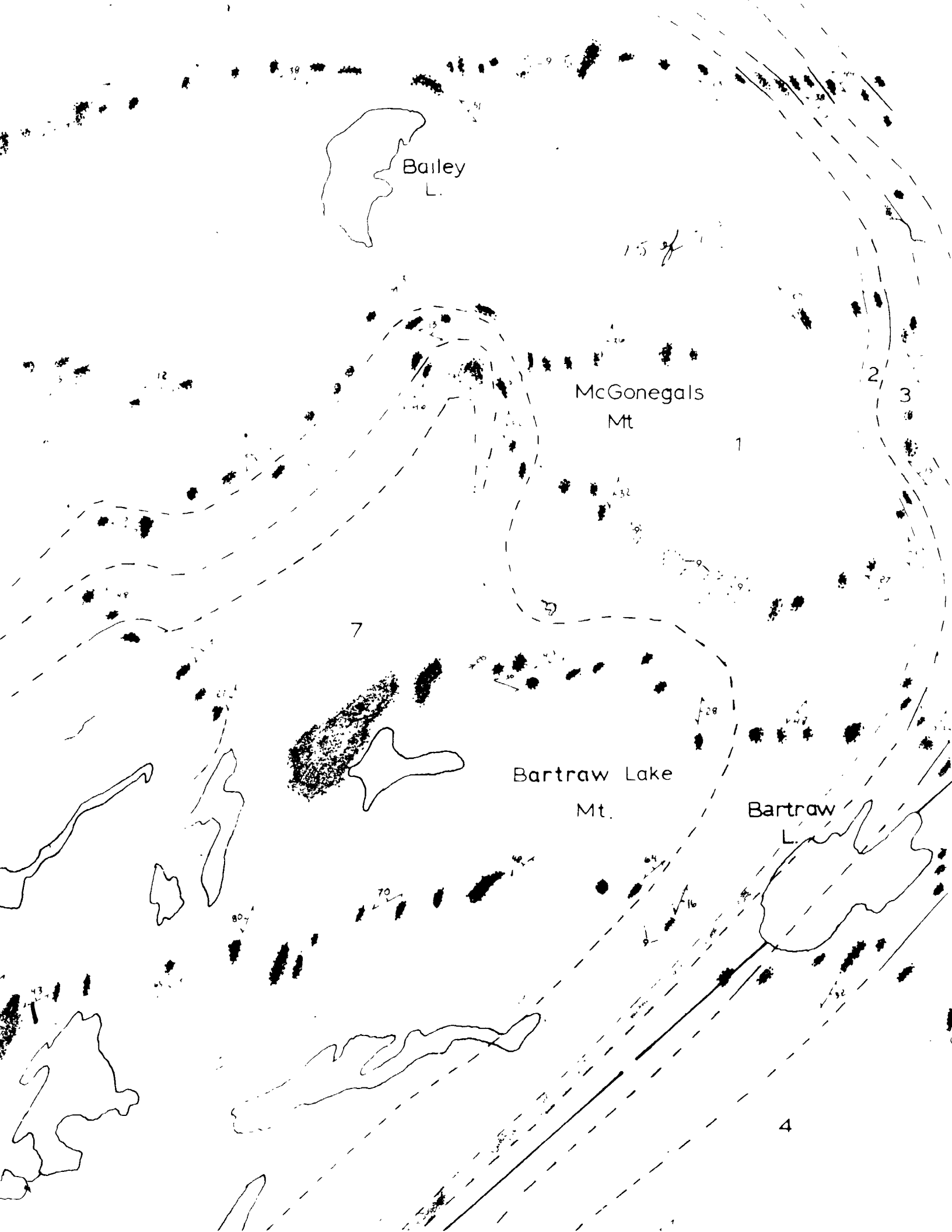
harlie L.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Bartraw Lake
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Bartraw
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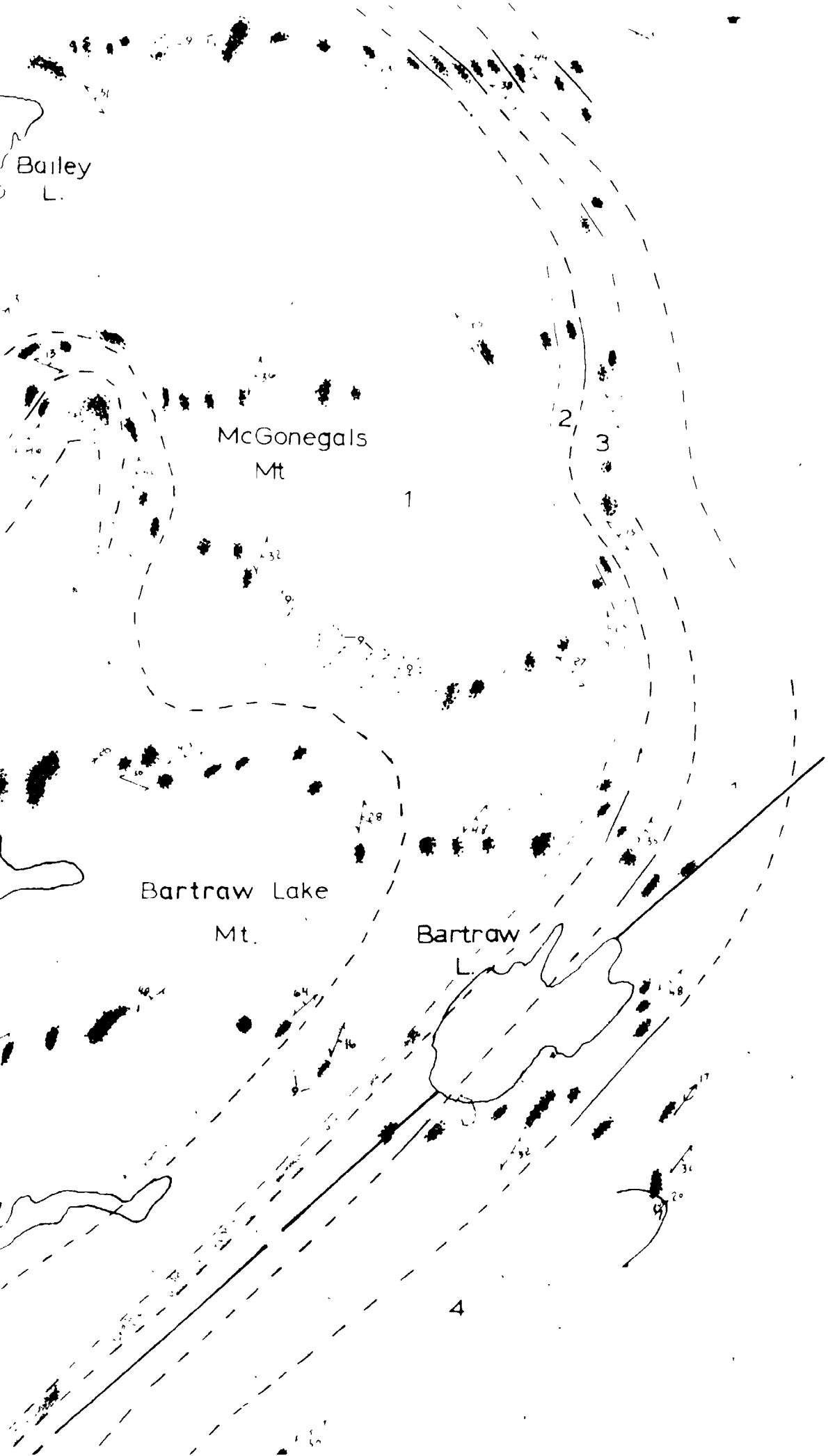
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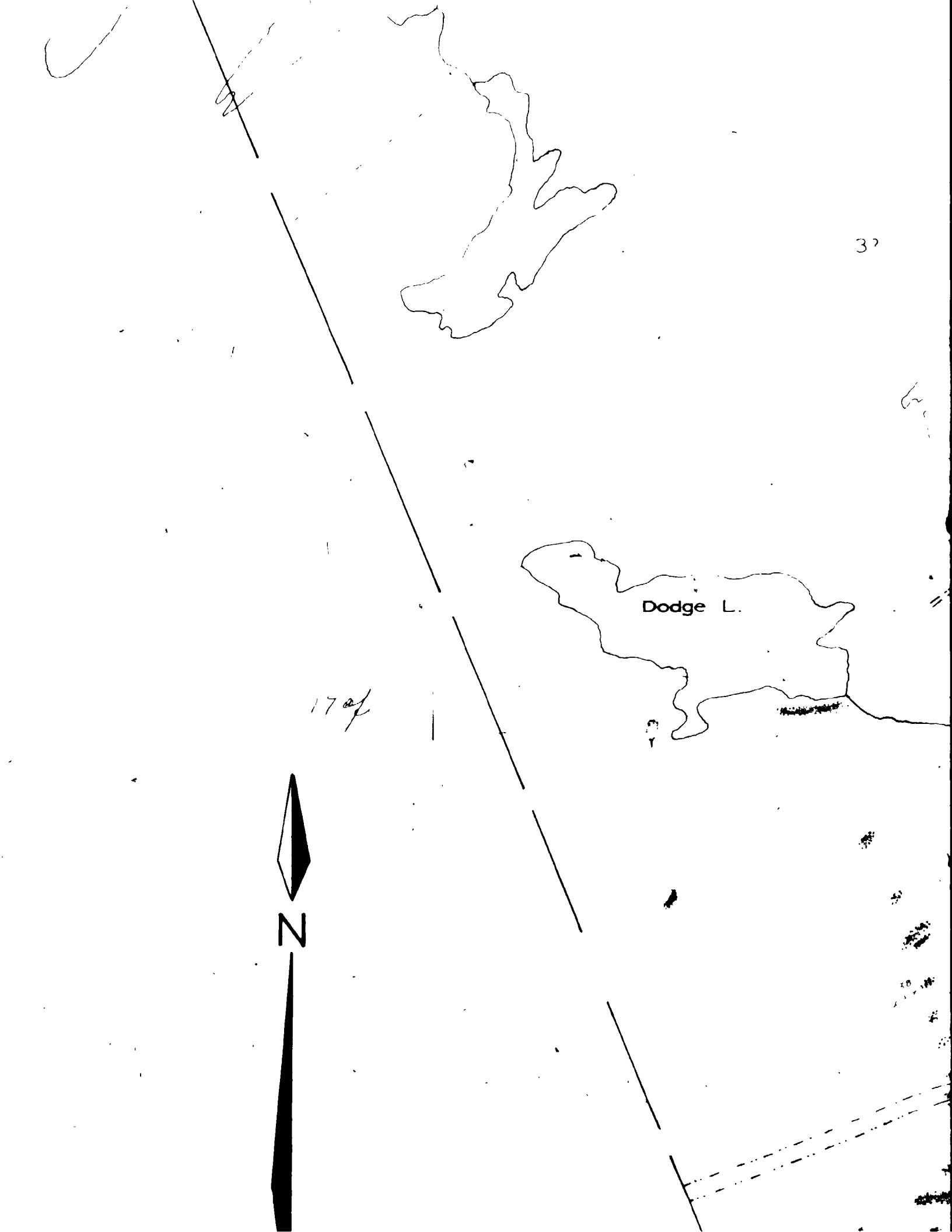
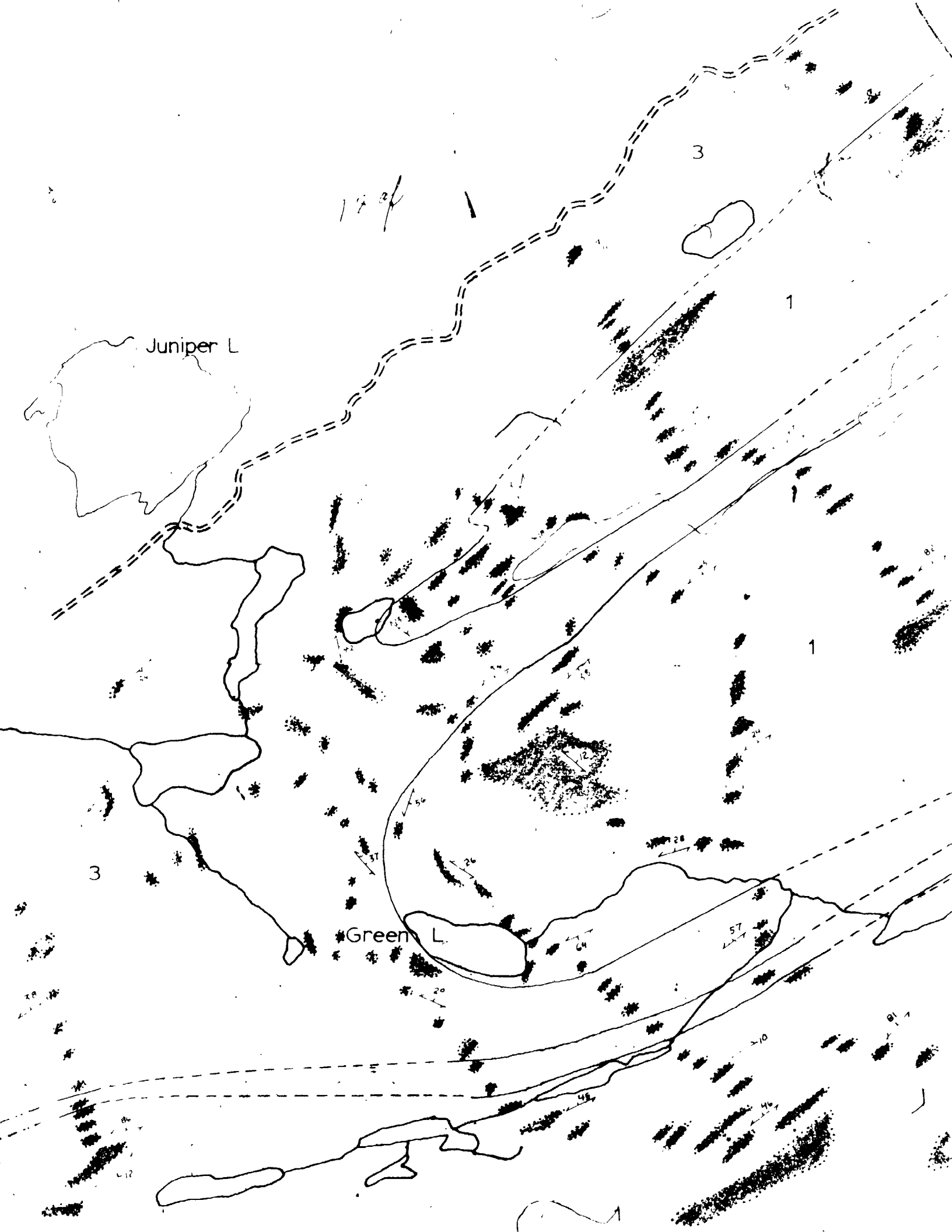
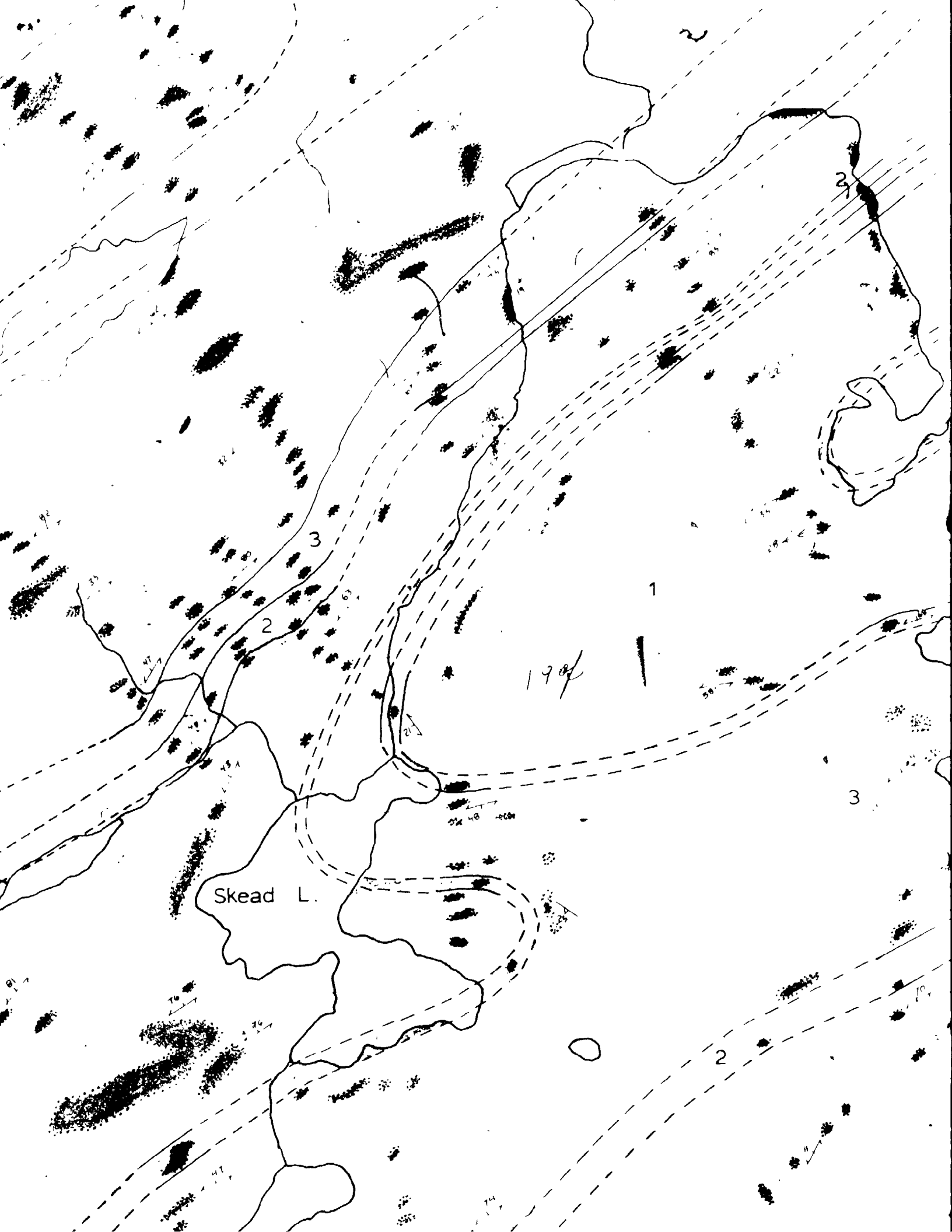

Dodge L.
N


Juniper L
Green L
3
1
1
3


Skead L.
1
2
3
2
3
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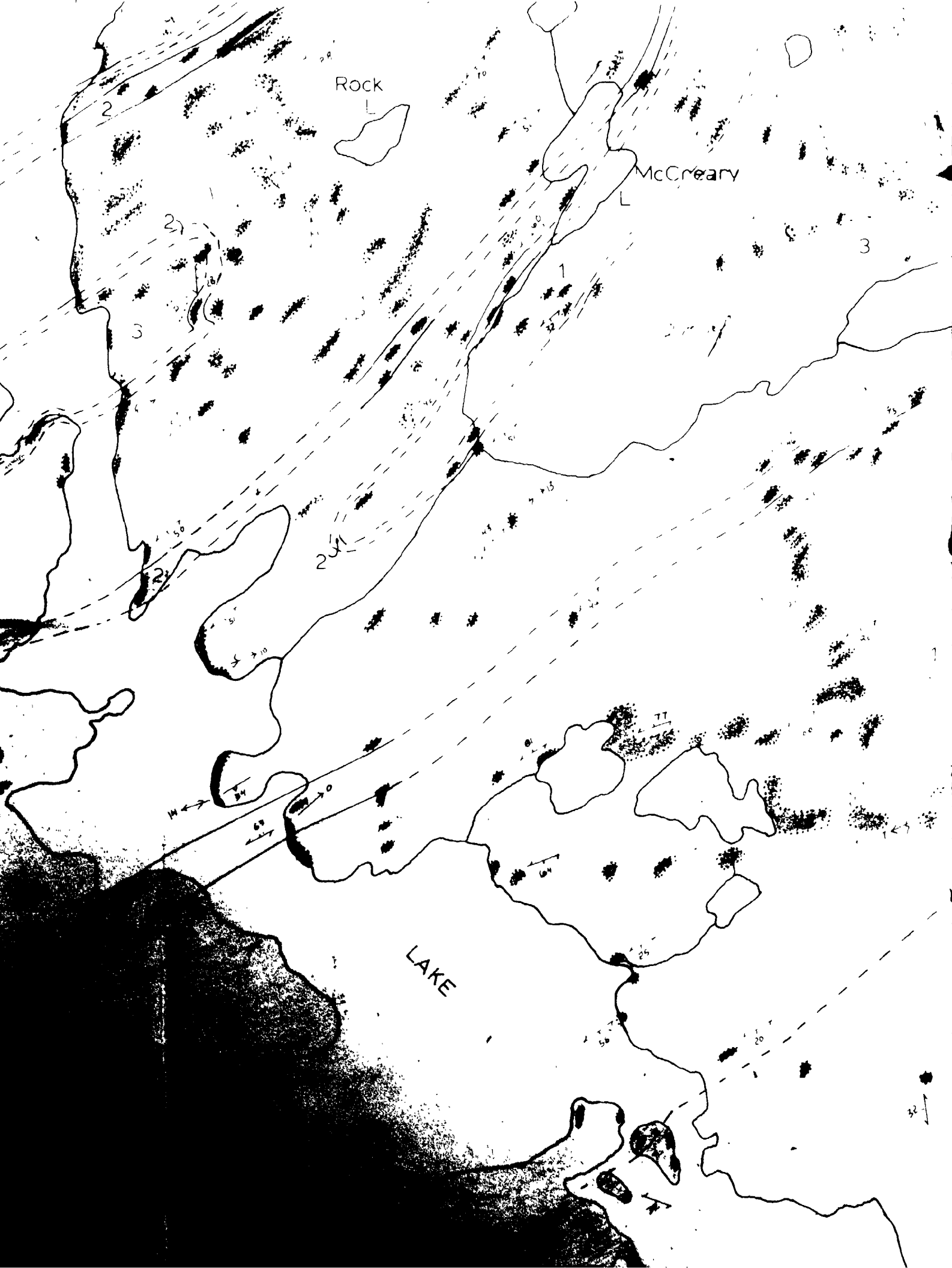
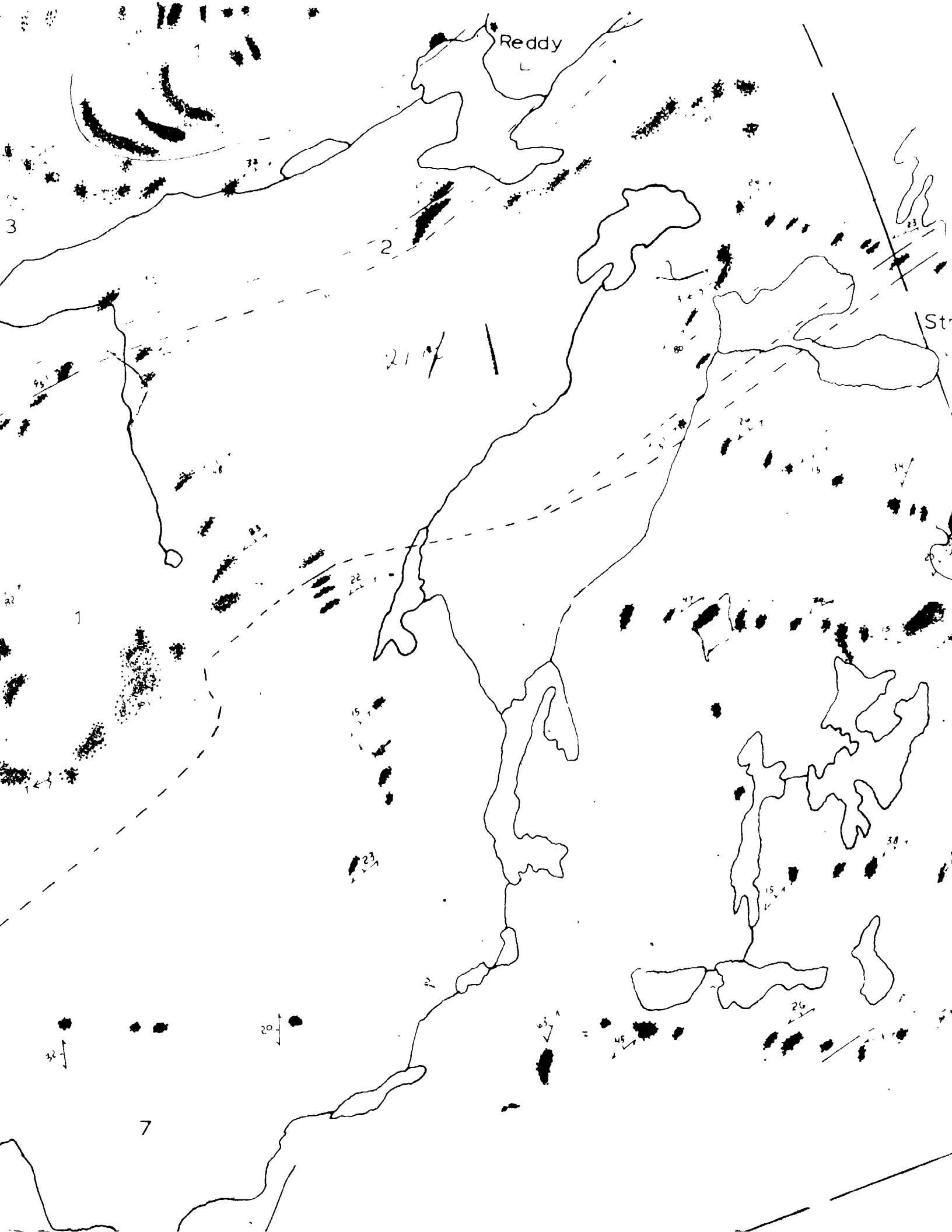
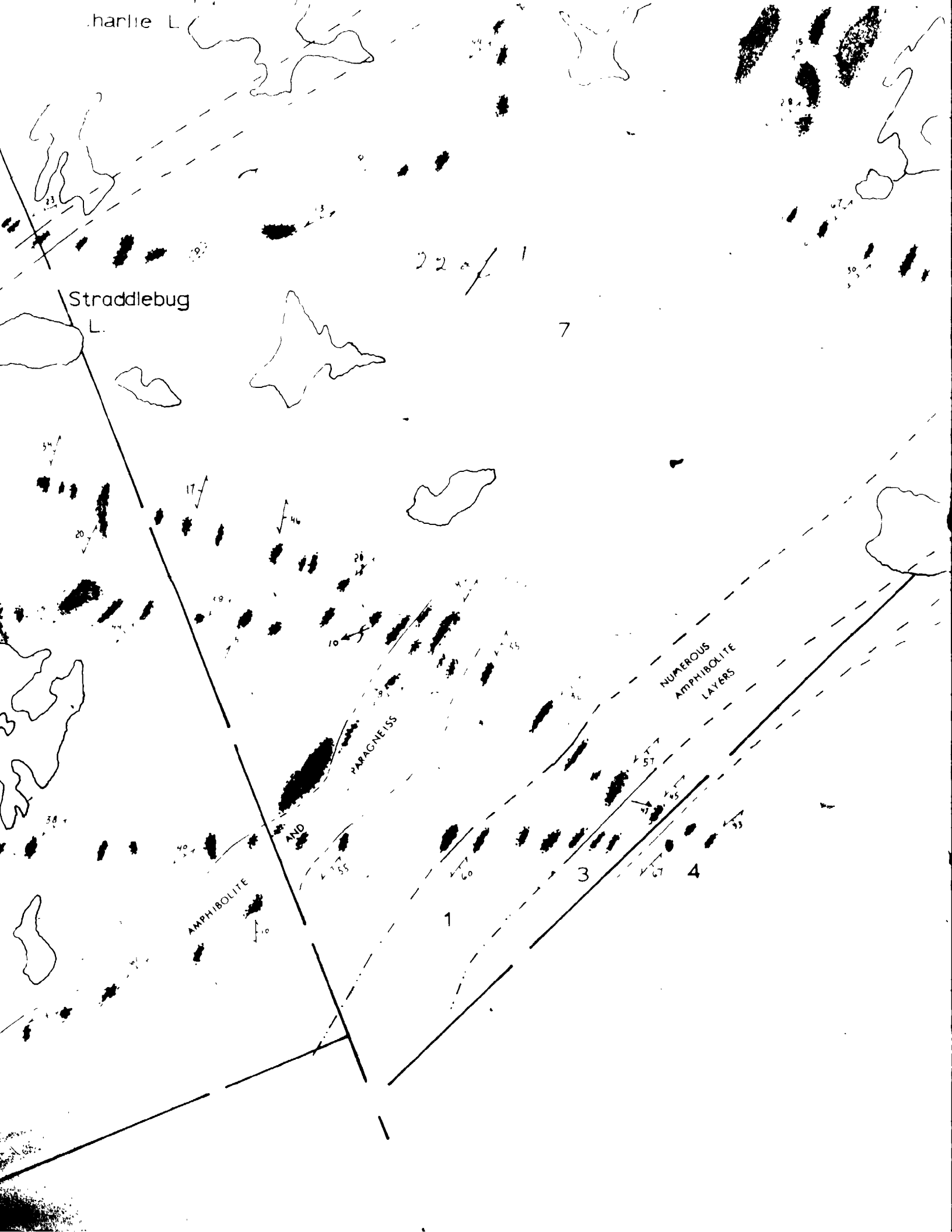
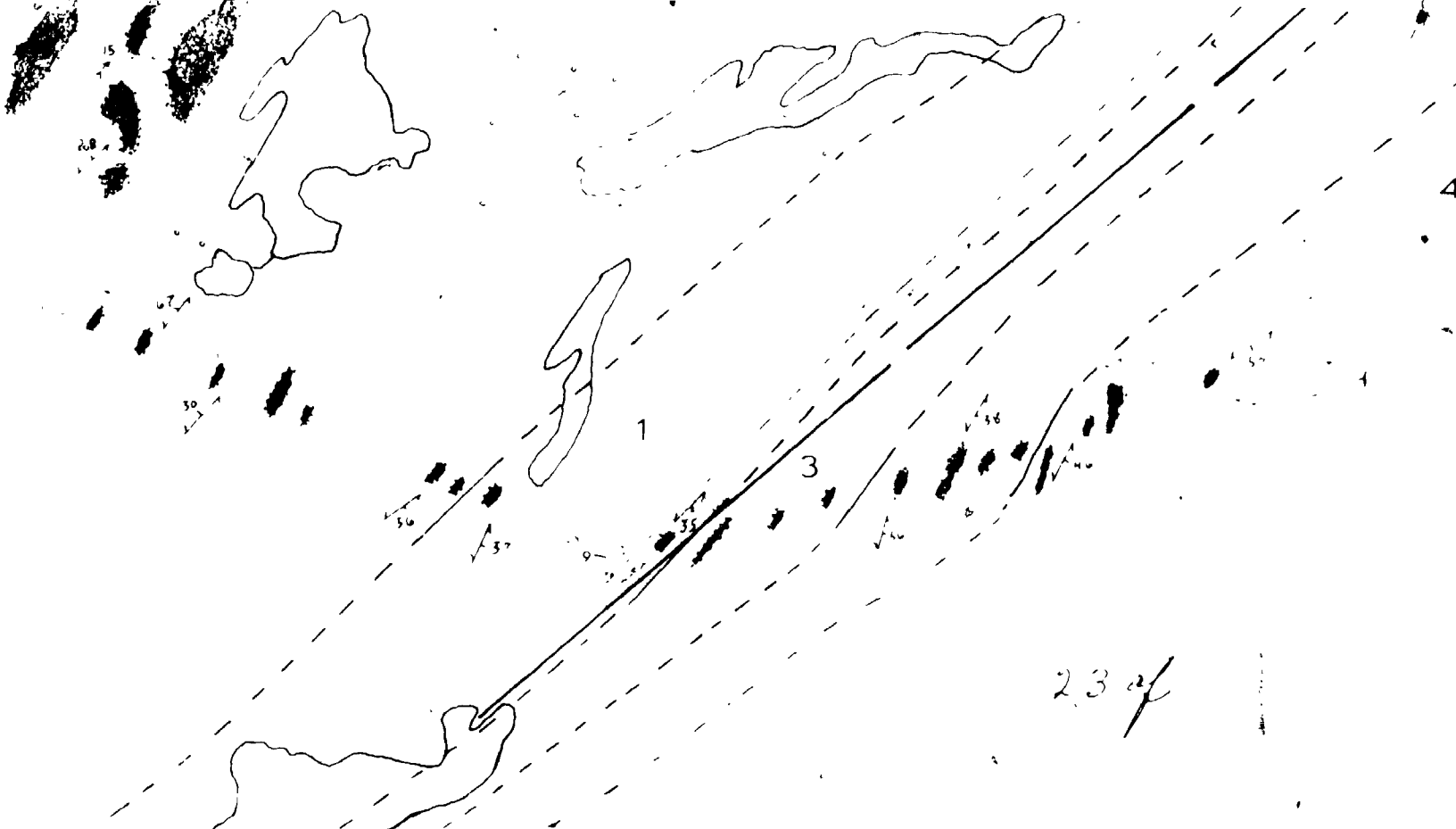
1984


Rock
L
McCreary
L
LAKE


Reddy
3
2
1
7


harlie L.
Straddlebug L.
7
NUMEROUS AMPHIBOLITE LAYERS
PARAGNEISS
AND
AMPHIBOLITE
1
3
4



LEGEND

GRENVILLE SUPERGROUP A
PLUTONIC RO

9 PEGMATITE

Intrusive Contac

METAMORPHOSED PLU

8 SYENITE: Red, inequigranular, homogeneous augen gneiss

TONALITE: Predominantly white to grey, medium to coars without hornblende and biotite. Contains unit

Intrusive Contac

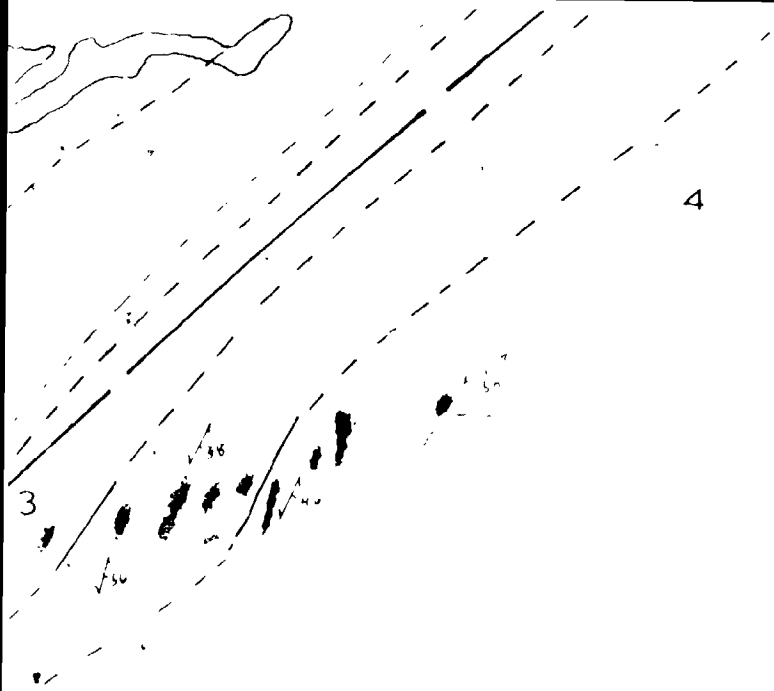
METAMORPHOSED SEDIM

UNDIFFERENTIATED MARBLE AND PARAGNEISS: Thinly to paragneiss.

MIXED PARAGNEISS: Sillimanite, garnet, amphibole, biotit gneisses. Contains units of quartzo-feldspathi

PARA-AMPHIBOLITE: Fine to medium-grained, slightly folia feldspar-quartz-biotite-plagioclase paragneiss clinopyroxene, layered amphibolite gneiss an volcanics.

3 CALCITIC MARBLE: White, medium-grained, homogeneous



LEGEND

GRENVILLE SUPERGROUP AND ASSOCIATED PLUTONIC ROCKS

MATITE

24 of

Intrusive Contact

METAMORPHOSED PLUTONIC ROCKS

NITE: Red, inequigranular, homogeneous augen gneiss.

ALITE: Predominantly white to grey, medium to coarse-grained, homogeneous orthogneiss, with or without hornblende and biotite. Contains units of dioritic, and anorthositic orthogneiss.

Intrusive Contact

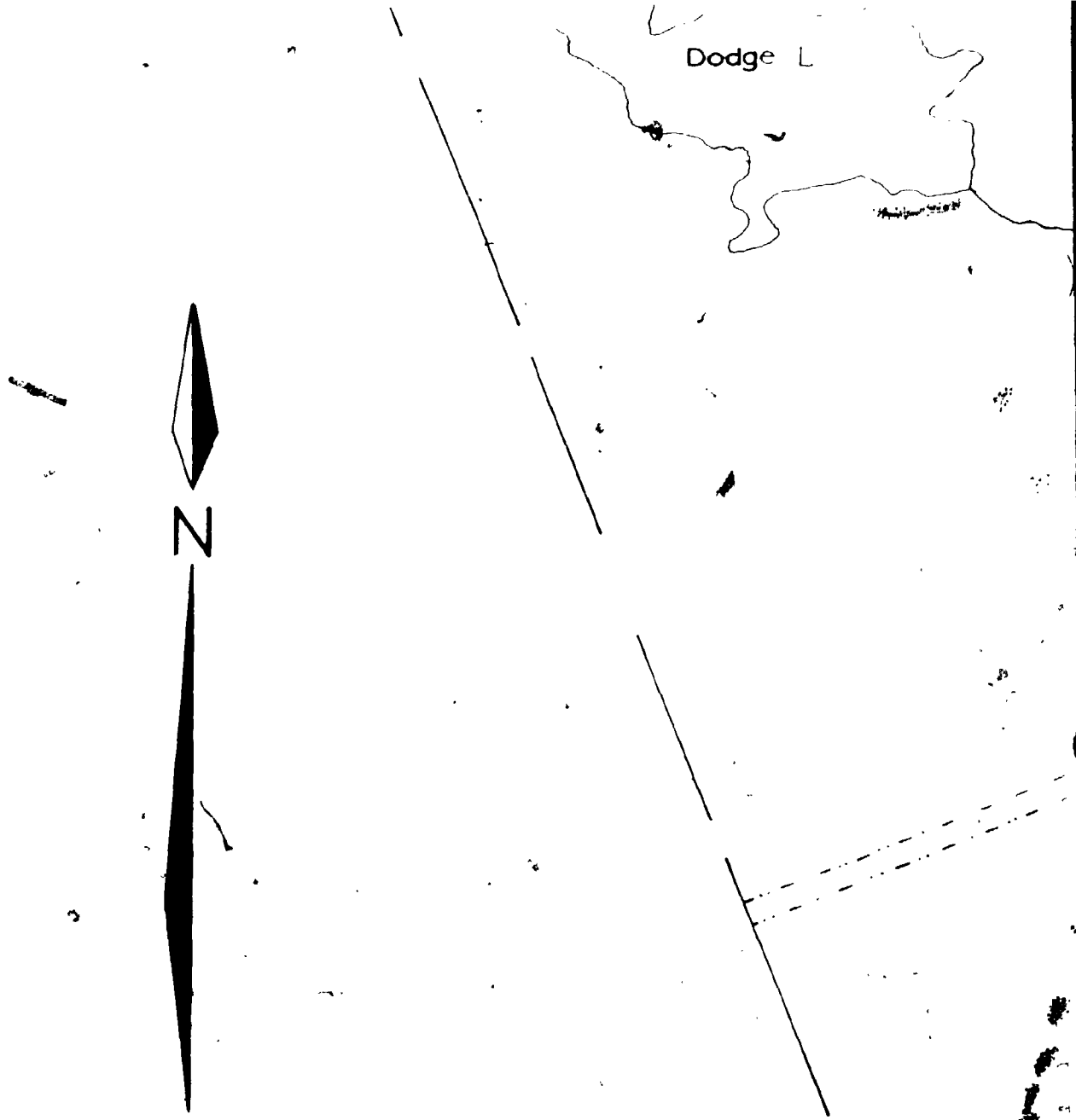
METAMORPHOSED SEDIMENTARY ROCKS

IFFERENTIATED MARBLE AND PARAGNEISS: Thinly to thickly inter-layered calcitic marble and paragneiss.

ED PARAGNEISS: Sillimanite, garnet, amphibole, biotite, quartz, plagioclase-bearing schists and gneisses. Contains units of quartzo-feldspathic gneiss and amphibolite.

A-AMPHIBOLITE: Fine to medium-grained, slightly foliated, garnet amphibolite interlayered with K-feldspar-quartz-biotite-plagioclase paragneiss with or without hornblende and clinopyroxene, layered amphibolite gneiss and calcitic marble. May in part contain mafic volcanics.

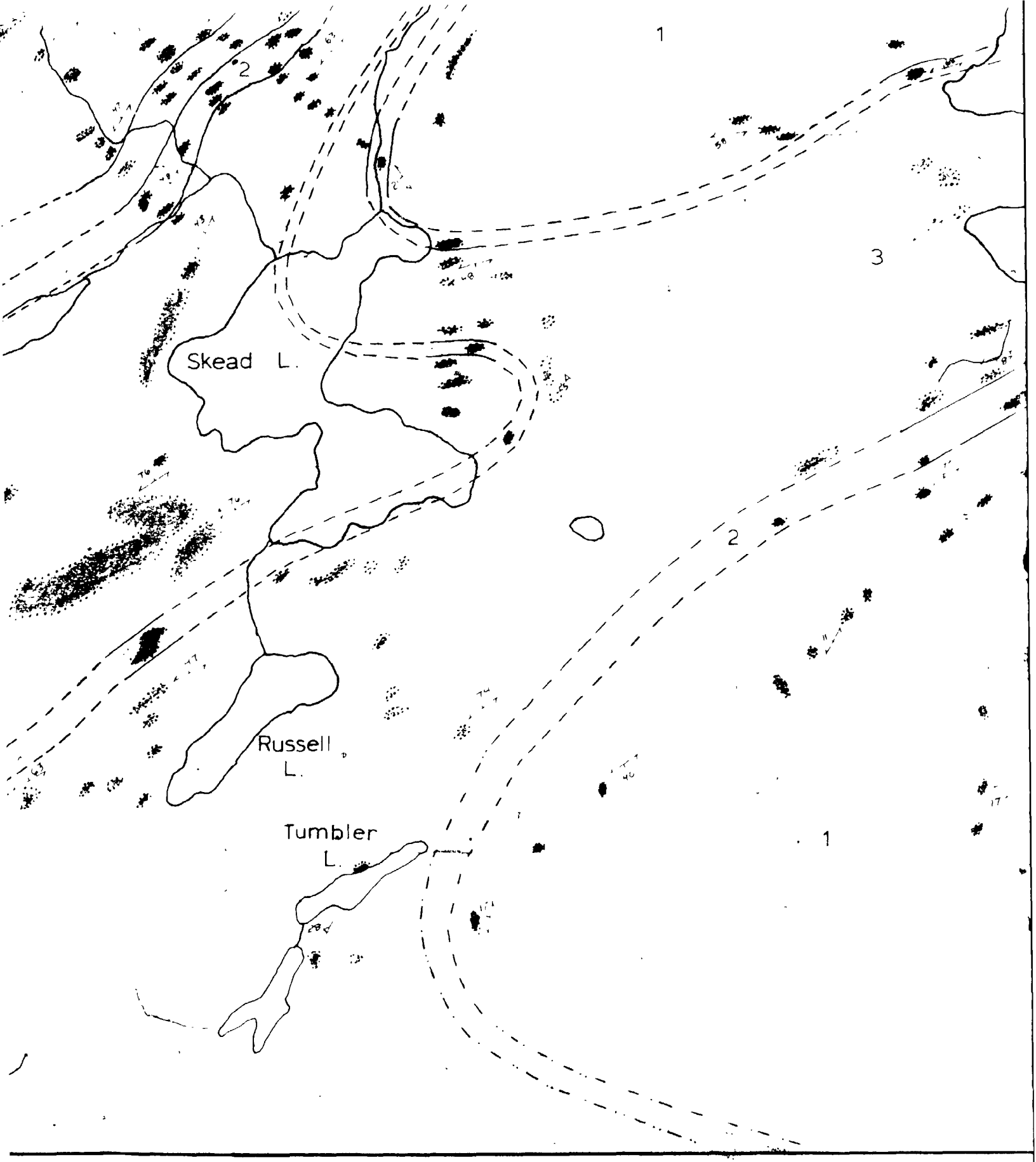
CITIC MARBLE: White, medium-grained, homogeneous, phlogopite-graphite marble, with or



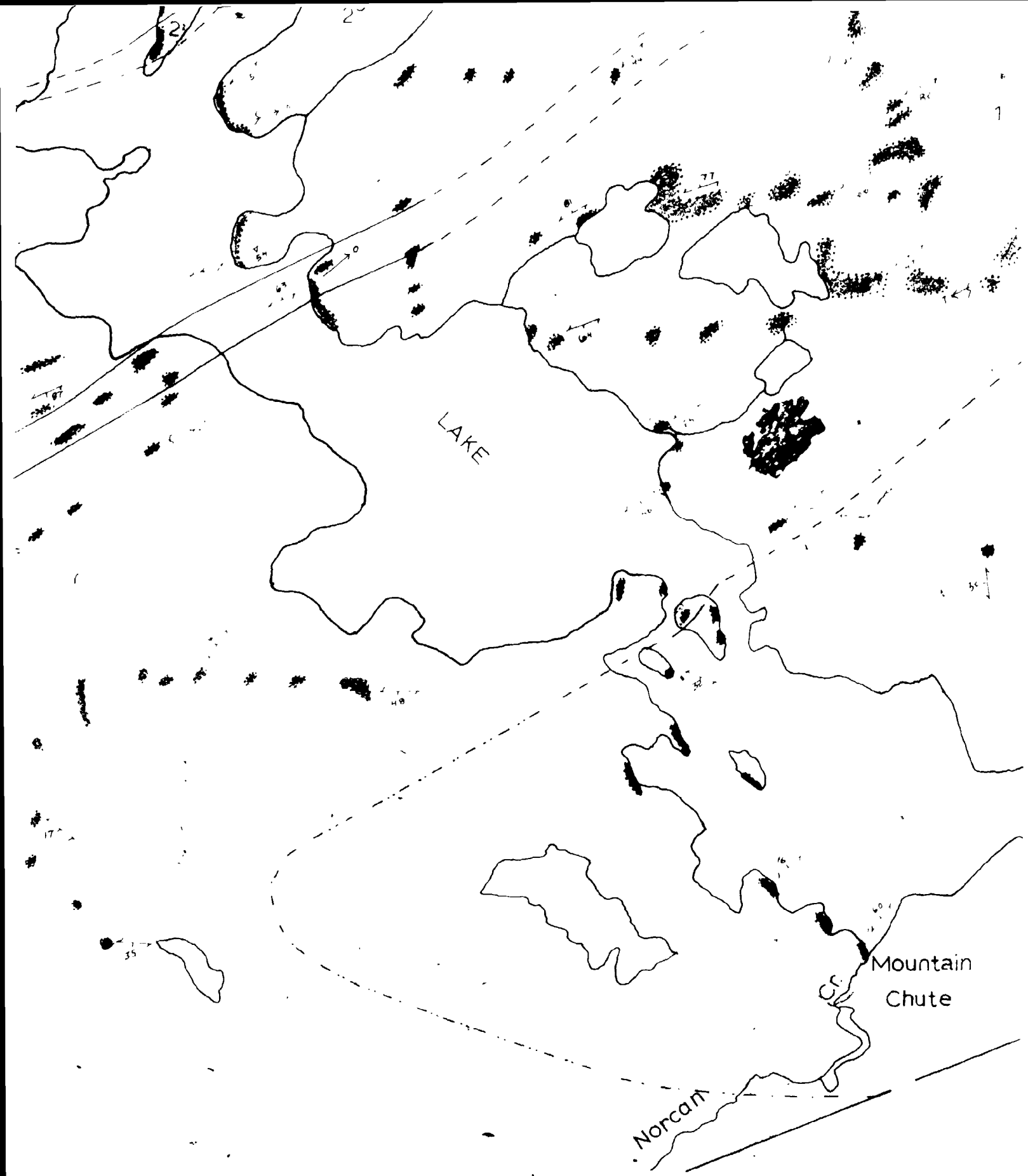
SYMBOLS

Strike and dip of foliation (inclined, vertical) .....
Strike and dip of foliation parallel to bedding (inclined, vertical) .....
Trend and plunge of minor fold axes: antiform, synform, showing sense of asymmetry .....
Trend and plunge of mineral lineation .....
Geologic contact (observed or readily inferred, inferred, speculative) .....
Outcrop or closely spaced outcrops .....
Township line .....

26 ft 1



27 of



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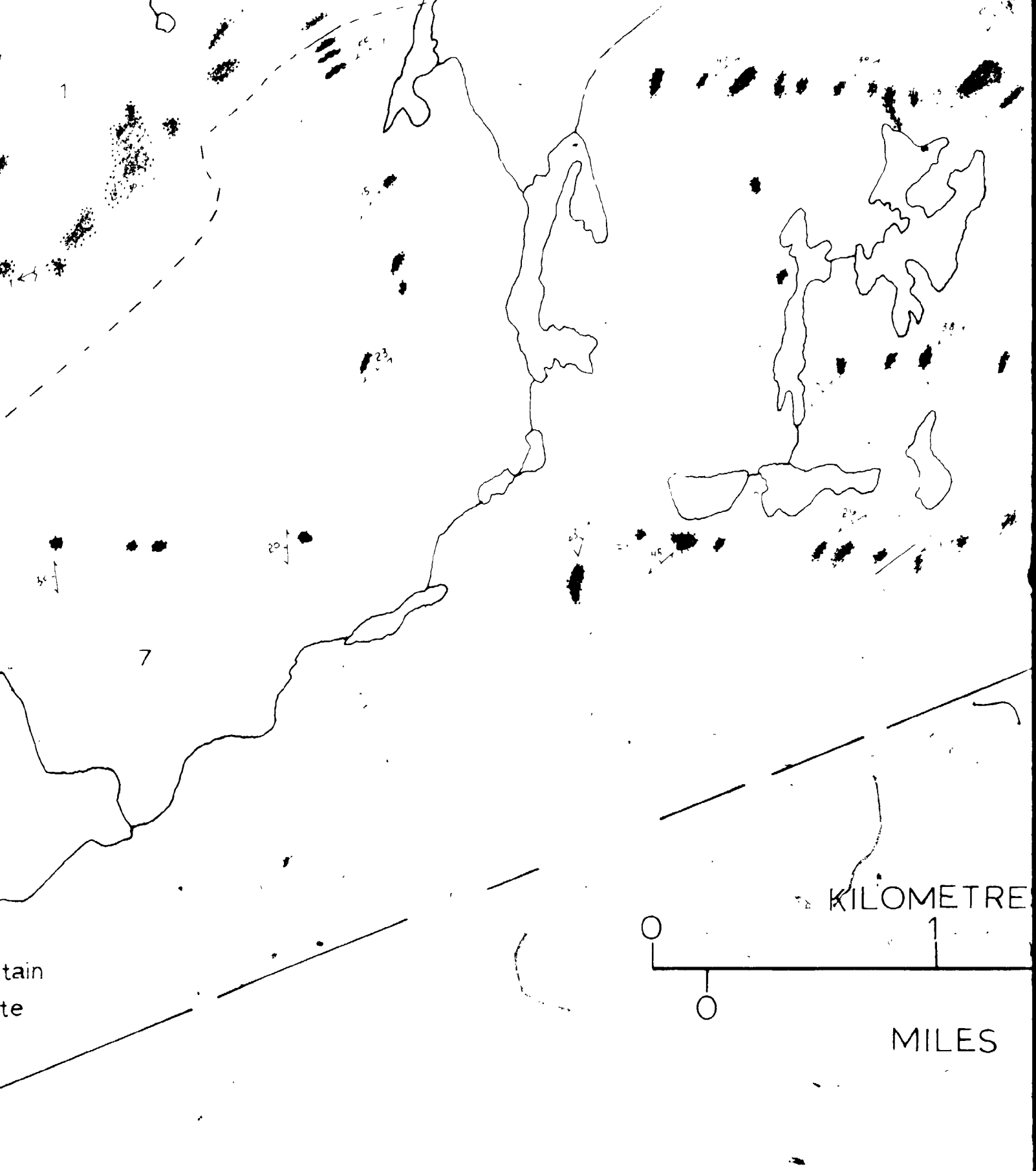
KILOMETRE
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MILES
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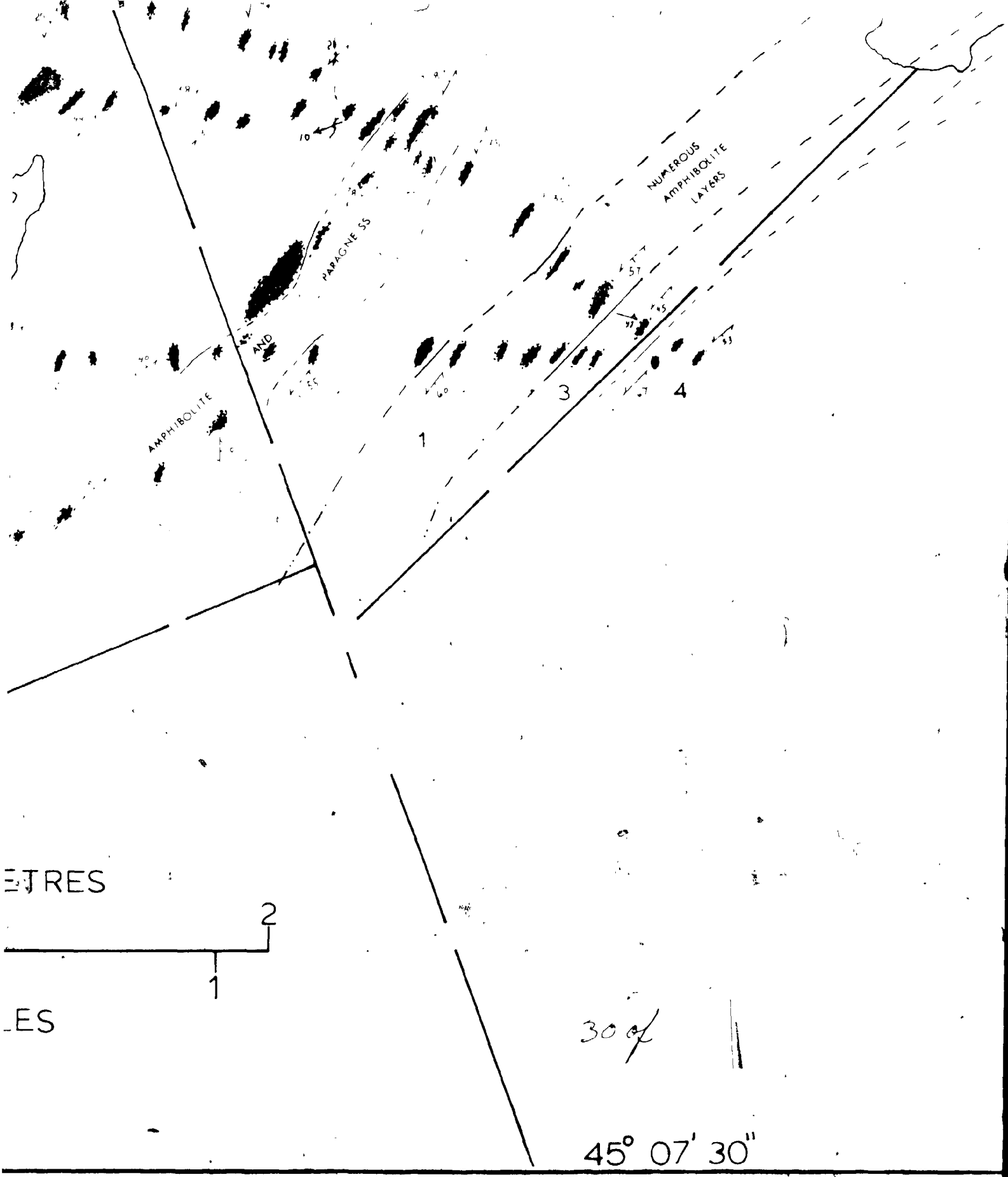

NUMEROUS
AMPHIBOLITE
LAYERS
PARAGNESS
AND
AMPHIBOLITE
1
3
4
ETRES
2
1
LES
30 of
45° 07' 30"

LEGEND

GRENVILLE SUPERGROUP AND ASSOCIATE PLUTONIC ROCKS

9 PEGMATITE

Intrusive Contact

METAMORPHOSED PLUTONIC ROCKS

8 SYENITE: Red, inequigranular, homogeneous augen gneiss

TONALITE: Predominantly white to grey, medium to coarse-grained, homogene without hornblende and biotite. Contains units of dioritic, and anor

Intrusive Contact

METAMORPHOSED SEDIMENTARY ROCK

6 UNDIFFERENTIATED MARBLE AND PARAGNEISS: Thinly to thickly inter-layered paragneiss.

5 MIXED PARAGNEISS: Sillimanite, garnet, amphibole, biotite, quartz, plagioclase gneisses. Contains units of quartzo-feldspathic gneiss and amphib

4 PARA-AMPHIBOLITE: Fine to medium-grained, slightly foliated, garnet amphibol feldspar-quartz-biotite-plagioclase paragneiss with or without horn clinopyroxene, layered amphibolite gneiss and calcitic marble. Ma volcanics.

3 CALCITIC MARBLE: White, medium-grained, homogeneous, phlogopite-graphite without biotite, diopside and tremolite. Contains units of amphibol interlayered marble and paragneiss.

METAMORPHOSED ROCKS OF UNCERTAIN C (LARGELY VOLCANICS?)

2 AMPHIBOLITIC GNEISS: Magnetite-biotite-quartz-hornblende-plagioclase gneiss interlayered amphibolite and granite gneiss.

1 HETEROGENEOUS GNEISS: Grey, fine to medium-grained, biotite-hornblende q coarsely interlayered with pink, fine to medium-grained granite gn amphibolite and paragneiss.

LEGEND

GRENVILLE SUPERGROUP AND ASSOCIATED PLUTONIC ROCKS

PEGMATITE

Intrusive Contact

METAMORPHOSED PLUTONIC ROCKS

SYENITE: Red, inequigranular, homogeneous augen gneiss

TONALITE. Predominantly white to grey, medium to coarse-grained, homogeneous orthogneiss, with or without hornblende and biotite Contains units of dioritic, and anorthositic orthogneiss

Intrusive Contact

METAMORPHOSED SEDIMENTARY ROCKS

UNDIFFERENTIATED MARBLE AND PARAGNEISS: Thinly to thickly inter-layered calcitic marble and paragneiss

MIXED PARAGNEISS. Sillimanite, garnet, amphibole, biotite, quartz, plagioclase-bearing schists and gneisses. Contains units of quartzo-feldspathic gneiss and amphibolite.

PARA-AMPHIBOLITE: Fine to medium-grained, slightly foliated, garnet amphibolite interlayered with K-feldspar-quartz-biotite-plagioclase paragneiss with or without hornblende and clinopyroxene, layered amphibolite gneiss and calcitic marble. May in part contain mafic volcanics.

CALCITIC MARBLE: White, medium-grained, homogeneous, phlogopite-graphite marble, with or without biotite, diopside and tremolite. Contains units of amphibolite and thinly interlayered marble and paragneiss.

?

METAMORPHOSED ROCKS OF UNCERTAIN ORIGIN (LARGELY VOLCANICS?)

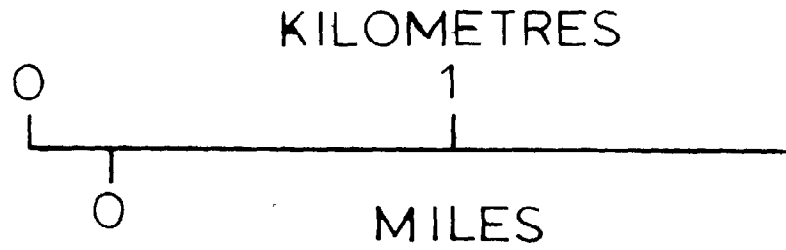
AMPHIBOLITIC GNEISS: Magnetite-biotite-quartz-hornblende-plagioclase gneiss, or thinly interlayered amphibolite and granite gneiss.

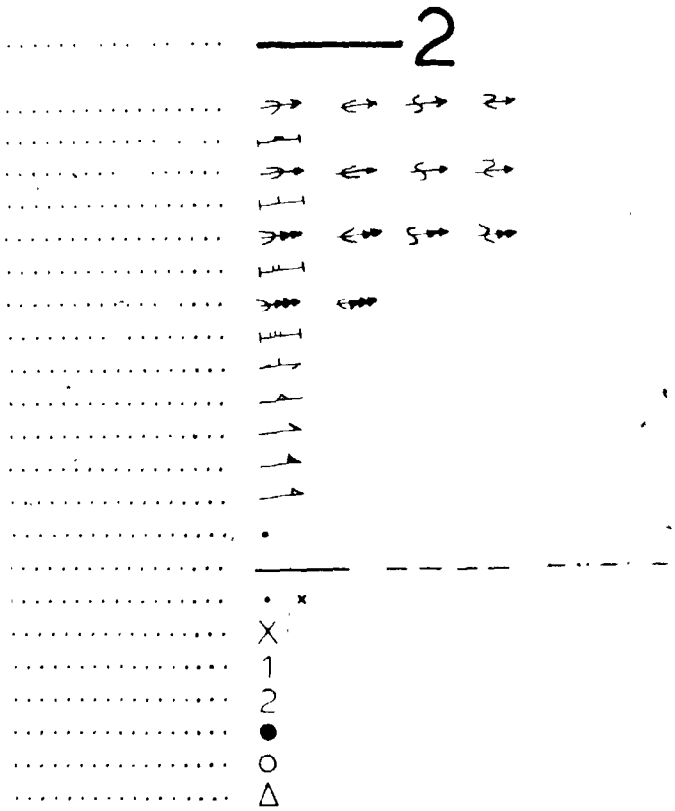
HETEROGENEOUS GNEISS: Grey, fine to medium-grained, biotite-hornblende quartz diorite gneiss coarsely interlayered with pink, fine to medium-grained granite gneiss. Contains units of amphibolite and paragneiss.

1 of 1

LEGEND

Subarea boundary and number
Trend and plunge of fold axes of uncertain generation:
antiform, synform, showing sense of asymmetry
Strike and dip of fold axial plane of uncertain generation
Trend and plunge of F1 fold: antiform, synform, showing sense of asymmetry
Strike and dip of F1 fold axial plane
Trend and plunge of F2 fold: antiform, synform, showing sense of asymmetry
Strike and dip of F2 fold axial plane
Trend and plunge of F3 fold: antiform, synform
Strike and dip of F3 fold axial plane
Strike and dip of foliation
Strike and dip of foliation parallel to bedding
Trend and plunge of mineral lineation
Trend and plunge of long axis of boudinage structure
Trend and plunge of mullion structure
Data station
Geologic contact (observed or readily inferred, inferred, speculative)
Stereonet diagrams: poles to foliation, foliation parallel to bedding
fold axes of uncertain generation
F1 fold axes
F2 fold axes
mineral lineations
long axes of boudins
mullion structure

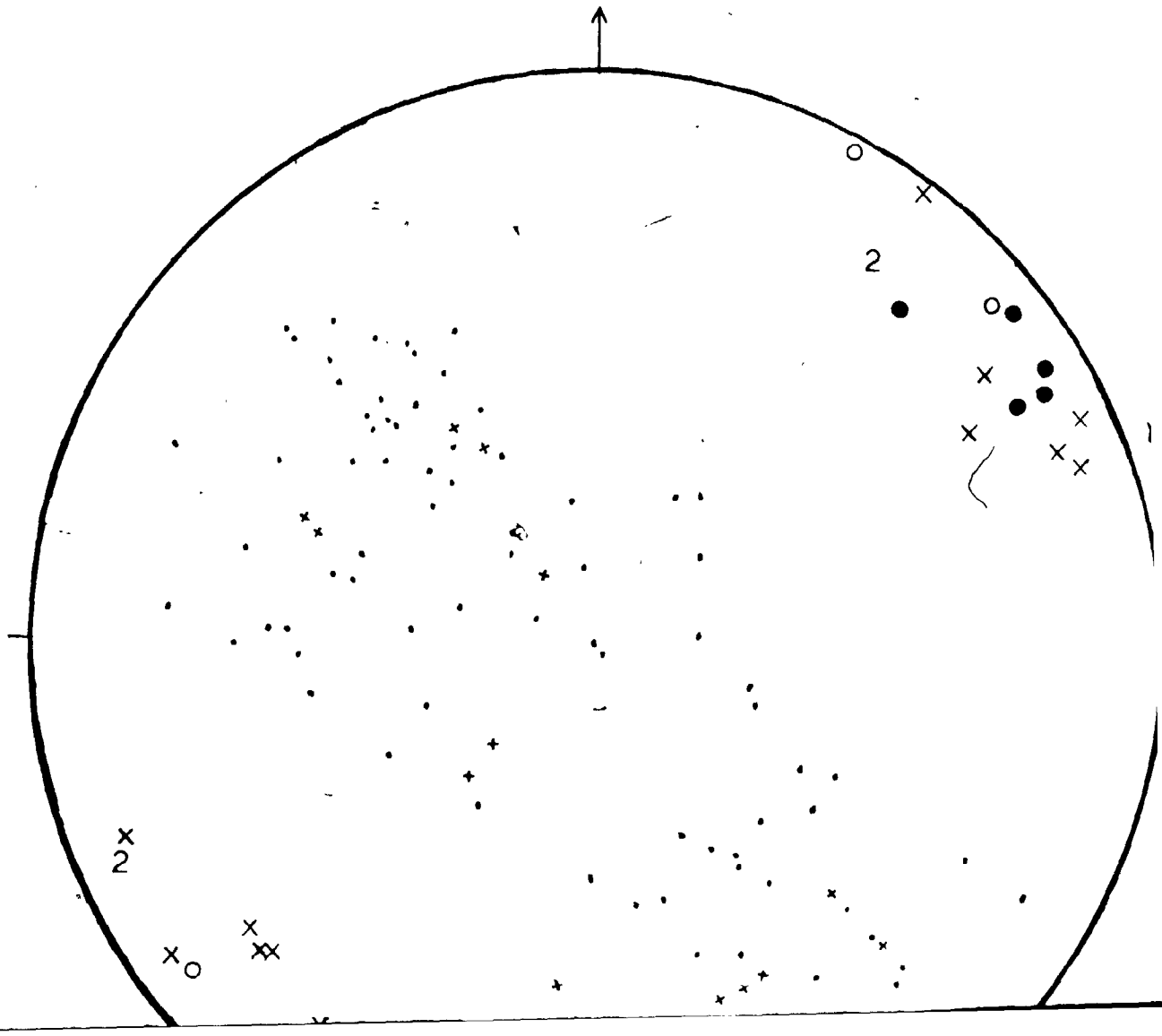




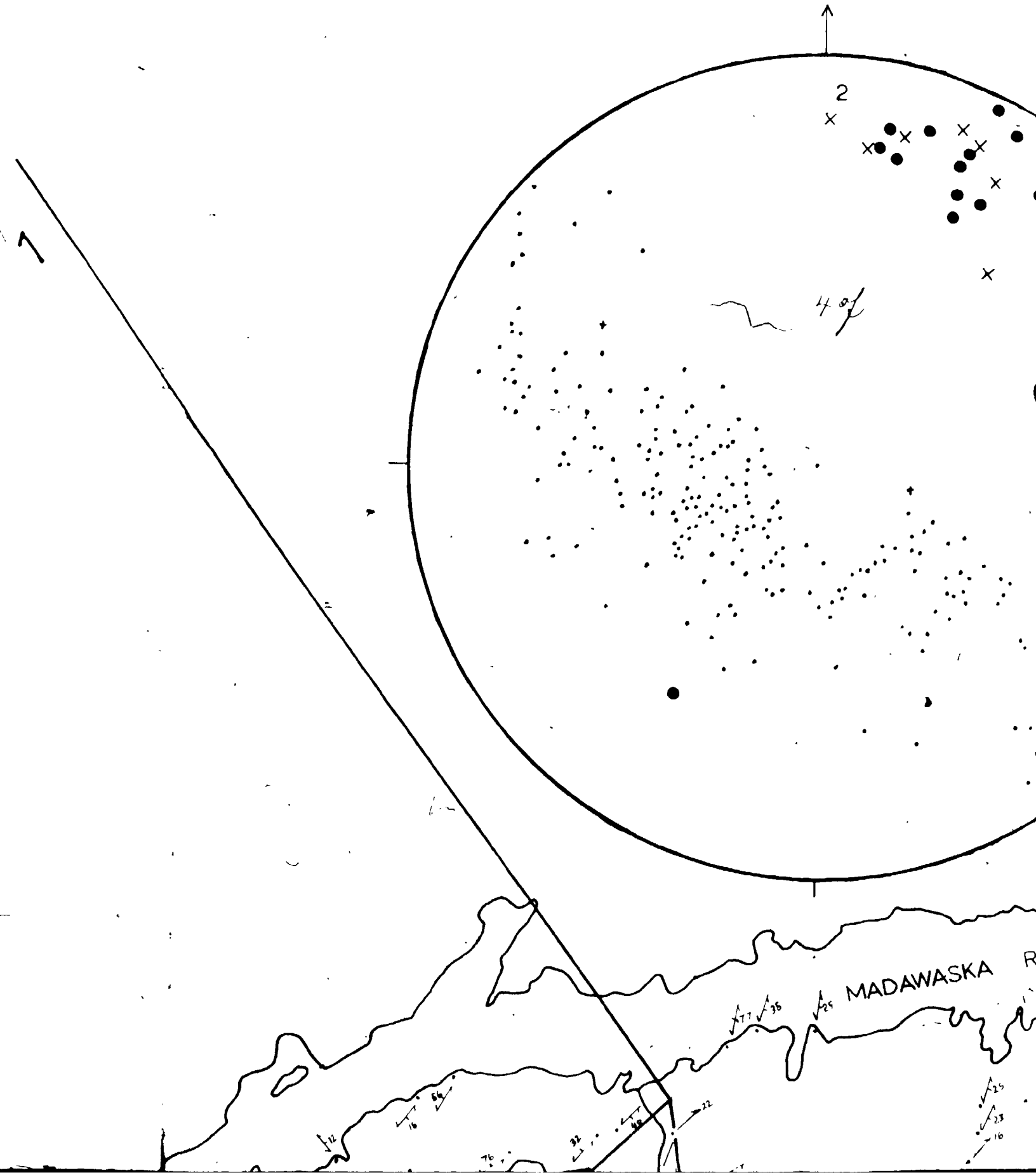
2 of 1

STRUCTURAL GI

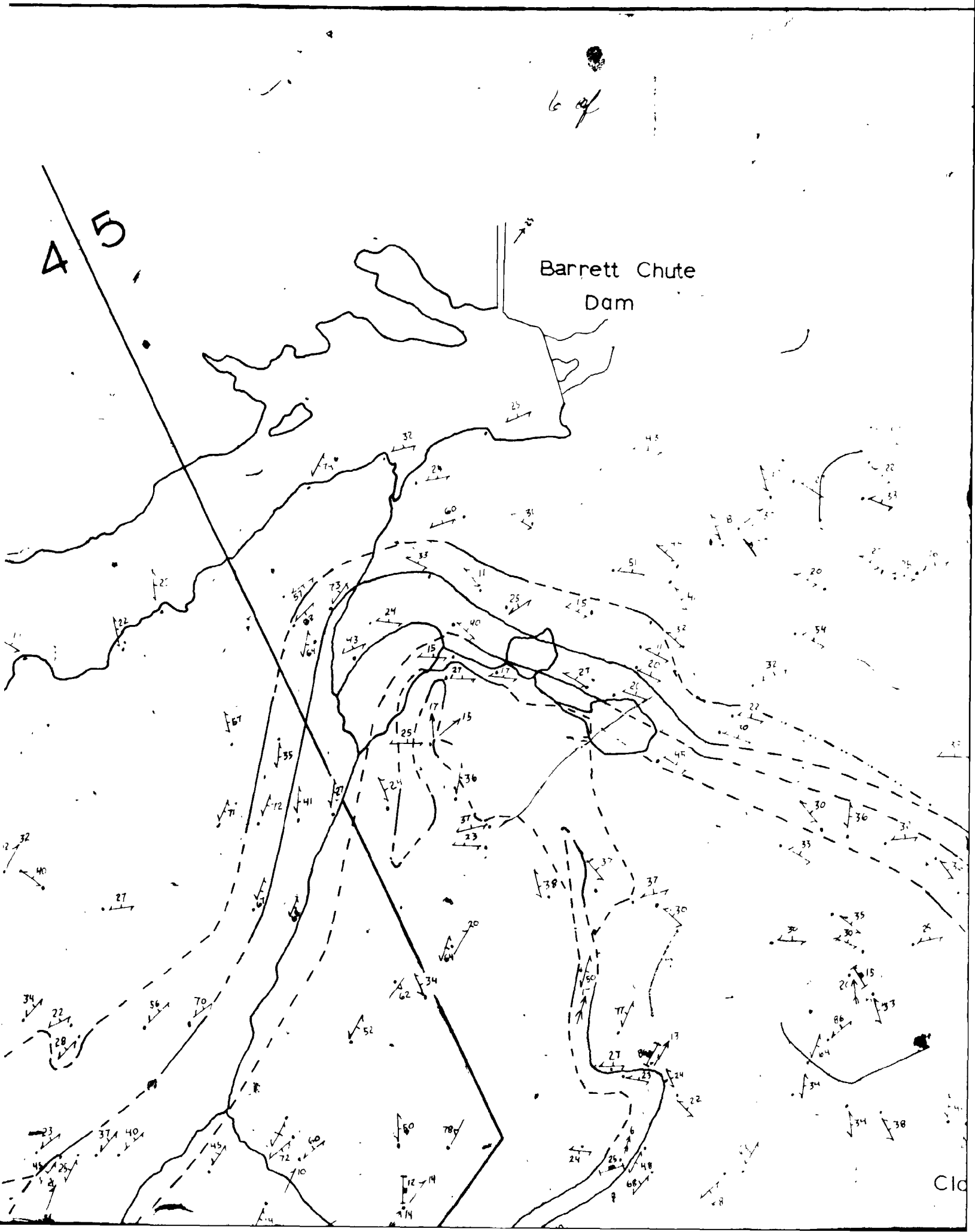
3 of



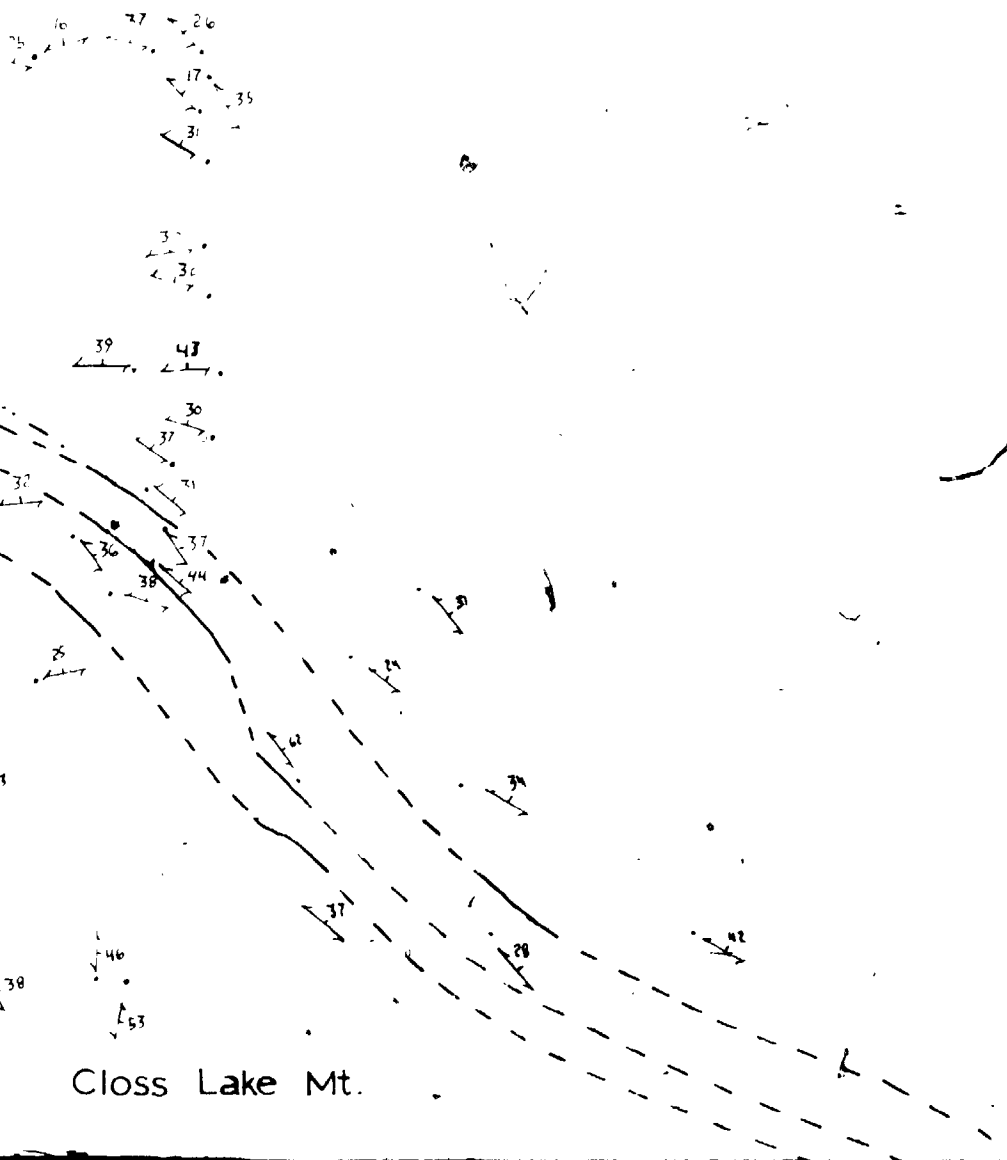
EOLOGY OF THE NORC



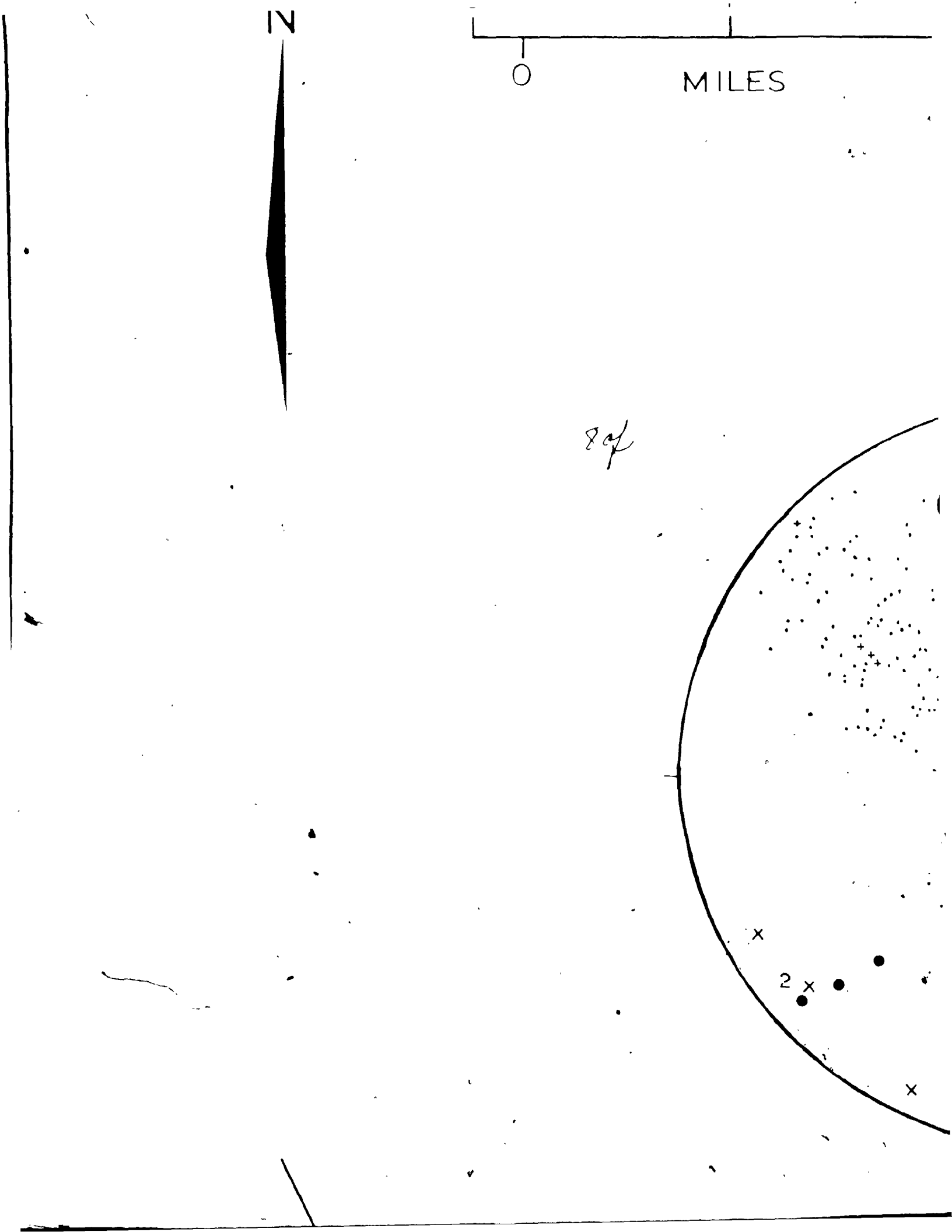
RCAN LAKE AREA


Barrett Chute
Dam

7 of



Closs Lake Mt.


N
0
MILES
8 of
2

2

1

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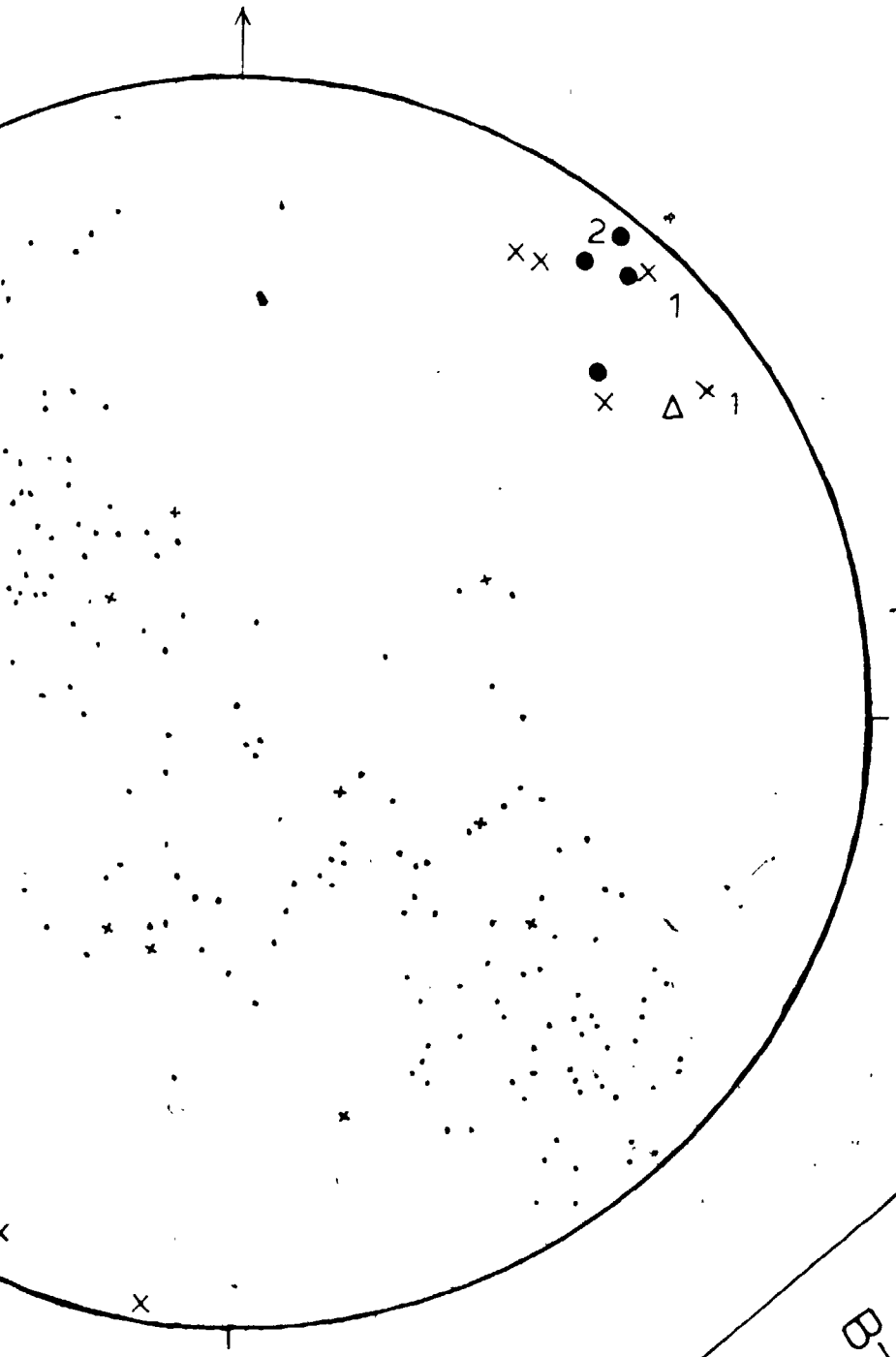

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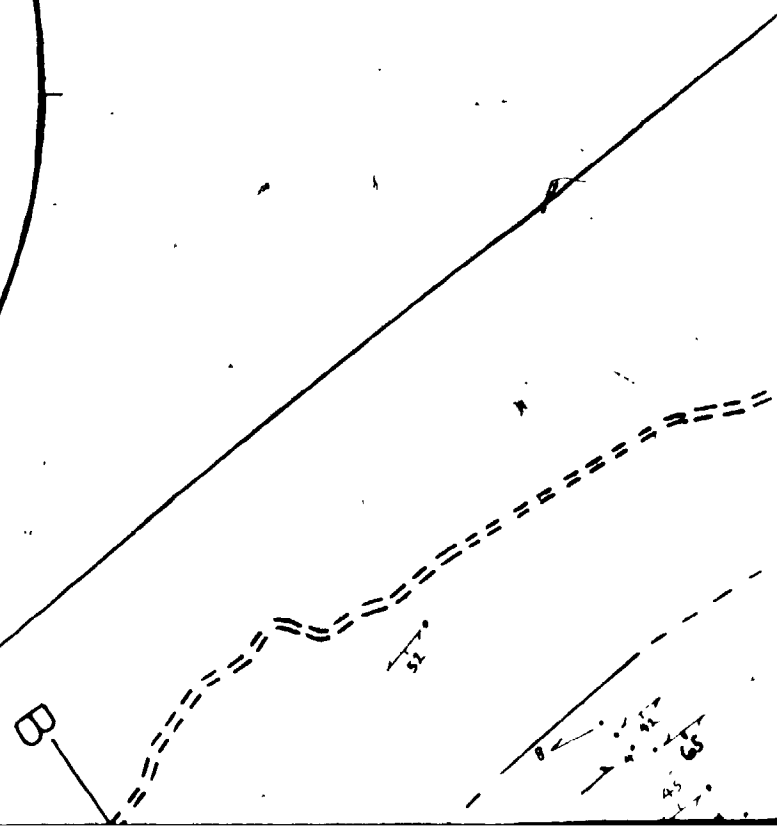
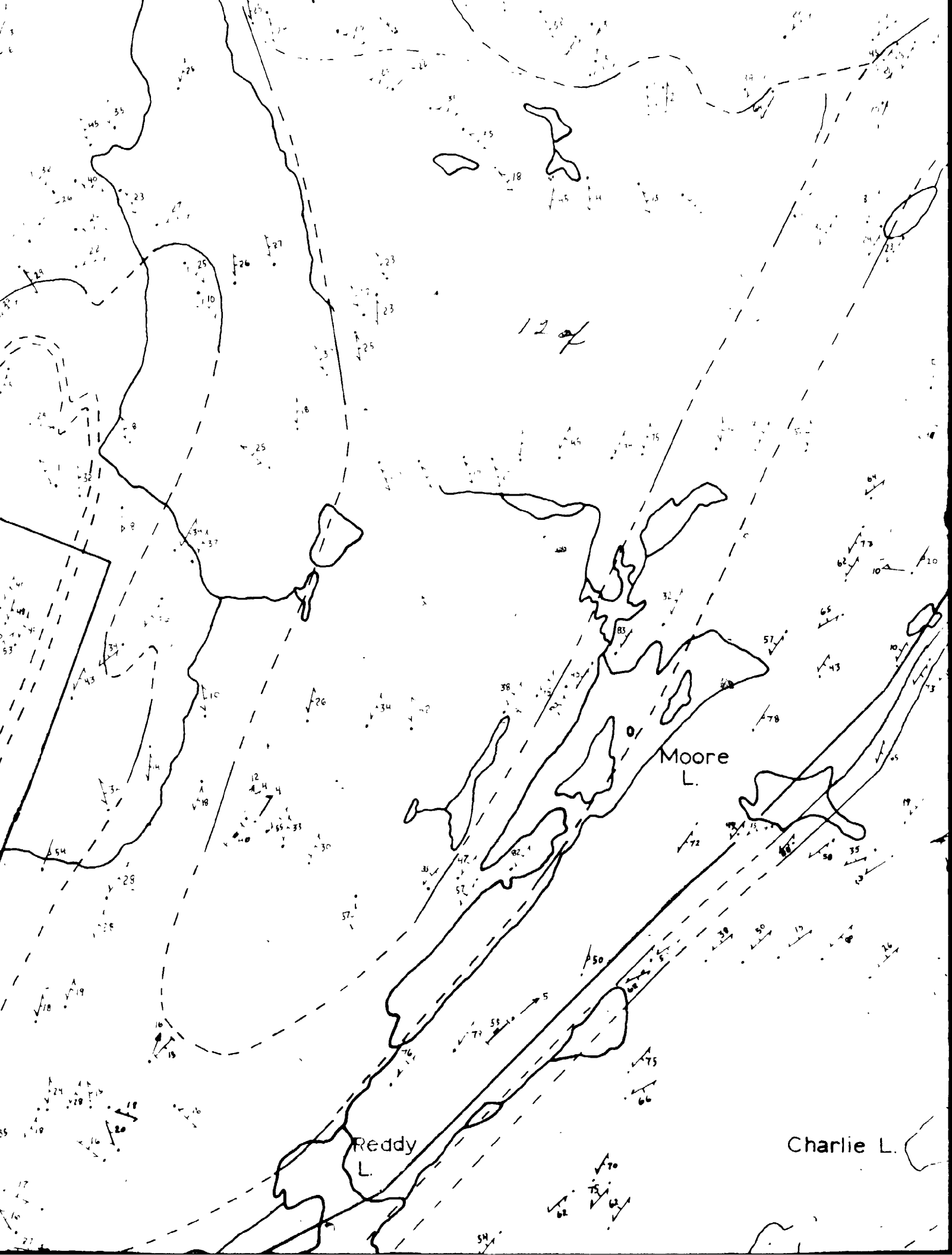
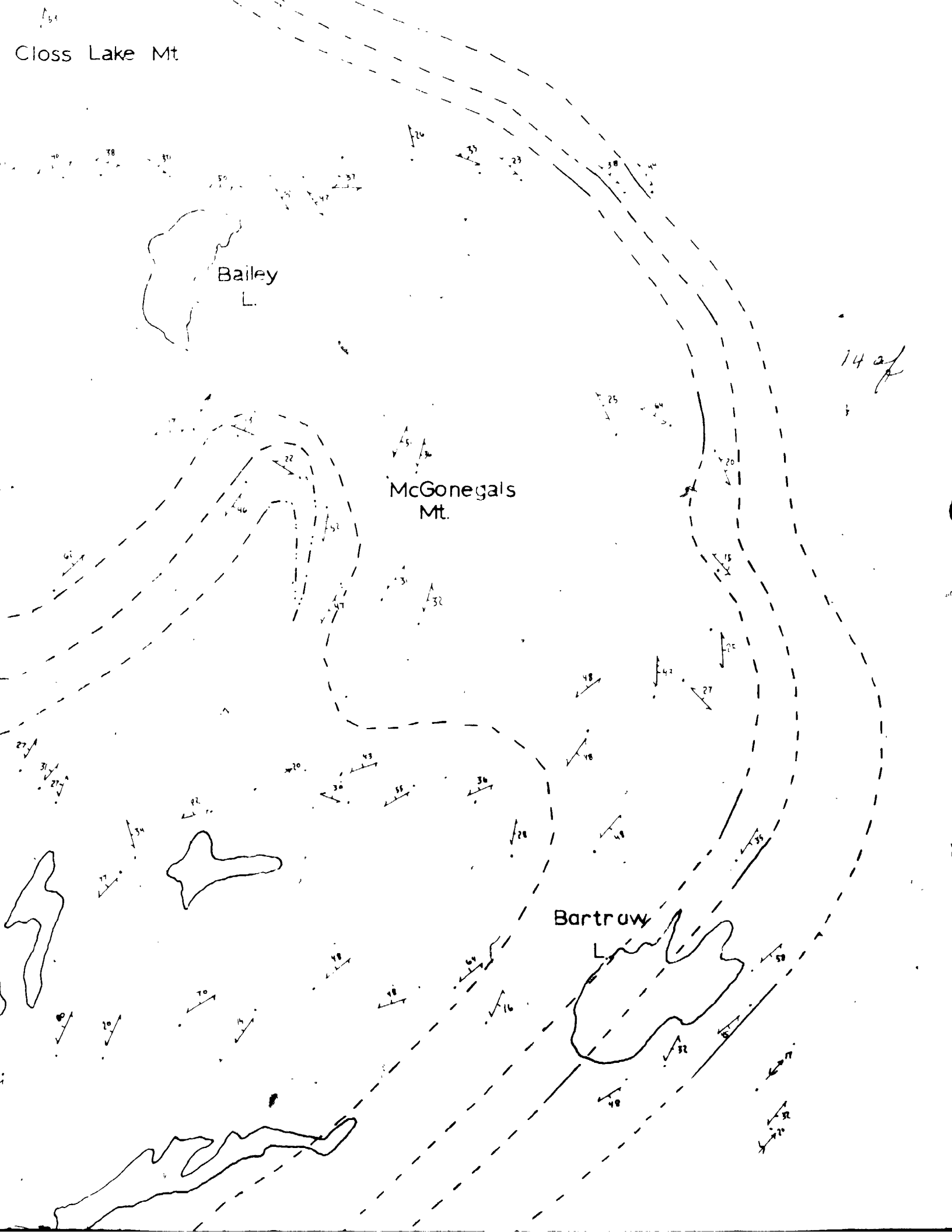



NORCAN


Mt.
McCreary

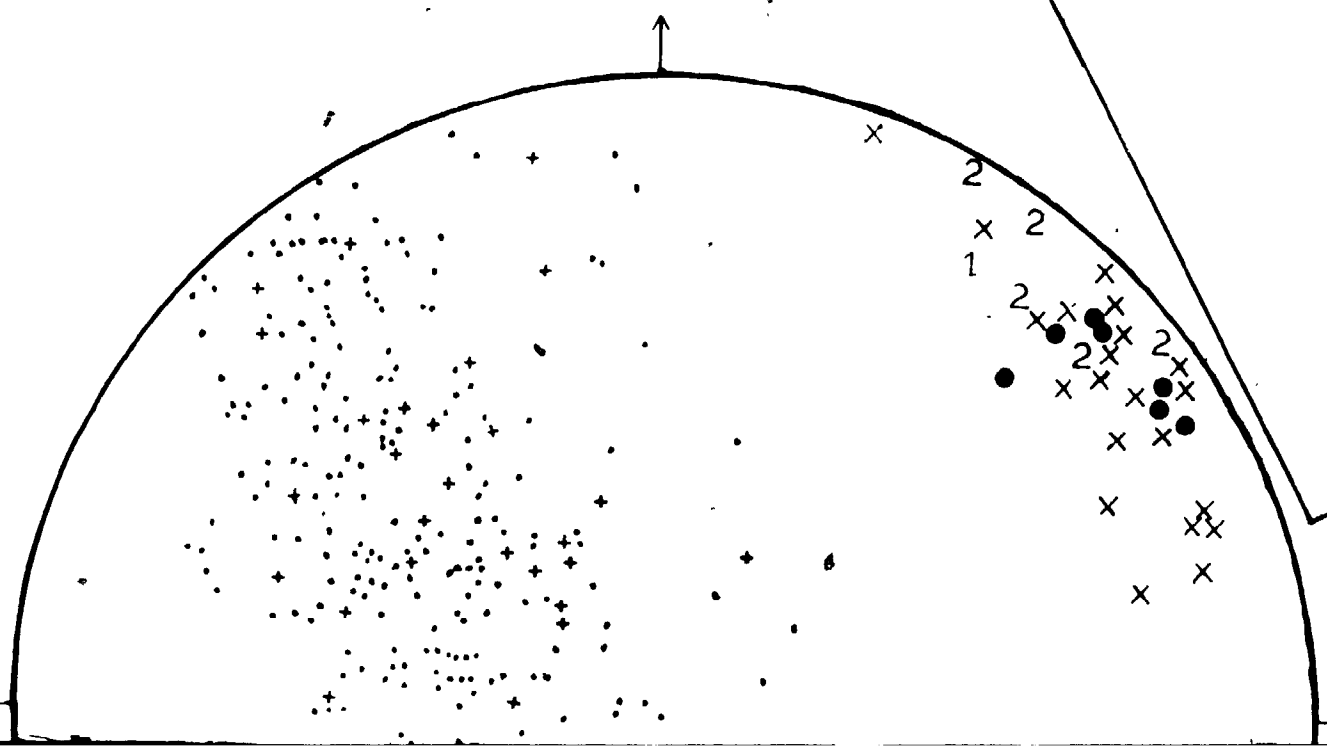

Moore L.
Reddy L.
Charlie L.

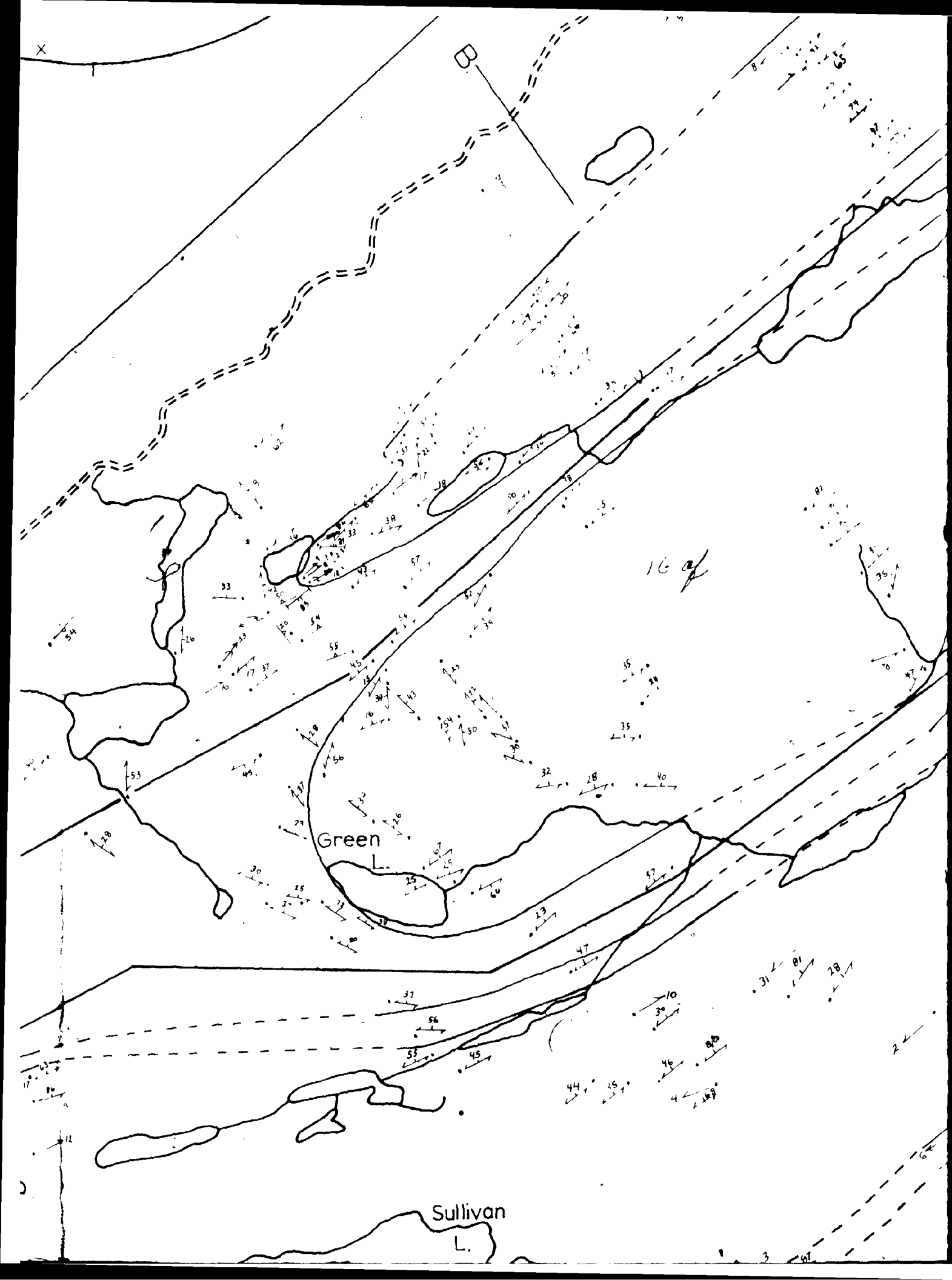

Closs Lake Mt
13 of 1

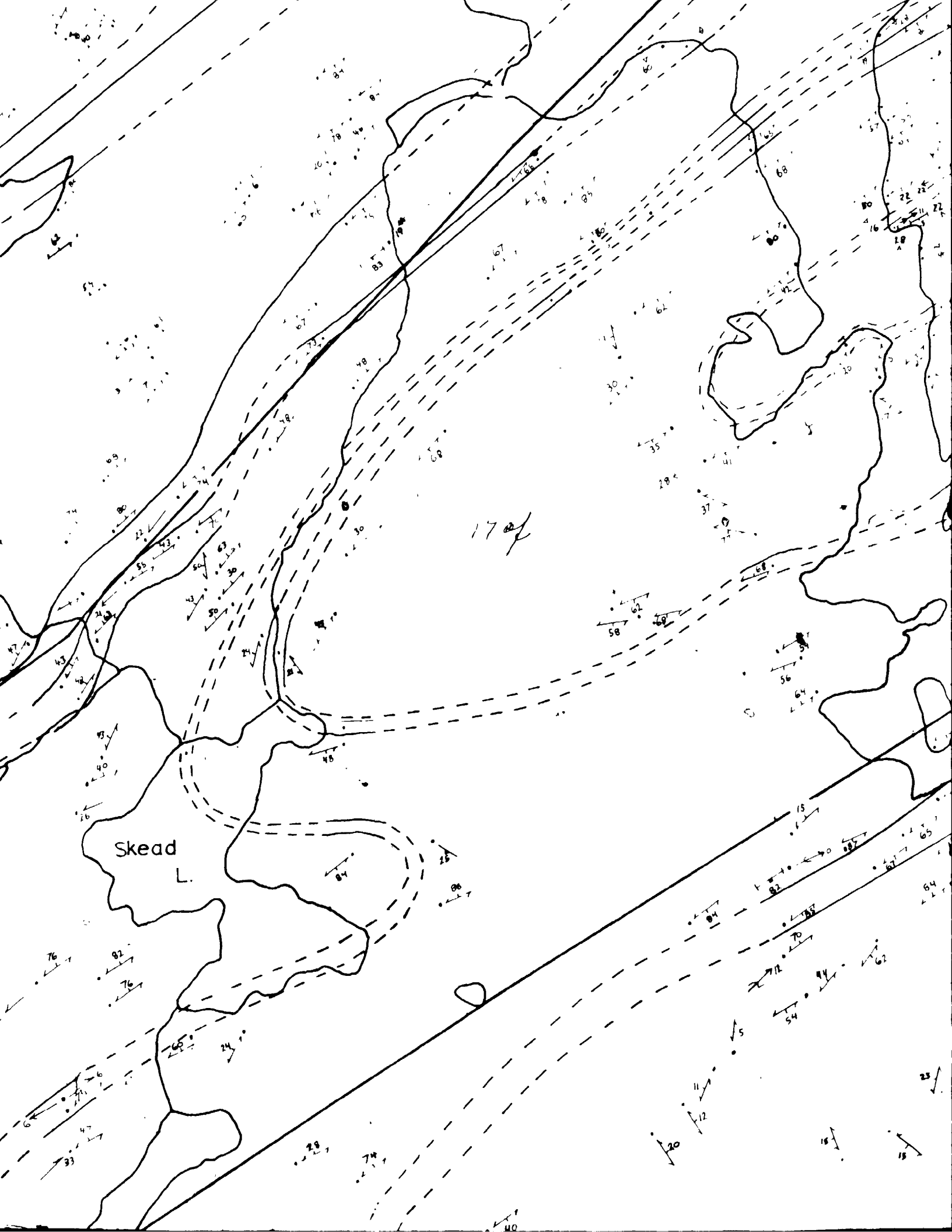

Closs Lake Mt
Bailey
L.
McGonegals
Mt.
Bartrow
L.

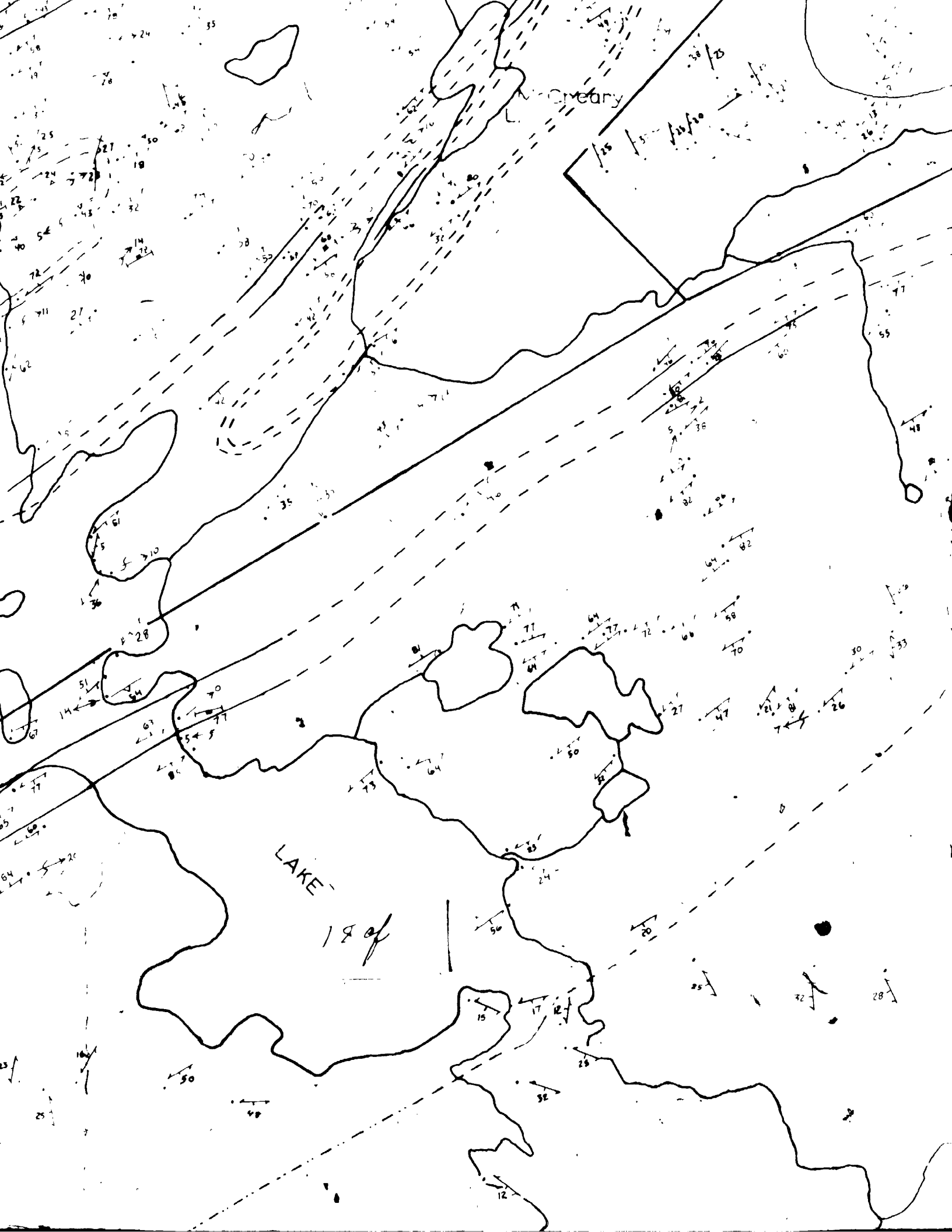
3

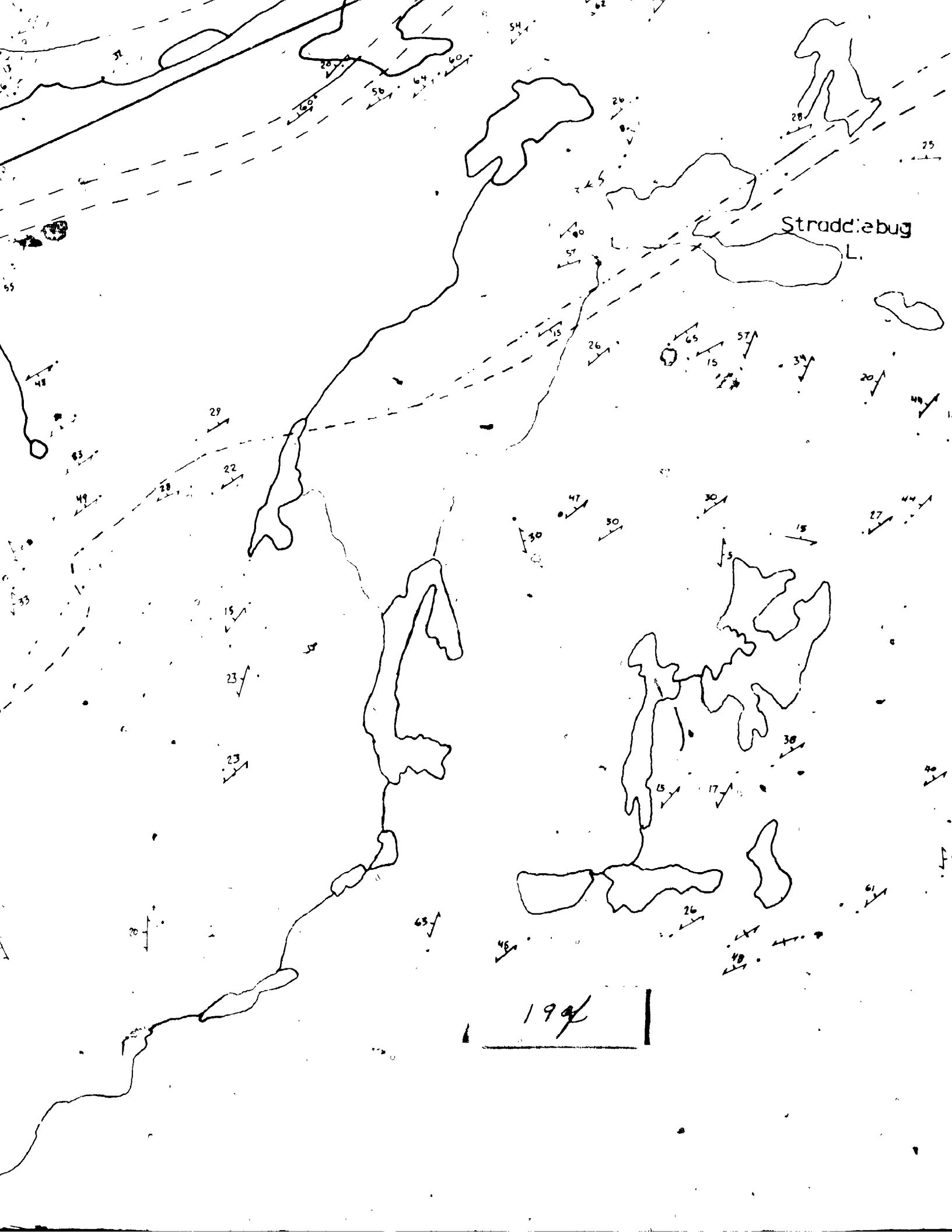
15 of

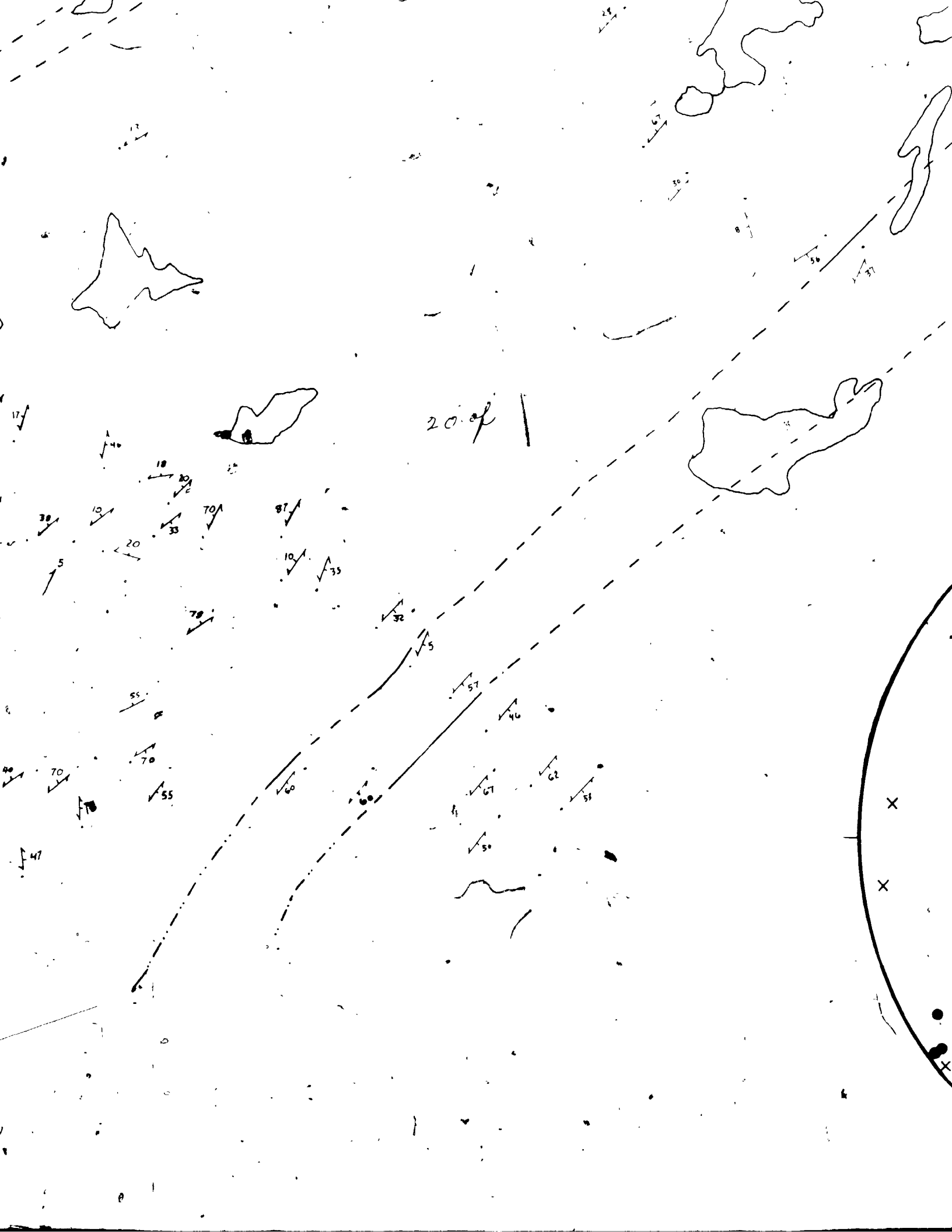


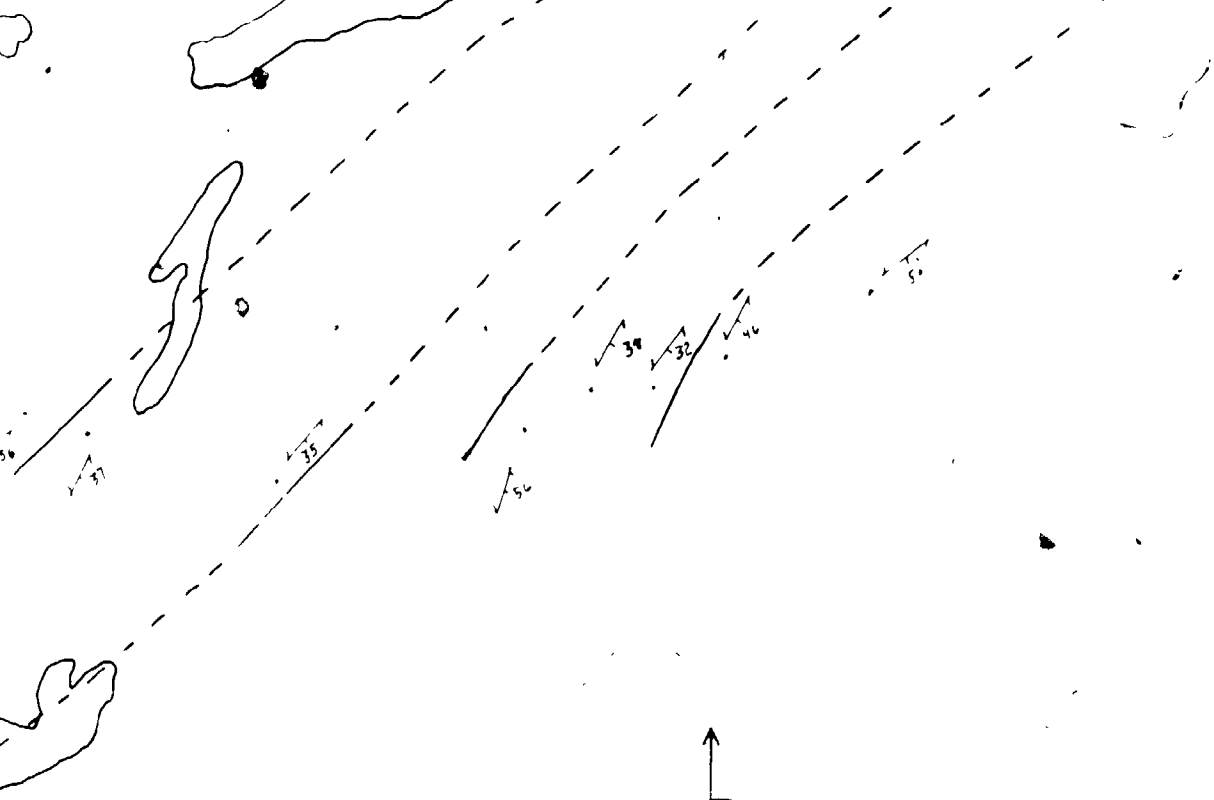

B
16 a
Green
L.
Sullivan
L.

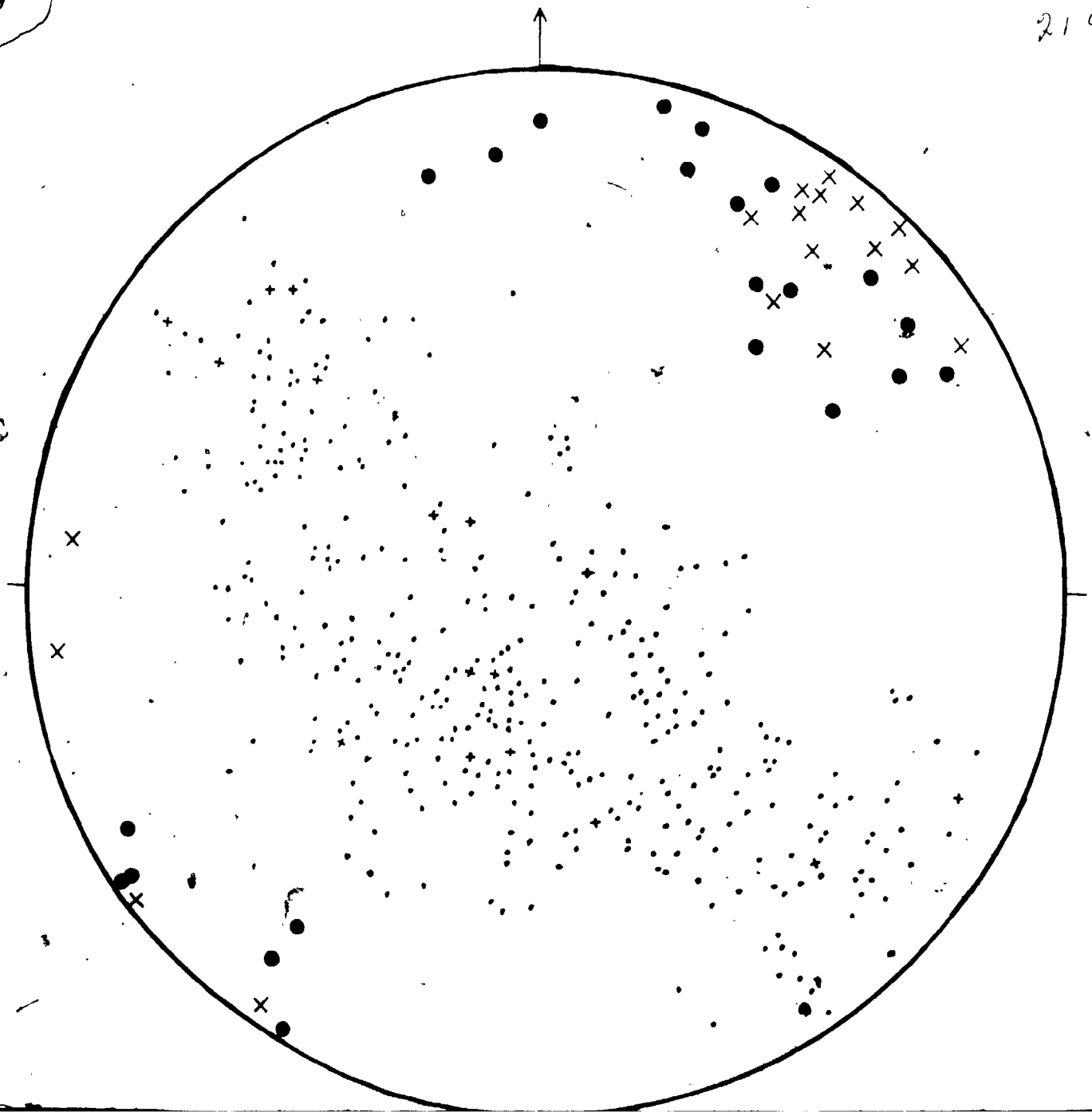

Skead
L.

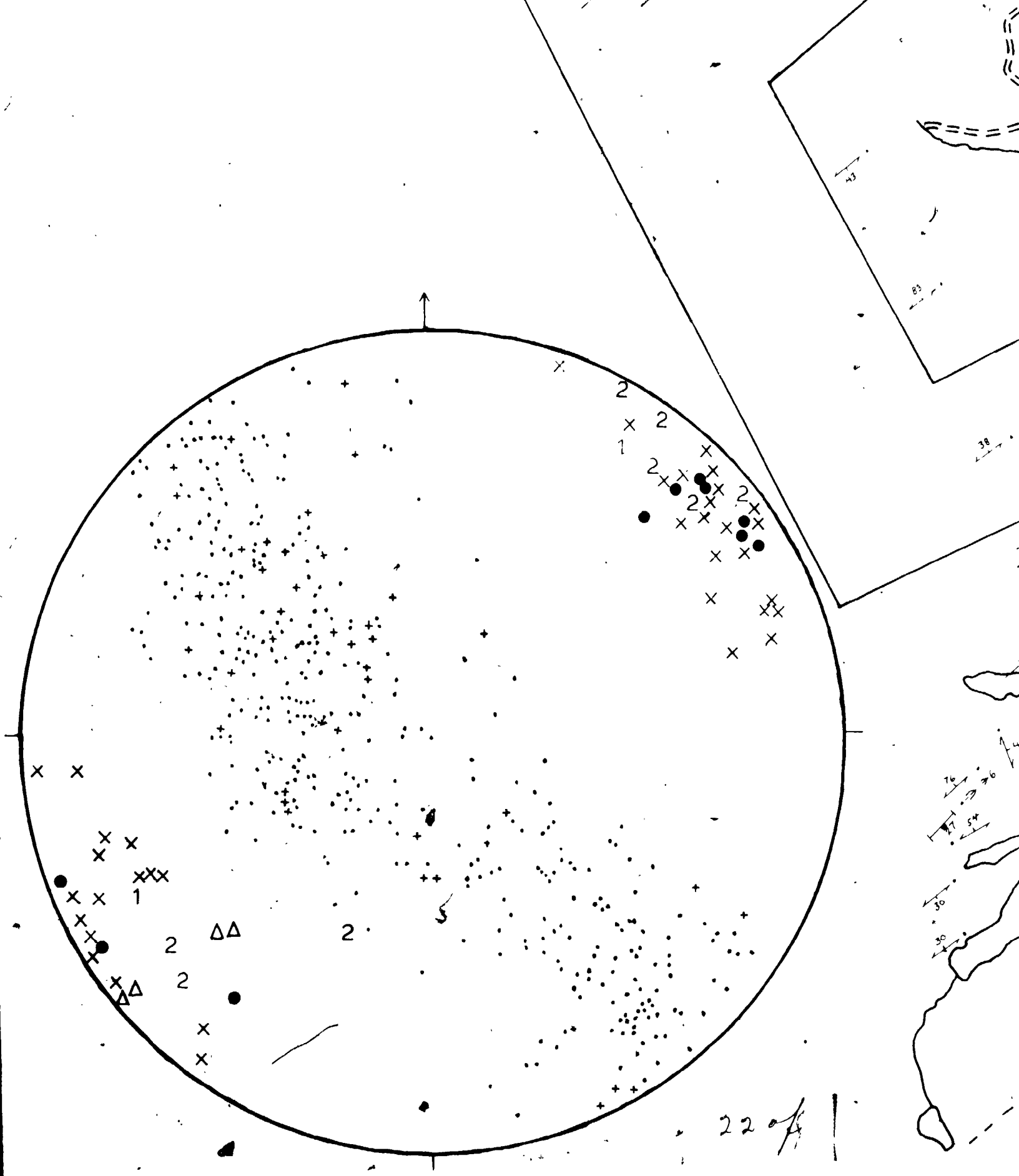

McCreary
LAKE
18 of


Straddlebug
L.
194


20 of



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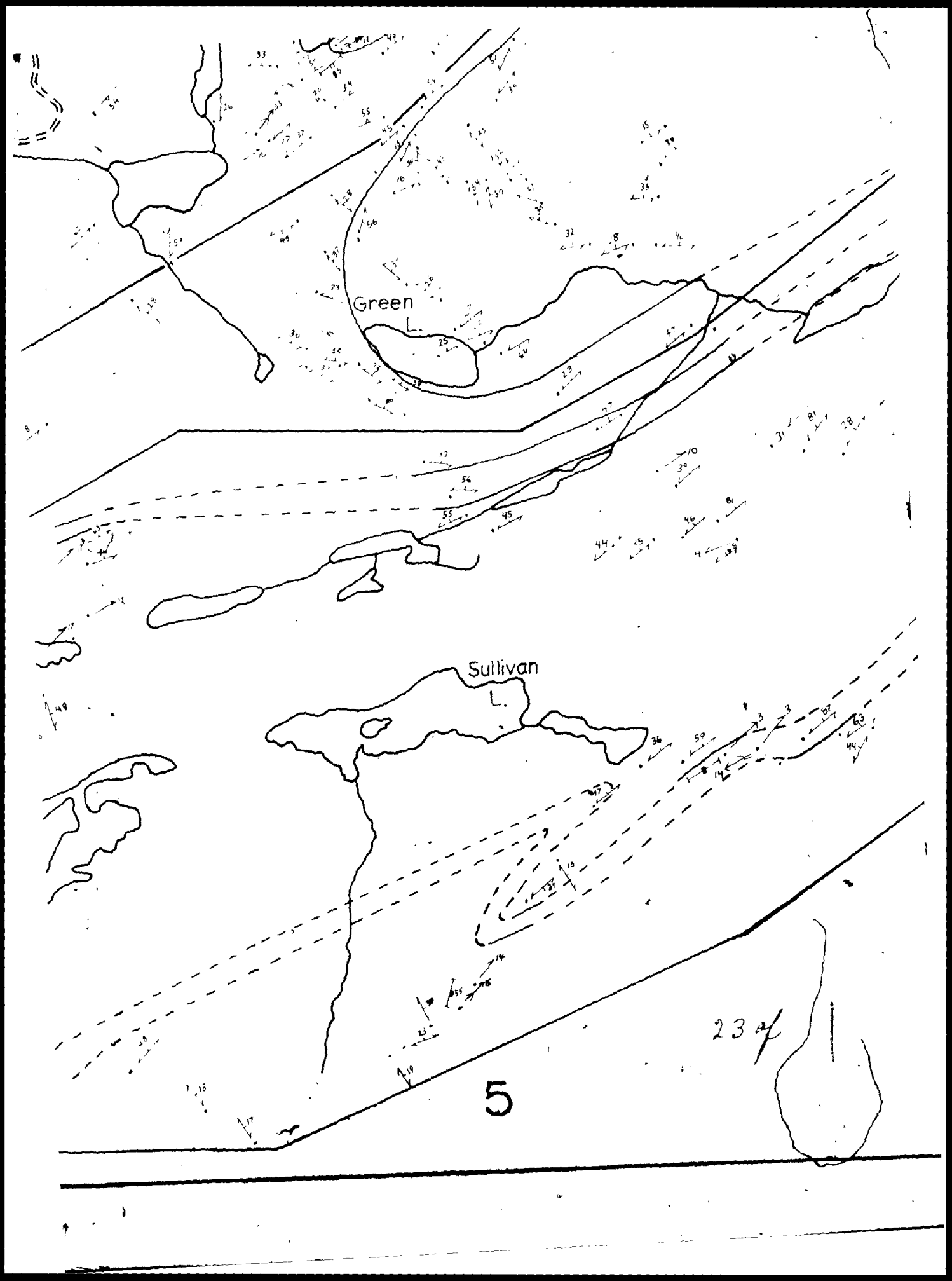
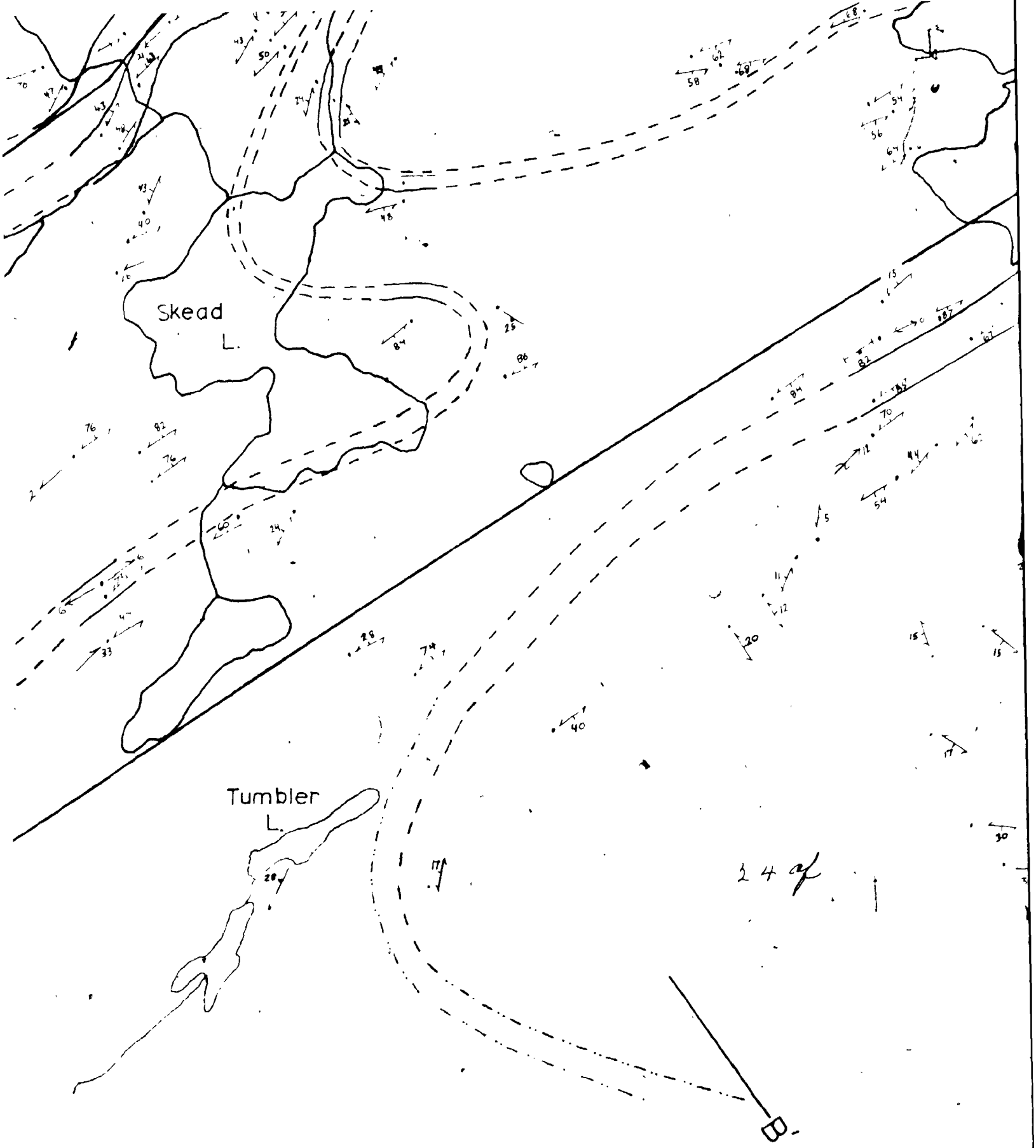
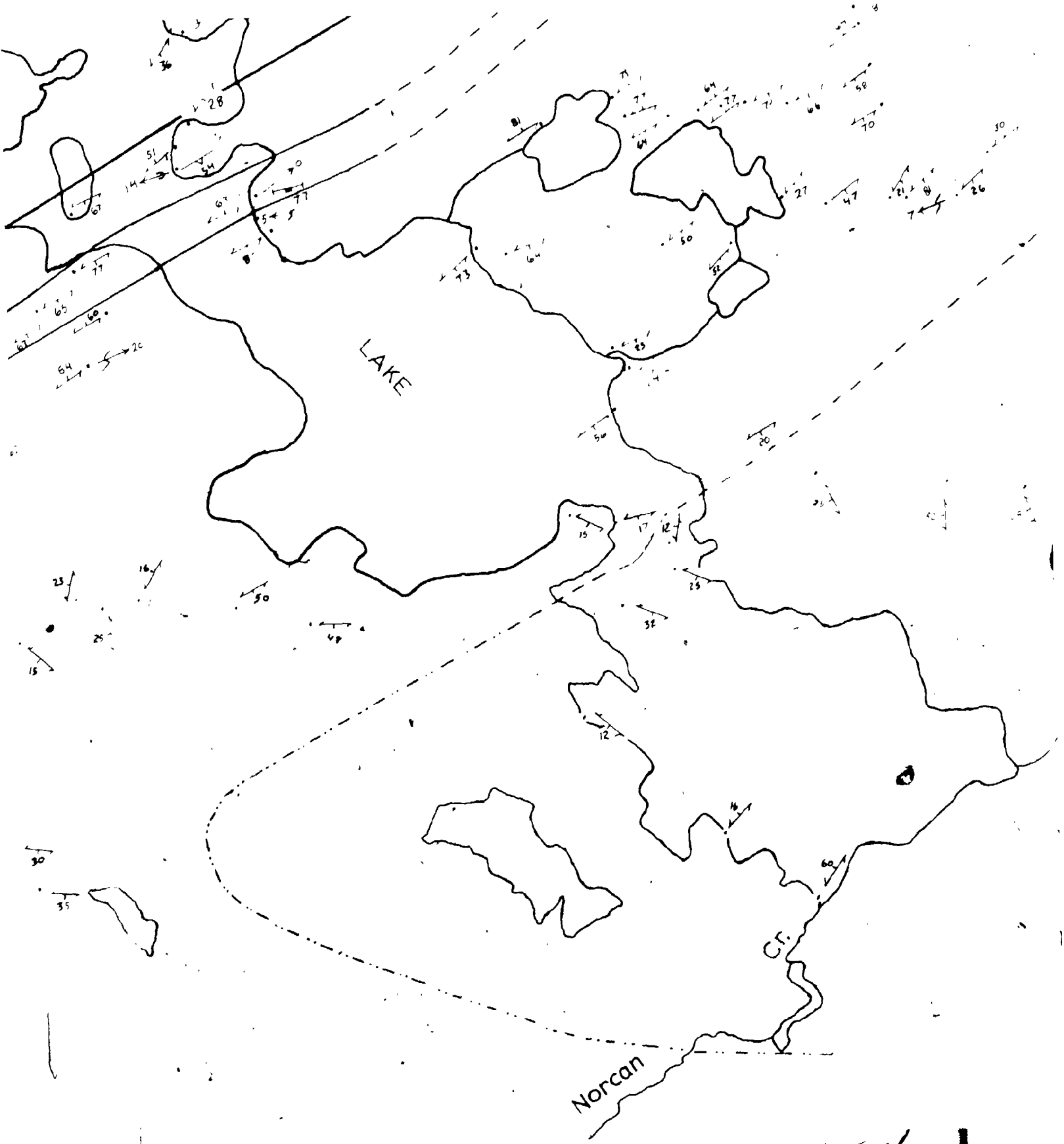

Green
L.
Sullivan
L.
5


Skead
L.
Tumbler
L.
B



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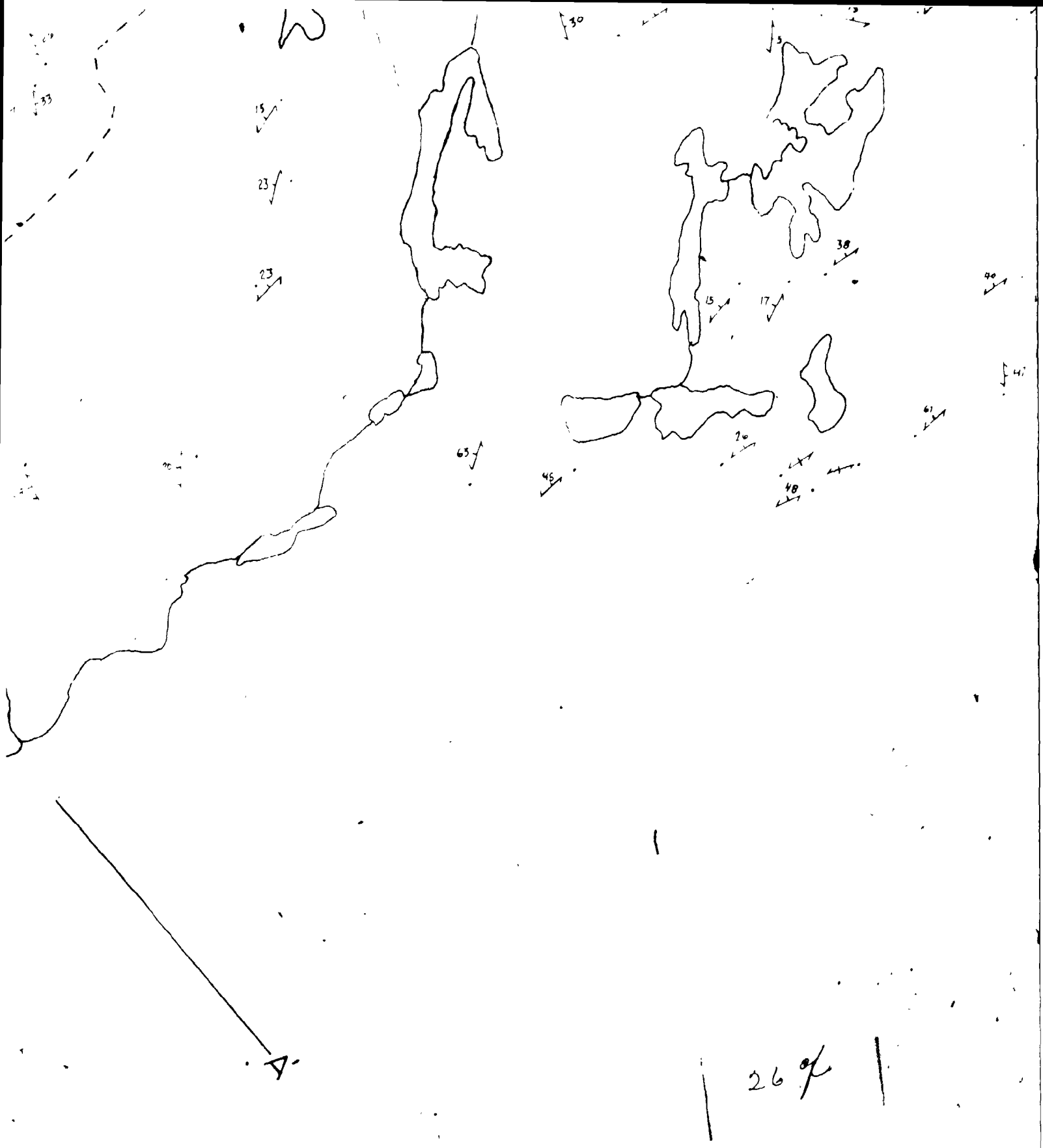
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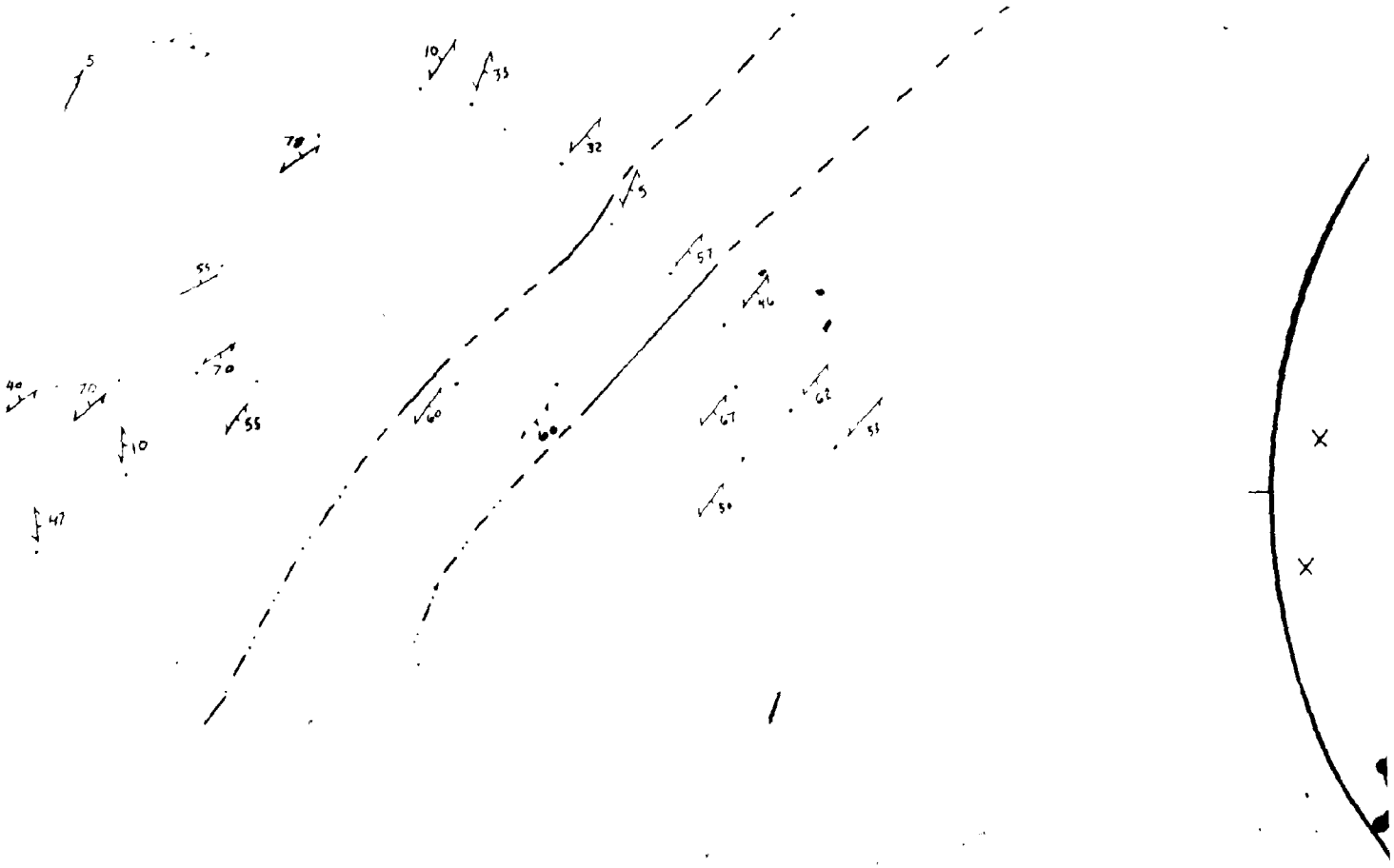

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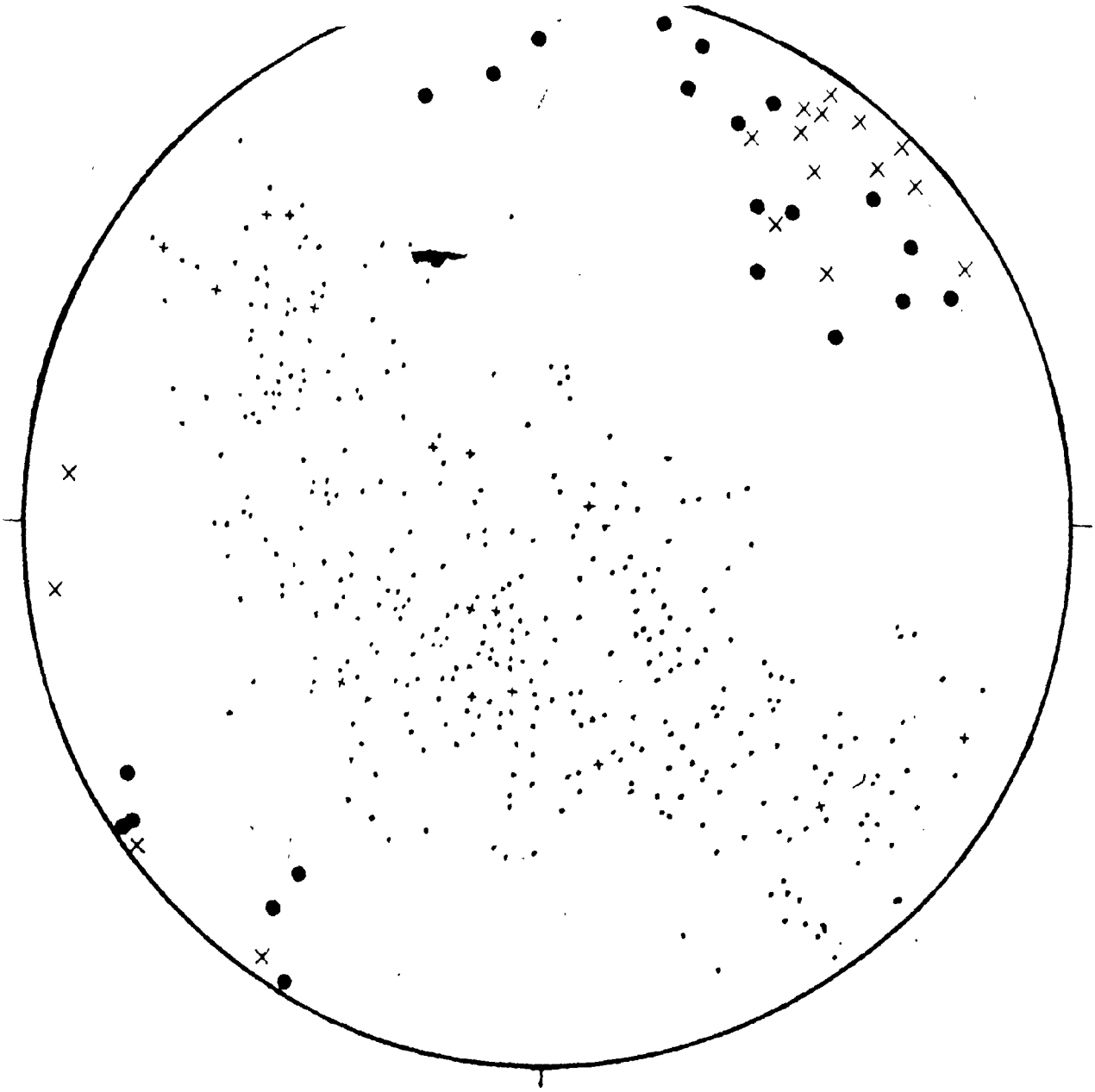


LEGEND

F1 axial surface trace (defined, assumed) showing pitch direction of major folds .....

F1 antiform, synform .....

F2 axial surface trace (defined, assumed) showing pitch direction of major folds .....

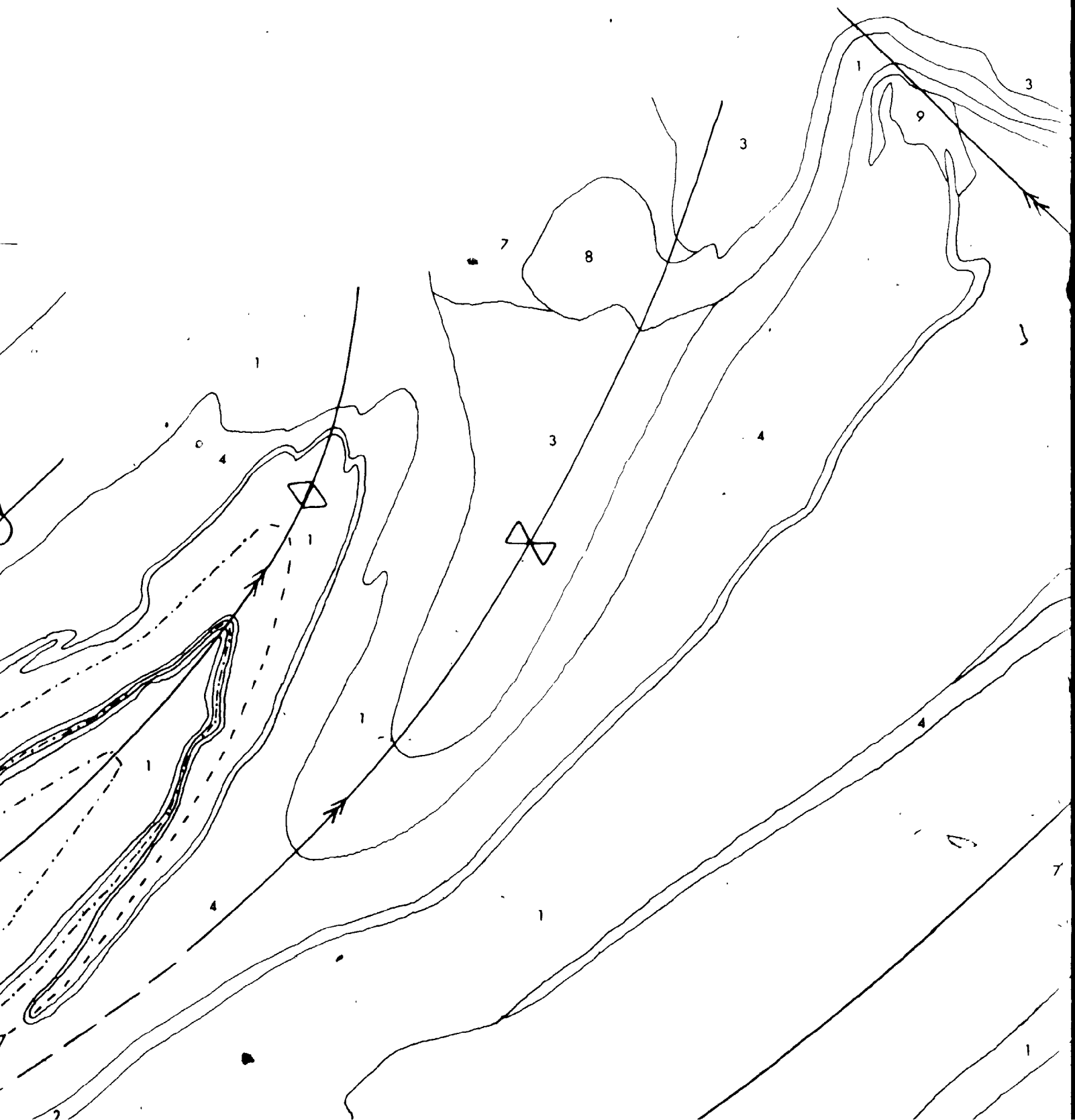
F2 upright antiform, synform .....

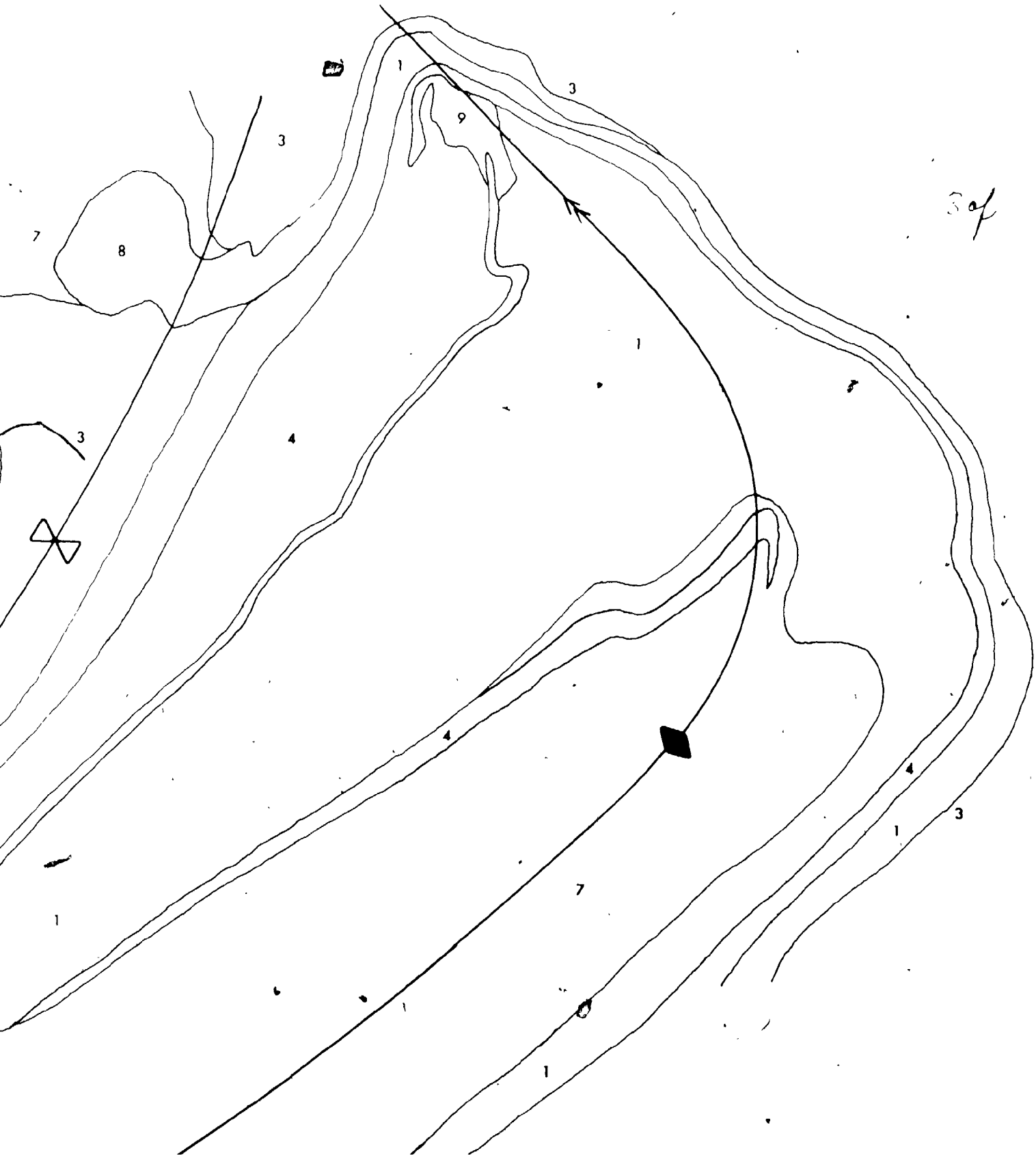
F2 overturned antiform, synform .....

F2 antiform, synform .....

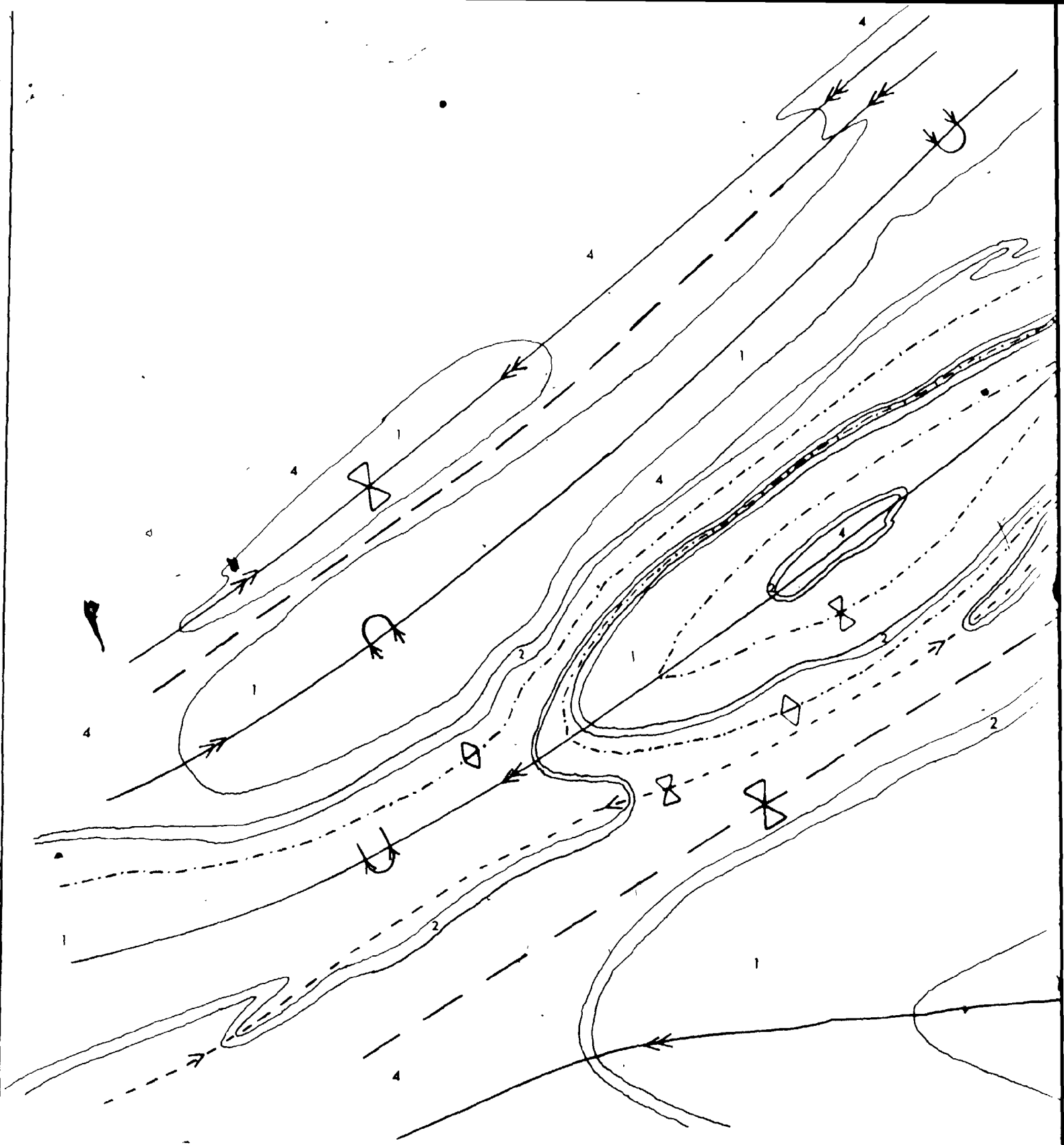
Geologic contact (units 1 to 9 designated as in figure 3) .....

2 of



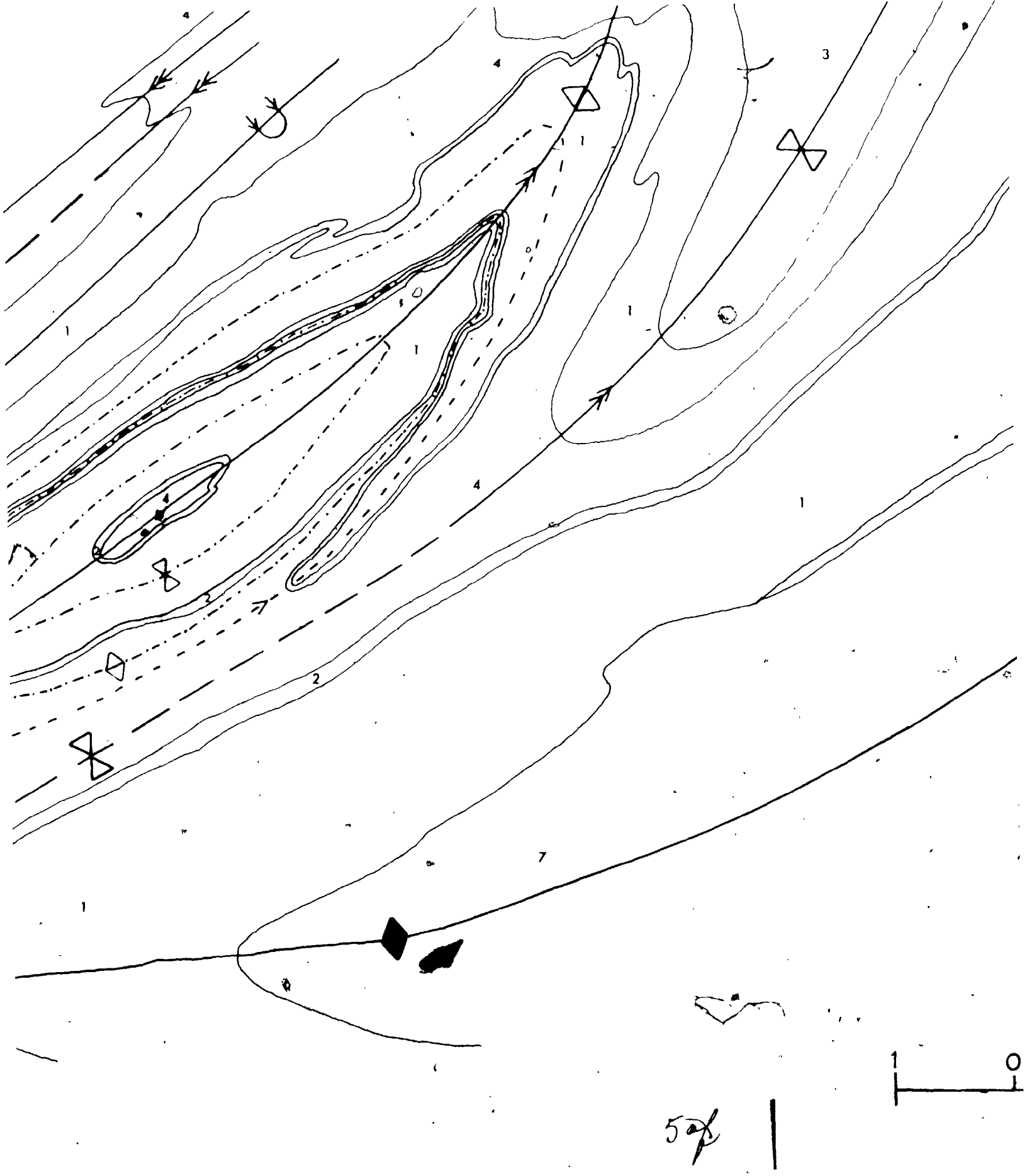


3 of

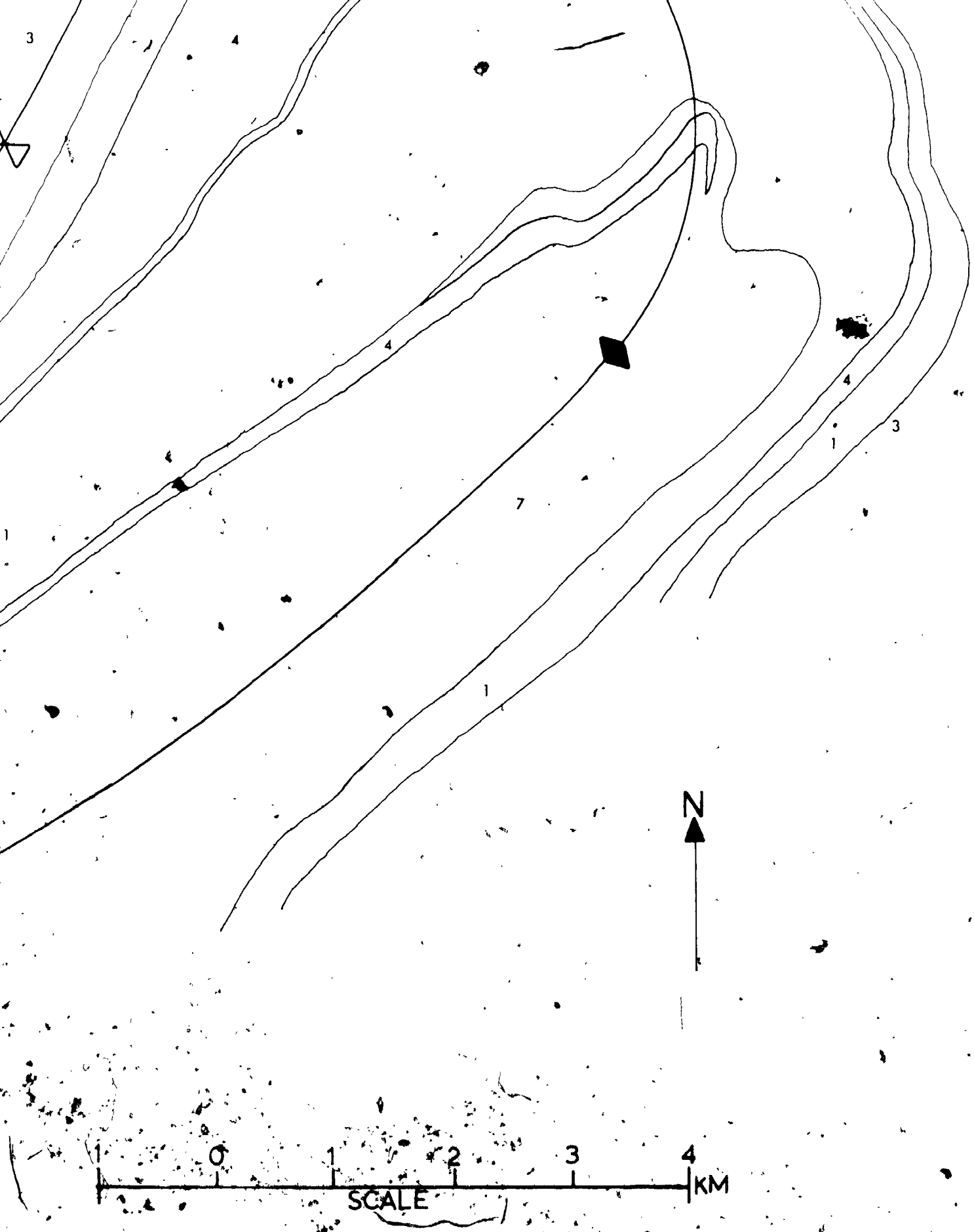


4 of 1

Axial Surface Tr



al Surface Traces of the Norcan Lake



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1 of

2 of

LEGEND

7 Tonalite

4 Para-amphibolite

3 Calcitic marble

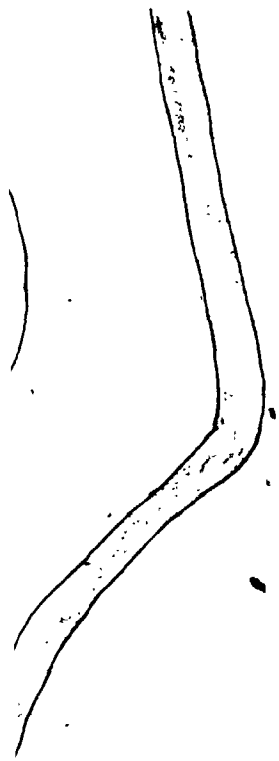
2 Amphibolitic gneiss

1 Heterogeneous gneiss

NOTES: Outcrop widths extrapolated from plan view to depth using a constant value of 20° for the plunge of the major structures

Section lines located on Figure 4

Plunge direction reverses across section lines

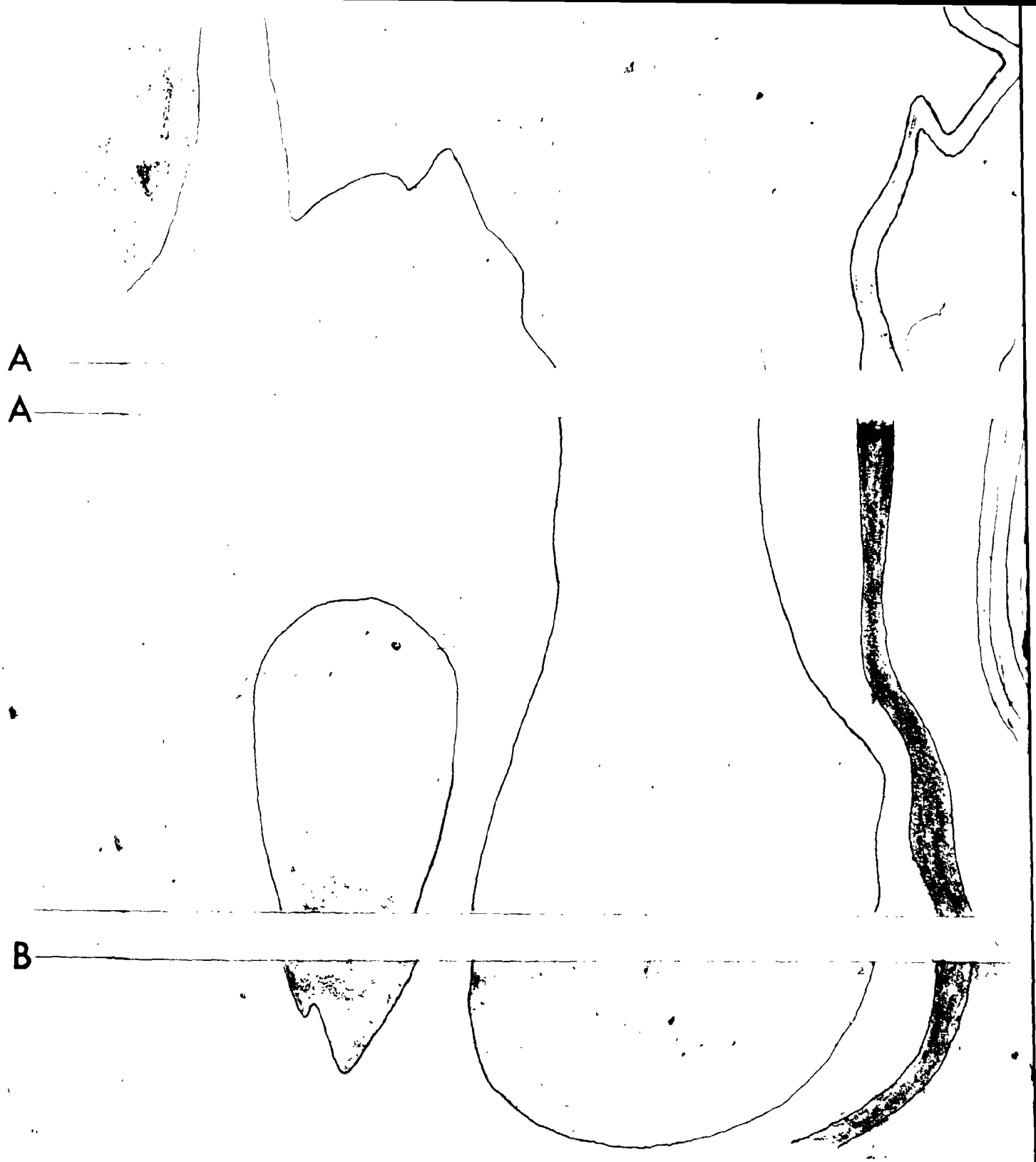


A'

A'

0 1 2

KM


A
A
B

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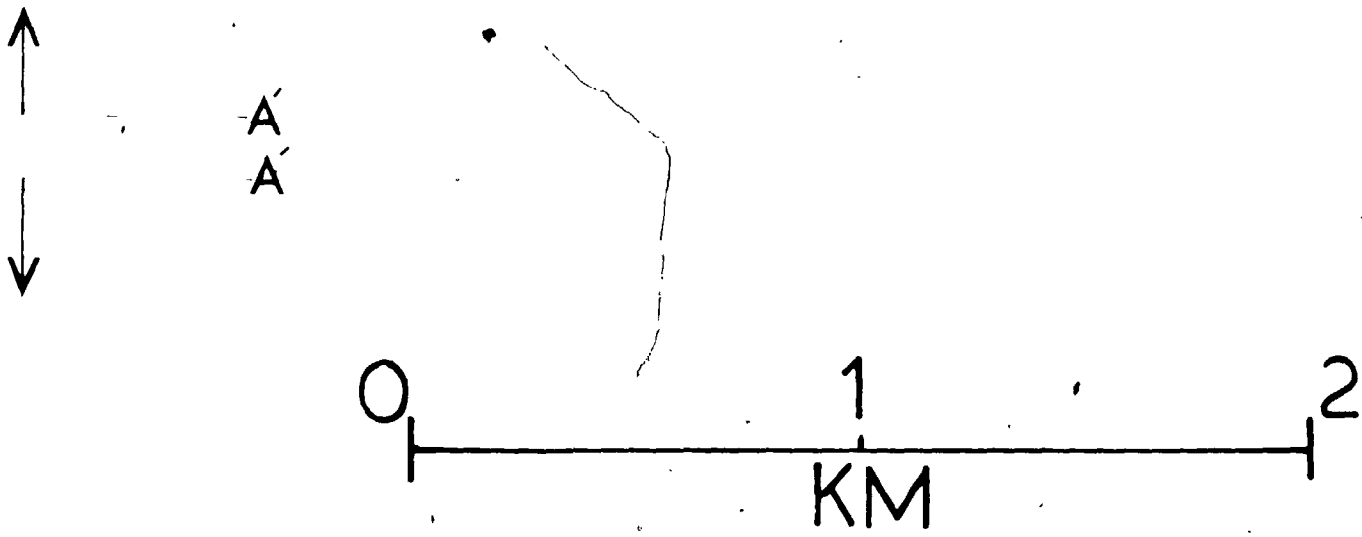
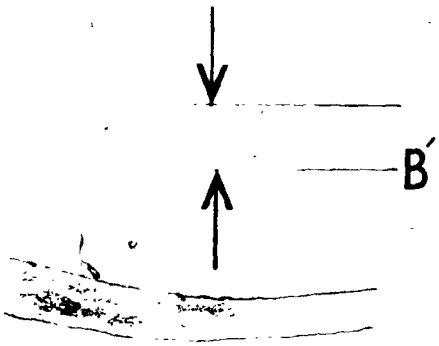
RIGHT SECTIONS

SS SECTIONS CONSTRUCTED NORMAL TO REGIONAL PLUNGE



HORIZONTAL = VERTICAL SCALE



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END

1 7 1 2 8 1

FIN