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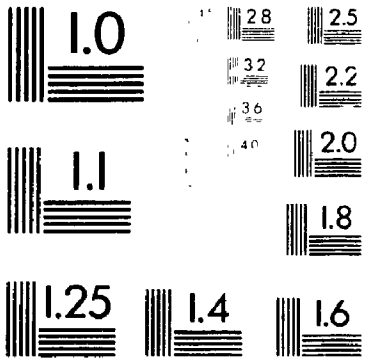

GEOLOGY OF THE WILSON ISLAND GROUP, GREAT SLAVE LAKE, NORTHWEST TERRITORIES

by

BRADFORD J. JOHNSON, B.S.

A thesis submitted to

the Faculty of Graduate Studies and Research

in partial fulfilment of

the requirements for the degree of

Master of Science

Department of Geology

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ABSTRACT

The Early Proterozoic (1.9 Ga) Wilson Island Group is exposed in a fault-bounded belt in the East Arm of Great Slave Lake, NWT. Intercalated mafic and felsic flows, volcaniclastic rocks, and fluvial to shallow-marine(?) arkose and conglomerate ("Reinhardt formation") were deposited in a tectonically active basin that probably was underlain by extending continental crust. This assemblage grades vertically into sandstone and dolostone of probable marginal- and shallow-marine origin ("Safety Cove formation"). Sandstone, argillaceous siltstone, concretionary pisolitic ironstone, and mudstone ("Basile formation") above the dolostone form an overall transgressive sequence.

North-south compressional deformation and synchronous metamorphism in the Early Proterozoic are manifested in east- to southeast-plunging folds and penetrative foliation with a down-dip stretching lineation. These structures were subsequently deformed by kink folds and dissected by transcurrent faults of the McDonald fault system.

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1. INTRODUCTION

Proterozoic rocks of the Athapuscow basin are exposed around the East Arm of Great Slave Lake, Northwest Territories, forming a northeast-trending fold belt between the Slave and Churchill structural provinces of the Canadian Shield (Fig. 1). Rocks along the shores of Great Slave Lake were first described by Bell (1902), in conjunction with his investigation of the economic mineral potential of the area in 1899. Stockwell (1933; 1936a, b) produced the first regional geological maps of the East Arm, and subdivided the supracrustal rocks into four groups: in ascending order (1) Point Lake-Wilson Island Group, (2) Union Island Group, (3) Great Slave Group, and (4) Et-then Series. The Point Lake-Wilson Island Group comprised two distinct "phases" of markedly different lithologic and metamorphic character, and of unknown relationship to one another. In later compilations, Brown (1950a, b) correlated the Point Lake phase with the Archean Yellowknife group (now called Yellowknife Supergroup), and reclassified the Wilson Island phase as a separate group. Mapping and stratigraphic investigation by Hoffman (1968) led to elevation of the Great Slave Group to its present supergroup status. Evolution of the nomenclature applied to the East Arm strata is summarized in Table 1, and the distribution of these strata is shown in Figure 2A.

The Union Island Group and the Great Slave Supergroup were deposited in Athapuscow Aulacogen (Hoffman, 1973; Hoffman *et al.*, 1974), contemporaneously with sedimentation in the Epworth and Kilohigok basins farther north (Fig. 1). The Wilson Island Group, however, is older than the aulacogen and has no known equivalents in the other

FIGURE 1. Distribution of Proterozoic basins relative to Slave (S) and Churchill (C) Provinces and Great Slave Lake (GSL). AT: Athapuscow; EP: Epworth (Wopmay Orogen); KH: Kilohigok.

	STOCKWELL, 1936a	BROWN, 1950	HOFFMAN et al., 1977	
PROTEROZOIC	ET-THEN SERIES	ET-THEN GROUP	ET-THEN GROUP	PROTEROZOIC
	GREAT SLAVE GROUP ? UNION ISLAND GROUP	GREAT SLAVE GROUP ? UNION ISLAND GROUP	GREAT SLAVE SUPER-GROUP[2]: Christie Bay Gp., Pethei Gp., Kahochella Gp., Sosan Gp. UNION ISLAND GROUP	
ARCHEAN ?	POINT LAKE-WILSON ISLAND GROUP: Point Lake phase ? Wilson Island phase	YELLOWKNIFE GROUP[1] ? WILSON ISLAND GROUP	WILSON ISLAND GROUP	ARCHEAN

Horizontal solid lines (_____) denote unconformities.

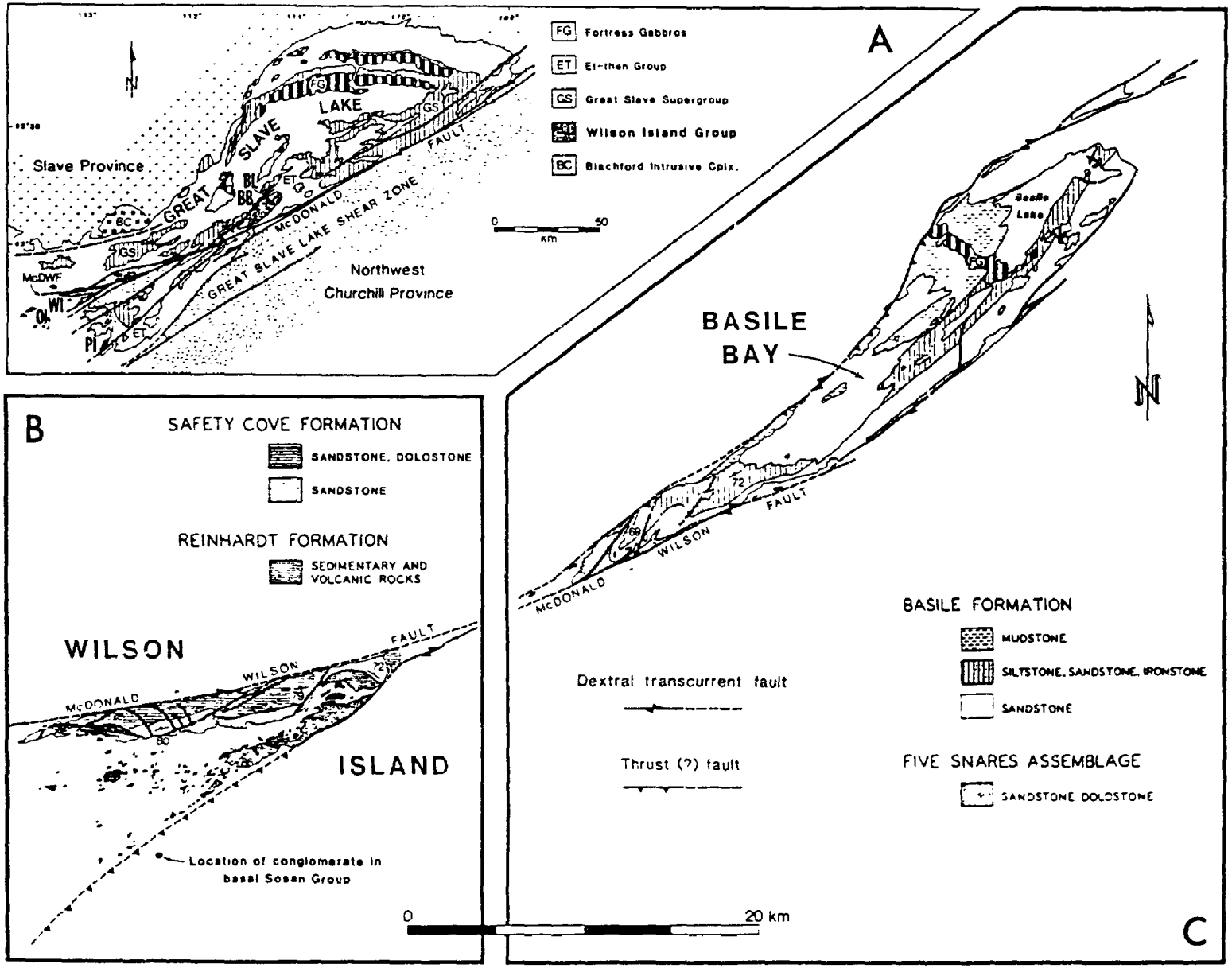
1. Classification of J.F. Henderson (1938); now called Yellowknife Supergroup (J.B. Henderson, 1970).

2. Classification of Hoffman (1968).

TABLE 1. STRATIGRAPHIC NOMENCLATURE APPLIED TO THE ROCKS OF THE EAST ARM by Stockwell (1936a), Brown (1950), and Hoffman *et al.* (1977).

FIGURE 2.

Regional geology of the East Arm of Great Slave Lake (**A**), showing distribution of the Wilson Island Group. **BB**: Basile Bay; **BL**: Basile Lake; **OI**: Outpost Islands; **PI**: Petitot Islands; **WI**: Wilson Island; **McDWF**: McDonald-Wilson Fault. Adapted from Hoffman, 1985. Generalized geology of the Wilson Island area (**B**) and Basile Bay area (**C**), based on the present study. Following Hoffman (1985), some faults are labelled with both thrust and transcurrent fault symbols, to indicate that they probably have been loci of both types of movement.



basins (Hoffman, 1973).

The Wilson Island Group, a succession of metasedimentary and metavolcanic rocks at least 8 kilometres thick, is distinguished from the younger Proterozoic successions in the East Arm by its ubiquitous penetrative deformation and greenschist to amphibolite facies metamorphic grade. Together with small intrusions known collectively as the Butte Granite, the Wilson Island Group is preserved as a fault-bounded sliver called the Wilson Island Terrane (Hoffman, 1985), that extends from the Outpost Islands northeastward to Basile Lake (Fig. 2). Abutting the Wilson Island Terrane to the south is a block of predominantly granitoid rocks and gneisses which Hoffman (op cit.) has termed the Simpson Islands Terrane; to the north are nappes of Great Slave Supergroup rocks that verge northwestward toward the Slave autochthon (Hoffman et al., 1977; Hoffman, 1985). The northeast trend of the Wilson Island Terrane, and of the East Arm fold belt as a whole, is parallel to the Great Slave Lake Shear Zone (Hanmer and Lucas, 1985) and the McDonald Fault system.

The minimum age of the Wilson Island Group is based on a U-Pb zircon date of 1928 ± 11 Ma (Bowring et al., 1984), obtained from samples of a felsic porphyry within the lower part of the exposed succession. Except where it is overlain with angular unconformity by the Murky Formation of the Et-then Group, stratigraphic contacts between the Wilson Island Group and other rocks are nowhere exposed. However, clasts of strongly deformed metasedimentary rocks of probable Wilson Island Group provenance occur within a conglomerate near the base of the Sosan Group (lowermost Great Slave Supergroup), suggesting that deformation and erosional unroofing of the Wilson Island Group preceded deposition of the Great Slave Supergroup (Hoffman et al.,

1977). This conglomerate is exposed on an island 6.5 kilometres south of Wilson Island (Fig. 2B).

Mapping of rocks in the western part of the East Arm, including the Wilson Island area, was carried out by Reinhardt (1969). Sedimentological aspects of the Wilson Island Group, focusing on paleocurrents and geochemistry of the sandstones, were reported by Yeo (1976a, b). Mapping of the Proterozoic rocks of the East Arm at 1:50,000 scale was completed by Hoffman et al. (1977; Hoffman, 1977). The present investigation is based on detailed mapping during 1984 and 1985, with emphasis on the description and analysis of the stratigraphy and structure of the Wilson Island Group.

2. STRATIGRAPHY

Incomplete stratigraphic successions of the Wilson Island Group are well exposed in the Wilson Island area (Map 1) and on the mainland around Basile Bay (Map 2). The rocks at Basile Bay have been displaced northeastward relative to those on Wilson Island, along the dextral McDonald-Wilson Fault (Hoffman et al., 1977). In both areas the strata generally are steeply inclined, with tops to the northwest. Rocks at the top of the section on Wilson Island lithologically resemble those in the lower part of the section at Basile Bay. The stratigraphic packages at Wilson Island and Basile Bay therefore appear to represent, respectively, the lower and upper parts of a continuous succession.

The rocks of both Wilson Island and Basile Bay have been metamorphosed to the greenschist facies, but in the following stratigraphic descriptions they are discussed in terms of their pre-metamorphic protoliths. Thus, quartzites are called sandstones, slates are called mudstones, and so forth. In the Outpost Islands, rocks of the Wilson Island Group have been metamorphosed to the amphibolite facies. The stratigraphic position of these rocks relative to those in the other areas is uncertain.

Provisional names for three formations and for one informal assemblage have been assigned to the strata at Wilson Island and Basile Bay. In ascending order (based on the correlation outlined above) these are the Reinhardt and Safety Cove formations, the Five Snares assemblage, and the Basile formation (Table 2). A "turbidite" unit (Hoffman, 1977), exposed on islands northeast of Wilson Island, may in part be correlative with the Basile formation (Hoffman et al., 1977).

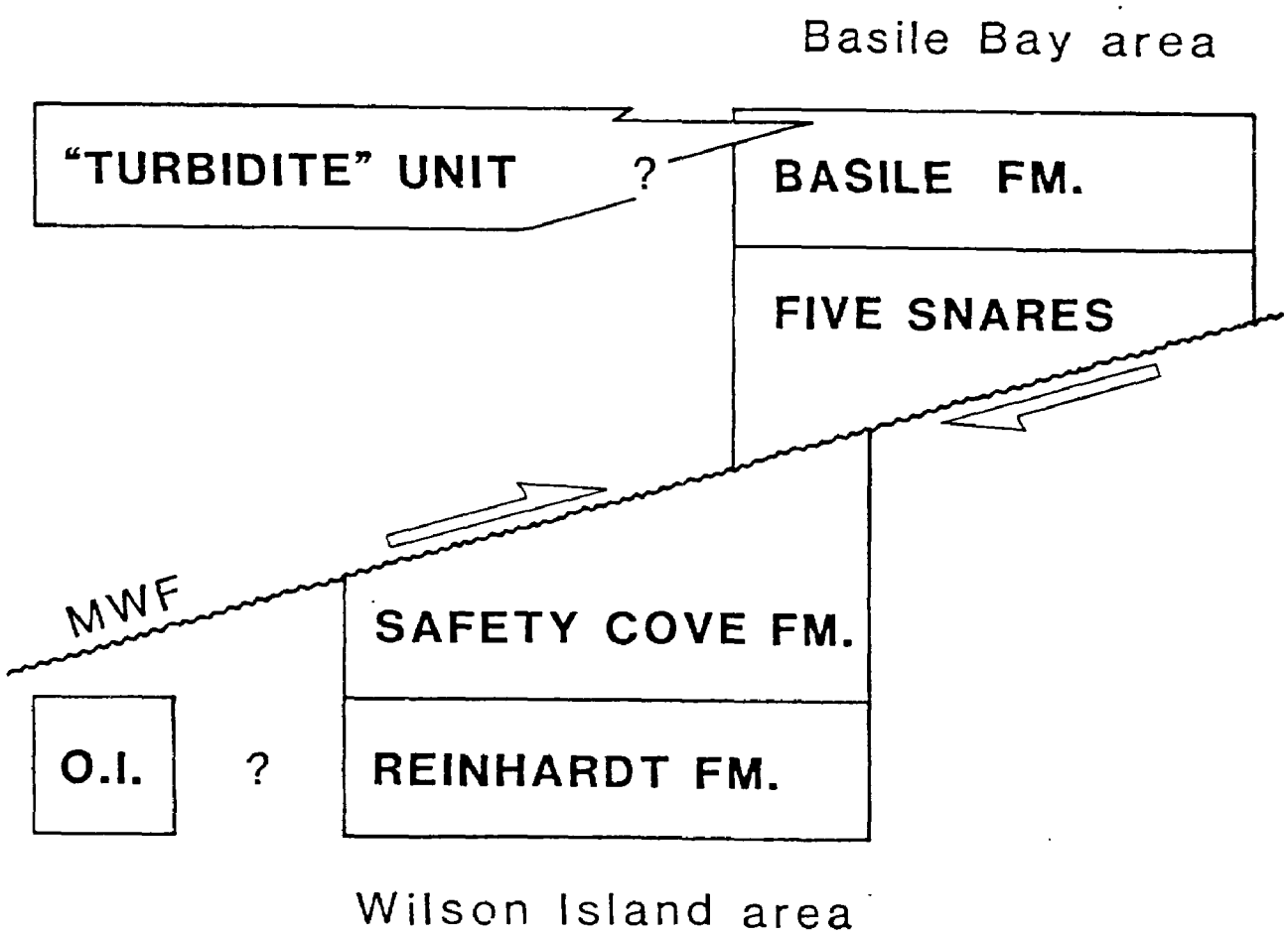


TABLE 2. INTERNAL STRATIGRAPHIC RELATIONS OF THE WILSON ISLAND GROUP.

MWF: McDonald-Wilson Fault. O.I.: Outpost Islands

REINHARDT FORMATION

An assemblage of intercalated mafic and felsic metavolcanic rocks and clastic metasedimentary rocks, exposed on the southern part of Wilson Island and on the "Reinhardt islands" (a group of islands southwest of Wilson Island, named in honour of the late E. W. Reinhardt), is here informally named the Reinhardt formation. Two members and four informal units (Table 3) are named on the basis of lithology and stratigraphic relationships to distinctive volcanic units (Figure A). Stratigraphic terminology is assigned to the volcanic rocks where applicable.

LOWER IGNEOUS COMPLEX

The Lower Igneous Complex is exposed along the southern shoreline of Wilson Island and on the southeasternmost Reinhardt Islands. The basal part of this unit is truncated by the Inconnu Thrust.

Basalt flows form about 70 percent of the Lower Igneous Complex. The flows are massive, and commonly have autobrecciated tops. They are intruded by fine-grained diabase sills that may be genetically related. Amygdules for the most part are evenly dispersed, but are concentrated near the tops of the flows, and occur as local concentrations in the central parts. Mineralogically, the basalts are composed of felted needles of actinolite (typically 50-60% of the rock volume), euhedral to anhedral interstitial plagioclase (albite to oligoclase; up to 30%), granular masses of epidote, disseminated opaque minerals (mainly pyrite and (?)magnetite; up to 10%), and minor quartz and carbonate. Many of the amygdules are concentrically zoned, consisting of, from rim to core: quartz, intergrowths of epidote/chlorite/actinolite, and albite. The epidote commonly forms bladed radial structures. Carbonate, mag-

netite or pyrite, and hematite are disseminated in the amygdules.

Felsic porphyritic intrusions are abundant in the Lower Igneous Complex. A unique variety, exposed only on the southernmost Reinhardt Islands, contains phenocrysts of microcline (?) up to 10 millimetres across. Other porphyritic intrusions resemble their extrusive equivalents higher in the section.

The felsic porphyries are cut by black, aphyric, mafic dikes. These commonly are schistose and tongue-shaped, up to a few metres wide, with irregular margins and diffuse to sharp contacts with the felsic country rock. Partially digested enclaves of felsic porphyry occur near the margins of these dikes, and mafic inclusions, compositionally identical to the dikes, occur within the felsic rock. A thin section from one of these mafic dikes shows the following modal mineralogy: biotite (40%), epidote (20%), actinolite (15%), quartz (20%), plagioclase (less than 5%), garnet (1%), sericite, and opaques.

Clastic sedimentary rocks within the assemblage include biotite-rich silty mudstone, less common fine- to coarse-grained sandstone, and rare graded lapilli tuff beds about 15 centimetres thick, which collectively form less than 10 percent of the complex. The silty mudstones generally are parallel laminated (rarely ripple laminated), and display abundant load and flame structures and convolute bedding. Trough crossbeds are common in the arenites. The clastic units form discontinuous beds that generally are less than 10 metres thick.

LOWER CLASTIC MEMBER

The Lower Clastic Member consists of nearly 600 metres of medium-grained feldspathic sandstone, argillaceous to sandy siltstone, basalt

flows, and sandy lapilli-ash tuff. Two submembers are separated by a thick composite basalt flow unit:

Submember 1

The base of the member is marked by a bed of poorly sorted lithic granule arkose that overlies intercalated siltstone and mafic metapelite at the top of the Lower Igneous Complex. This bed is 2-3 metres thick. It contains abundant quartz granules and pebbles, highly altered plagioclase and potassium feldspar, metavolcanic and metasedimentary lithoclasts, and metamorphic biotite and chlorite.

Medium-grained feldspathic sandstones (mostly arkosic arenites) form 70 percent of Submember 1. These arenites weather white to orange-pink and display trough crossbedding, parallel and ripple laminations, and both normal and reverse graded bedding. The arenite beds typically are tabular, and are separated by intercalated thin beds and laminations of fine-grained sandstone and argillaceous siltstone. These fine-grained units weather light brown to grey and display ripple lamination and (less common) lenticular bedding.

In the upper part of Submember 1, arenites are complexly intercalated with sandy lapilli-ash tuffs, which weather grey to brown, are poorly sorted, and form massive, unstratified units. The lapilli-ash tuffs (cf. Schmid, 1981) contain coarse sand-sized grains of plagioclase (some angular), coarsely polycrystalline quartz (typically rounded), potassium feldspar, and lapilli-sized irregular fragments of slightly quartzose and locally chloritic biotite schist (Fig. 3A). Subangular pebbles of fine-grained sandstone and angular cobbles or blocks of amygdaloidal basalt are less common. Clasts as large as 8 centimetres locally are abundant, in which case the rock is termed a breccia-tuff, in keeping with the nomenclature of Schmid (op cit.).

The clasts are supported by a fine-grained to microcrystalline matrix of quartz, feldspar, biotite, and chlorite, with minor white mica, epidote, actinolite, sphene, zircon, and garnet.

The upper surfaces of tuffaceous units are characterized by flame structures that protrude into overlying arenite beds (Fig. 3A); the basal parts of such arenite beds contain angular clasts of biotite schist from the tuffs. The tuffaceous units of Submember 1 locally contain lenticular arenite intercalations and isolated enclaves of deformed arenite, up to 4 metres across, that are internally well stratified and display a convex-downward orientation. They also locally contain irregular-shaped clasts of sandy lapilli-ash tuff (Fig. 3B).

Submember 2

An unstratified sandy lapilli-ash tuff at least 25 metres thick overlies the basalt that separates Submembers 1 and 2. The lapilli-ash tuff unit is in turn overlain by a succession of arkosic arenites and intercalated siltstones, much like those of Submember 1, which together constitute about 75 percent of Submember 2. These units contain numerous small-scale fining-upward cycles.

Typical fining-upward cycles consist of medium- to coarse-grained sandstone (35-100 centimetres thick), gradationally overlain by units of interlaminated fine-grained sandstone and argillaceous siltstone (less than 30 centimetres thick), in which the amount of sandstone decreases upward. Sandstones in the lower beds of these cycles display trough crossbedding, and both normal and reverse graded bedding. The upper parts of some fining-upward cycles display a gradation from small-scale trough crossbedding or current-ripple lamination (commonly with reactivation surfaces) to planar lamination. In other cycles,

recurrent graded laminae of fine-grained sandstone and argillaceous siltstone give way upward to ripple-laminated siltstone units in which sandy layers are absent. Interfaces between cycles commonly are slightly erosional, and are characterized by ball-and-pillow structures, load and flame structures, and small-scale slump folds.

Trough-crossbedded arenites in about the middle of the submember contain abundant scour-based pebble lags. The pebble types include light brown to greenish grey siltstone, grey to black intermediate or mafic metavolcanic rocks, quartz, and fine-grained granitoid rocks.

Based on petrographic analyses of six thin sections, the arenites of the Lower Clastic member typically are moderately sorted, and are composed of quartz (subrounded; 40-50% of rock volume), potassium feldspar (mostly orthoclase and perthite, some microcline; subrounded to subangular; 10-20%), plagioclase (subangular to subrounded, less commonly angular; 15-35%), lithic fragments (mostly chert or recrystallized felsic volcanic rock, less common metasedimentary and mafic or intermediate metavolcanic rock; up to 15%), and intergranular metamorphic chlorite and biotite (10-15%). They are thus texturally submature and compositionally immature to submature. Adjacent quartz grains in some samples exhibit sutured contacts, which obscures their degree of roundness. Most of the quartz is coarsely polycrystalline (i.e., much coarser than chert; Fig. 3C) and displays sutured subgrain boundaries. The feldspars are slightly altered to sericite and show thin coatings of hematite. Some feldspars exhibit feldspar overgrowths. Detrital muscovite, zircon, garnet, and epidote occur in small quantities. The argillaceous rocks of the assemblage contain up to about 45 percent metamorphic biotite.

Numerous thin, laterally continuous, amygdaloidal basalt flows

that occur in the Lower Clastic member differ from those of the Lower Igneous Complex only in that the amygdules of the former more commonly exceed 2 centimetres in size.

PEBBLY SANDSTONE MEMBER

Tabular beds of arenite at the top of the Lower Clastic member are overlain with planar contact by a poorly sorted massive conglomerate, about 10 metres thick, that marks the base of the Pebbly Sandstone member. The basal conglomerate grades into trough-crossbedded arenite with intercalated fine-grained sedimentary units like those of the Lower Clastic member. Pebbly layers up to 30 centimetres thick commonly form lags in channels at the bases of small-scale fining-upward sequences (Fig. 3D). Pebbles within these layers commonly are imbricated. The predominant pebble types (in approximate decreasing order of abundance) are dark grey to black intermediate or mafic metavolcanic rocks, quartz, brown to grey siltstone, and both foliated and non-foliated granitoid rocks. Clasts of orthoquartzite and red felsic porphyry (petrographically identical to porphyries of the Reinhardt formation) are much less common. The intermediate volcanic clasts and some of the siltstone clasts are angular to subrounded, whereas other clasts typically are well-rounded. Clasts up to 8 centimetres are common. The largest cobbles, over 15 centimetres across, are of granitoid rocks and porphyry.

Thin sections show that the dark-coloured metavolcanic clasts are composed of plagioclase microlites in a fine-grained groundmass of chlorite or biotite and disseminated pyrite. The groundmass of some metavolcanic clasts additionally contains sericite, quartz, and epi-

FIGURE 3.

A. Lapilli-ash tuff, showing irregular clasts of biotite schist. Flame structures protrude into overlying arenite bed. Diameter of coin in lower right (and in other photographs) is 2 cm.

B. Irregular-shaped clast of lapilli-ash tuff in matrix of lighter-coloured lapilli-ash tuff. Pen is 15 cm long.

C. Photomicrograph of feldspathic arenite from Lower Clastic member, showing grains of plagioclase (twinned), minor potassium feldspar, and coarsely polycrystalline quartz (e.g., large grain, lower right) in a poorly sorted sandy matrix.

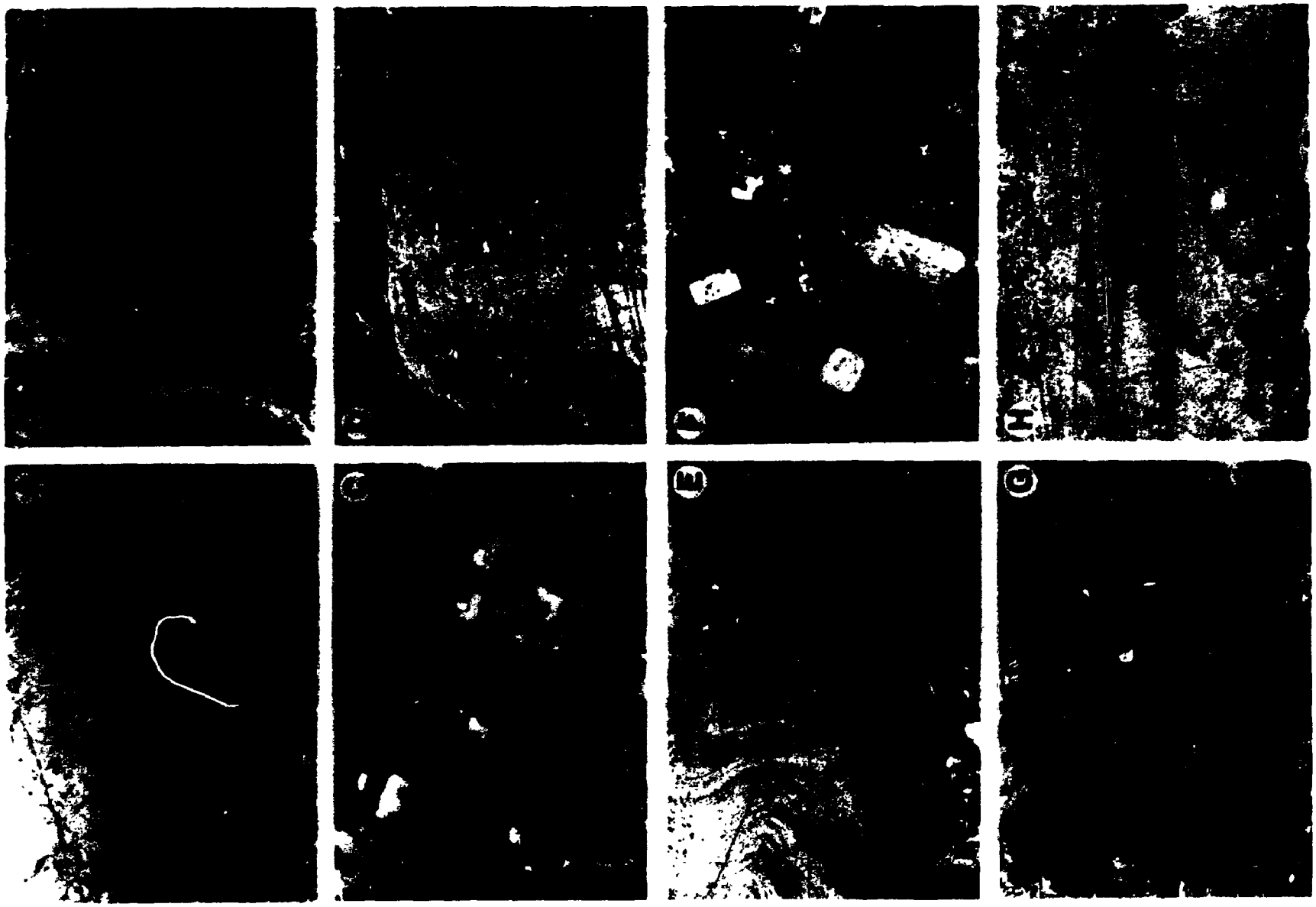
D. Trough crossbeds and scoured channels with pebble lags in feldspathic arenite of Reinhardt formation. Top to right; hammer for scale.

E. Flow banding in felsic volcanic rock, Reinhardt islands.

F. Photomicrograph of felsic porphyry, displaying zoned and twinned plagioclase phenocrysts. Cross-polarized light. Scale bar = 1 mm.

G. Felsic dike with mafic inclusions. Mafic margin of dike is chilled against basalt (left). Lower Igneous Complex.

H. Gently dipping cross strata in quartzose sandstone, accentuated by heavy mineral laminae. Lower Arkose member, Safety Cove formation.



dote.

Two amygdaloidal basalt flows are intercalated with the sedimentary units of the Pebbly Sandstone member (Figure A). On Map 1A, these flows are shown only in the hinge area of the Blind Bay anticline, where the combined effects of topography and the low dip of strata gives them a significant apparent thickness. The Pebbly Sandstone member is capped by a thick basalt unit.

FELSIC PORPHYRITIC ROCKS

Thick layers of felsic porphyry occur at numerous stratigraphic levels in the succession. The layers have concordant basal contacts and display lateral thinning. They are connected to a system of feeder dikes that is well exposed across the southern part of Wilson Island. A laterally extensive 20-metre-thick porphyry sill that intrudes the Lower Clastic member is also part of this system. Thick felsic units dominate the limited exposure in the western Reinhardt islands. The felsic rocks range from purple to pink or brick red; in zones of high strain they are light green. They commonly are flow banded (Fig. 3E), and display autoclastic breccias and quartz-filled amygdules near their upper margins. In some localities on the Reinhardt islands, the basal 2 metres of felsic units are fragmental.

Plagioclase phenocrysts, almost invariably present, commonly form up to 20 percent of the rock. Euhedral laths up to 6 millimetres long display Carlsbad, albite, and pericline twinning. The plagioclase consistently appears to be albitic, based on numerous petrographic measurements of extinction angles using the Michel-Levy method. Other feldspar phenocrysts are tabular to ovoid in outline and exhibit only Carlsbad twinning. In some samples these are clearly perthitic potas-

sium feldspar; in others they are optically positive, and thus may be albite or albitized potassium feldspar. Zoning is especially common in these ovoid feldspars, although some crystals of polysynthetically twinned plagioclase also are zoned (Fig. 3F). Some feldspars also display myrmekitic textures.

The purple felsic units contain biotite, pyrite, magnetite(?), chlorite, and epidote, typically in association with aggregates of plagioclase phenocrysts. The quartzo-feldspathic groundmass is aphanitic to fine-grained. The more siliceous pink units consist of alternate microcrystalline quartzo-feldspathic and sericitic layers, with phenocrysts of quartz and albite (or albitized potassium feldspar). Flow banding and spherulites are especially common in this variety, which probably is rhyolite. The units in which ferromagnesian and calcic minerals abound are probably close to dacite in composition.

Amphibole phenocrysts are characteristic of some felsic intrusive units of the Lower Igneous Complex, locally forming over 25 percent of the rock. Thin sections of two amphibole-phyric rocks were studied. In one sample the amphibole is a bright green pleochroic hornblende (maximum extinction angle = 24°), which has been extensively replaced by epidote, silica, and chlorite. The other sample displays chlorite pseudomorphs after prismatic amphibole, in turn partially replaced by silica. Feldspars in these rocks are highly altered, and are embayed by microcrystalline silica.

Dark grey plagioclase-phyric intrusions are spatially associated with siliceous porphyry dikes, commonly separating the latter from metasedimentary country rocks. Contacts between the two igneous phases range from gradational to sharp (and generally sheared). In addition

to sparse concentrations of plagioclase phenocrysts, the grey units contain actinolite, biotite, epidote, quartz, and opaques.

Some of the porphyry dikes are compositionally zoned. An example within the Lower Igneous Complex is well-exposed in the Reinhardt islands (Fig. 3G). This dike displays aphyric mafic border zones about 10 centimetres wide, with chilled outer margins. The mafic border phase shows distinct (but not chilled) contacts with purple biotite-rich porphyry, which grades toward the centre of the dike into siliceous pink porphyry. Mafic inclusions are dispersed in the porphyry, with decreasing abundance and increasing size away from the margins. The inclusions are mineralogically identical both to the mafic border phase of the dike and, as confirmed by a thin section, to the aphyric mafic dikes that intrude felsic porphyries in the Lower Igneous Complex (i.e., actinolite, biotite, epidote, quartz, garnet). Such mafic inclusions are common in intrusive felsic units throughout the assemblage, and are especially abundant in the Lower Igneous Complex. They typically are a few millimetres long, although the largest inclusions are over 100 centimetres long and 15-20 centimetres wide. A similar zonation is exhibited by composite units within the l[illegible] thick felsic volcanic sheet on Wilson Island, especially near the western end of the main system of feeder dikes. In this case (Fig. 4), each unit has only one chilled mafic margin.

Disseminated sulfides occur in some small zones where the felsic rocks are strongly sheared. Pyrite is concentrated in small gossans, both in these sheared felsic rocks and in basalts proximal to felsic intrusions.

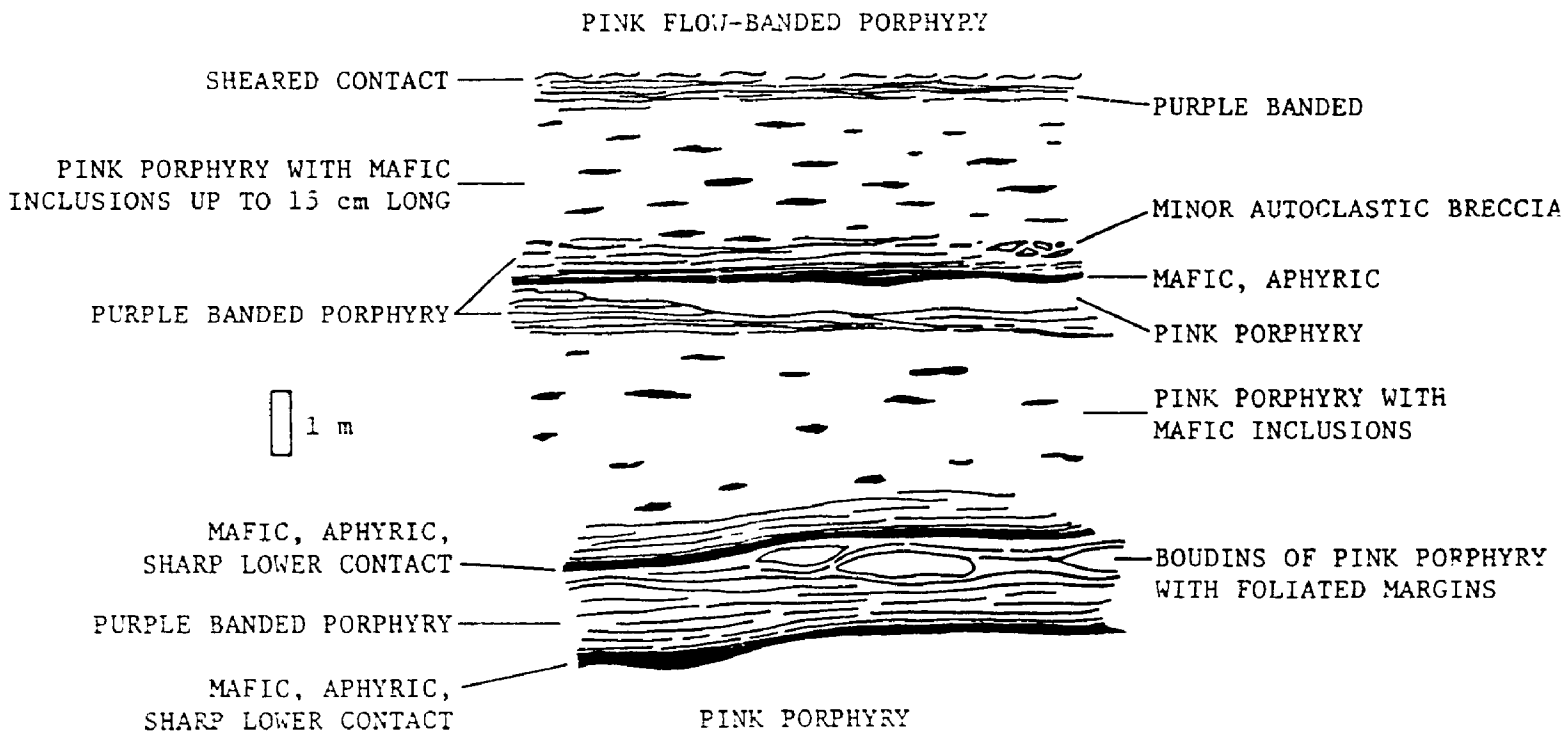


FIGURE 4. COMPOSITIONAL ZONATION IN PORPHYRY UNITS, NORTHEASTERNMOST REINHARDT ISLANDS (61°47'N, 112°59'W). FROM FIELD SKETCH.

GABBRO

Sills and dikes of metagabbro intrude sedimentary rocks of the Reinhardt formation. A gabbro sill in the Pebbly Sandstone member displays distinctive phenocrysts of actinolite that probably is pseudomorphous after tabular pyroxene.

UNNAMED SANDSTONE UNIT

A succession of arenites called the "Unnamed Sandstone unit" (unit Wro, Map 1A) is exposed in the core of the Blind Bay anticline, northeast of the "Scimitar" fault. Layers of argillaceous siltstone and basalt are present only near the top of this unit, whereas the lower stratigraphic levels consist almost entirely of medium-grained arkosic arenite. These arenites are comparable in texture and composition to arenites of the Lower Clastic member, although they contain only rare lithic clasts. The Unnamed Sandstone unit is inferred to be correlative with similar monotonous arenites (Wro?, Map 1A) which are exposed south of the Scimitar fault, above the basalt unit that caps the Pebbly Sandstone member.

UPPER PEBBLY UNIT

The unnamed sandstone unit is gradationally overlain by poorly sorted pebbly and granular arkoses with intercalated fine-grained arenites and argillaceous siltstones, called the Upper Pebbly unit. The base of the Upper Pebbly unit is marked by the lowermost pebbly or conglomeratic layer. Because of difficulty in defining the top of the unit, member status is not proposed.

Massive layers of medium-grained to granular arkose in successions up to 60 metres thick form about 65 percent of the unit. Subangular to

subrounded pebbles of quartz, feldspar, light brown siltstone, and pale green phyllite are dispersed through the sandy matrix and locally concentrated in planar layers. Quartz pebbles average 15-20 millimetres across, whereas the flat phyllite clasts commonly are 50-80 millimetres in maximum dimension. Round cobbles of fine-grained granitoid rock, and pebbles of mafic or intermediate volcanic rocks and red felsic porphyry (resembling the Reinhardt porphyries), also are present but are much less common.

Intercalated thin beds and laminae of sandstone and silty mudstone display planar lamination, graded bedding, wavy and lenticular bedding, and small-scale ball-and-pillow and load structures. The argillaceous components weather pale orange-brown to olive green; the arenites weather buff to light brown.

Rocks with similar characteristics are exposed at approximately the same stratigraphic position in the Reinhardt islands and on the southern end of Wilson Island. Extensive trough crossbedding is prominent in shoreline exposures of these units. Sandy lapilli-ash tuffs are intercalated with pebbly arkoses on one of the Reinhardt Islands, and below the upper thick felsic unit on the southern end of Wilson Island. Numerous basalt flows also occur within the Upper Pebbly unit.

BLIND BAY CONGLOMERATE

Abundant clasts of felsic porphyritic rocks, and a general abundance of large cobbles derived from a diverse assemblage of rock types, distinguish conglomeratic rocks exposed in Blind bay (this name comes from Stockwell and Kidd, 1931) and in the northern Reinhardt islands from other pebbly and conglomeratic units of the Reinhardt formation.

Rocks that exhibit these characteristics have been mapped as an informal unit within the Reinhardt formation called the Blind Bay conglomerate. The basal contact with the underlying Upper Pebbly unit is exposed at a single locality between the Inconnu Thrust and the southern shore of Blind bay, where conglomeratic layers of the Blind Bay conglomerate (containing abundant pebbles and cobbles of red rhyolite) gradationally overlie sandstones of the Upper Pebbly unit (containing mainly scattered pebbles of quartz and light-coloured phyllitic metasedimentary rocks)

On one of the Reinhardt islands (61°47'N, 113°03'30"W), conglomerates of the Blind Bay unit that rest on the autobrecciated top of a felsic volcanic flow contain 90 percent rounded clasts of the volcanic rock, 5-15 centimetres across, in a matrix of medium-grained arkosic arenite. Felsic volcanic clasts displayed elsewhere within the Blind Bay conglomerate include red and purple plagioclase-phyric varieties, pink flow-banded rhyolite, and light green rhyolite, each of which has *in situ* equivalents within the Reinhardt formation.

Other abundant clast types are intermediate or mafic metavolcanic rocks, foliated and non-foliated granitoid rocks, polycrystalline quartz, phyllitic metasedimentary rocks, and purple-grey banded metavolcanic rocks with weathered rims. Rounded pebbles of quartzitic arkose and subarkose (resembling rocks lower in the Reinhardt succession), siltstone, various quartz-bearing porphyries, and angular fragments of potassium feldspar also are locally common.

The pebbles, cobbles, and local boulders form very poorly sorted orthoconglomeratic layers, ranging from a few centimetres to a few metres thick, separated by medium- to very-coarse-grained feldspathic sandstones that commonly are many metres thick. Typical conglomerate

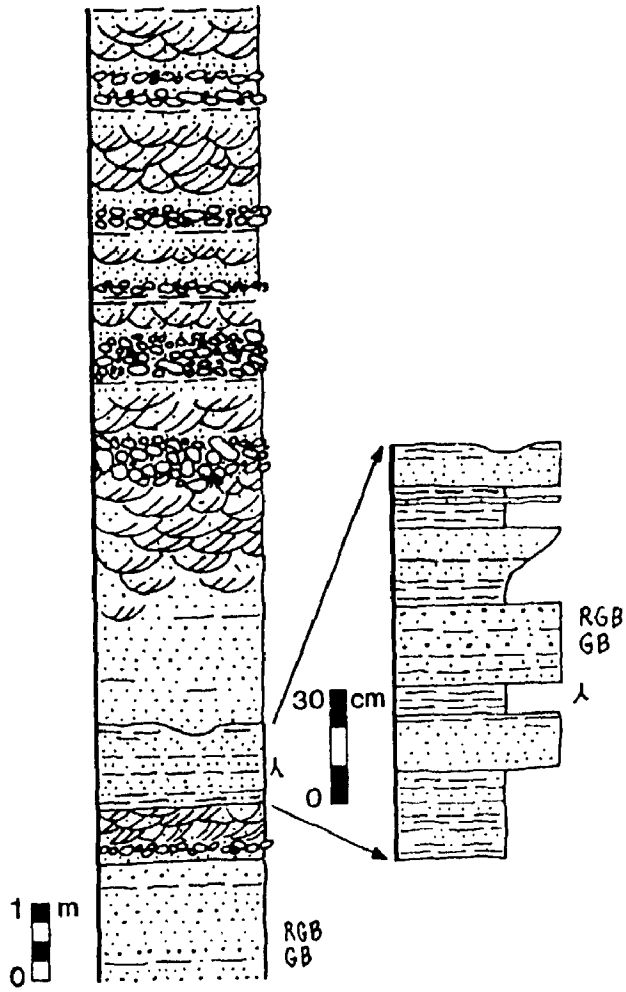
layers show marked lateral continuity, although some thin layers are lenticular; isolated pebbles and sparse pebbly lags also occur in the sandstones.

Small-scale fining-upward cycles within the Blind Bay conglomerate ideally exhibit a slightly erosional basal surface, above which a basal conglomerate grades successively into planar-laminated arenite, trough-crossbedded arenite, and planar-laminated arenite intercalated with silty mudstone (Fig. 5). The planar-laminated intervals at the tops of the cycles commonly comprise alternate layers of sandstone and silty mudstone, wherein both normal and reverse graded bedding are displayed (detailed section, Fig. 5). Trough-crossbedded units are especially well-preserved in the northernmost Reinhardt Islands; festoon crossbed sets with amplitudes of 5-25 centimetres constitute lenticular cosets 50-200 centimetres thick and many metres in lateral extent.

The top of the Blind Bay conglomerate (also the top of the Reinhardt formation) is marked by a distinctive, laterally extensive, granulestone unit several metres thick (Figure B). The poorly sorted granulestones comprise 80-85 percent subrounded quartz, 10-15 percent angular potassium feldspar, rare lithic fragments, and a matrix of white mica.

SAFETY COVE FORMATION

A mixed succession of quartzitic sandstone and carbonate that forms the main part of Wilson Island is here informally named the Safety Cove formation. The name Safety cove has been used in reference to a cove 3 kilometres from the western tip of Wilson Island (see Stockwell and Kidd, 1931). The Safety Cove formation sharply overlies



EXPLANATION

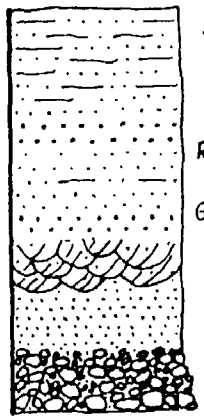


FIGURE 5. MEASURED STRATIGRAPHIC SECTIONS: TYPICAL SEGMENT OF BLIND BAY CONGLOMERATE.

the granulestone beds at the top of the Blind Bay conglomerate unit of the Reinhardt formation (Figure B).

Seven lithofacies are named to facilitate description of the Safety Cove formation:

Quartzose sandstone lithofacies

The quartzose sandstone lithofacies comprises fine- to medium-grained feldspathic sandstones that form tabular beds separated by thin argillaceous partings. The sandstones typically weather white, pink, or mauve. Most of these sandstones are subarkosic to arkosic arenites (classification of Pettijohn et al., 1973); feldspar grains generally constitute over 15 percent, and commonly over 25 percent, of their framework.

Carbonate-cemented sandstone lithofacies

The carbonate-cemented sandstones are similar to the quartzose sandstones, except that the grains are held together by interstitial carbonate cement. These sandstones weather white to pinkish grey and have a slightly porous appearance due to the dissolution of carbonate cement. Bedding generally is poorly defined.

Argillaceous sandstone lithofacies

The argillaceous sandstone lithofacies comprises very fine- to medium-grained feldspathic sandstones (commonly carbonate-cemented) that contain numerous silty and argillaceous laminae. Rocks of this lithofacies typically are grey on fresh surfaces and weather a distinctive light orange-brown to orange-pink. Intercalated mudstone laminae are pale green.

The sandstones commonly display festoon crossbedding, which is accentuated by the recessive weathering of carbonate cement. Wavy and lenticular bedding and ball-and-pillow structures are well developed in

units of alternate sandstone and mudstone. Petrographic studies have shown that the meta-argillaceous component in these rocks is a combination of biotite (which gives the rock its colour) and sericite.

Dolomitic sandstone lithofacies

Sandstones that display extensive cementation and replacement of the siliciclastic framework by brown-weathering crystalline (or recrystallized) dolomite are classified as dolomitic sandstones. They weather pink to white, with recessive brown streaks of carbonate that commonly parallel foresets of cross strata. The dolomitic sandstone lithofacies also includes quartzose sandstone with intercalated thin beds and tongues of dolostone.

Sandy dolostone lithofacies

Rocks that contain sand-sized siliciclastic grains and more than 50 percent carbonate are called sandy dolostones. The sandy dolostone lithofacies shows a compositional continuum, from rocks that contain 50 percent siliciclastic grains (intergradational with dolomitic sandstones) to massive brown dolostones with thin sandy laminae. Dolostones with intercalated thin beds of sandstone also are included in this lithofacies.

Argillaceous dolostone lithofacies

Argillaceous dolostones comprise alternate thin laminae rich in crystalline carbonate and pelitic mudstone, with fine silt-sized grains of quartz and feldspar dispersed throughout. The mudstone laminae typically are sericitic, but some are composed of chlorite or biotite/phlogopite. These rocks weather light brown to light greenish-grey or greyish-pink.

Laminated dolostone lithofacies

Laminated dolostones weather dark to medium brown and display planar to slightly undulose layers of alternate coarse- and fine-grained crystalline carbonate. In thin section the fine-grained layers are seen to contain sericite and about 5 percent biotite or phlogopite. Silt-sized grains of quartz and feldspar, present both in the coarse and fine laminae, constitute about 35 percent of the rock.

LOWER ARKOSE MEMBER

Above the granulestones that mark the top of the Reinhardt formation, greenish grey argillaceous rocks of the Lower Arkose member are intercalated with, and give way upward to, medium-grained quartzose sandstones. These sandstones are similar to those of the Blind Bay conglomerate. Dispersed granules of potassium feldspar and felsic volcanic rock are common, and sparse lags of quartz and feldspar pebbles occur locally, but pebble and granule layers are at most a few centimetres thick. The sandstones weather white to mauve, are medium- to thick-bedded, and exhibit planar lamination, trough crossbedding, and ripple lamination. Some intervals exhibit numerous heavy mineral laminations.

Argillaceous sandstones are abundant in some parts of the succession. They display festoon crossbedding, wavy and lenticular bedding, ball-and-pillow structures, and rhythmically interbedded laminae of sandstone and mudstone up to 5 centimetres thick. A slab of rock with the latter type of bedding shows that the sandy units within the rhythmite couplets internally comprise thin (1-5 mm) planar sandstone laminae, separated by drapes of mudstone. Each compound sandstone unit

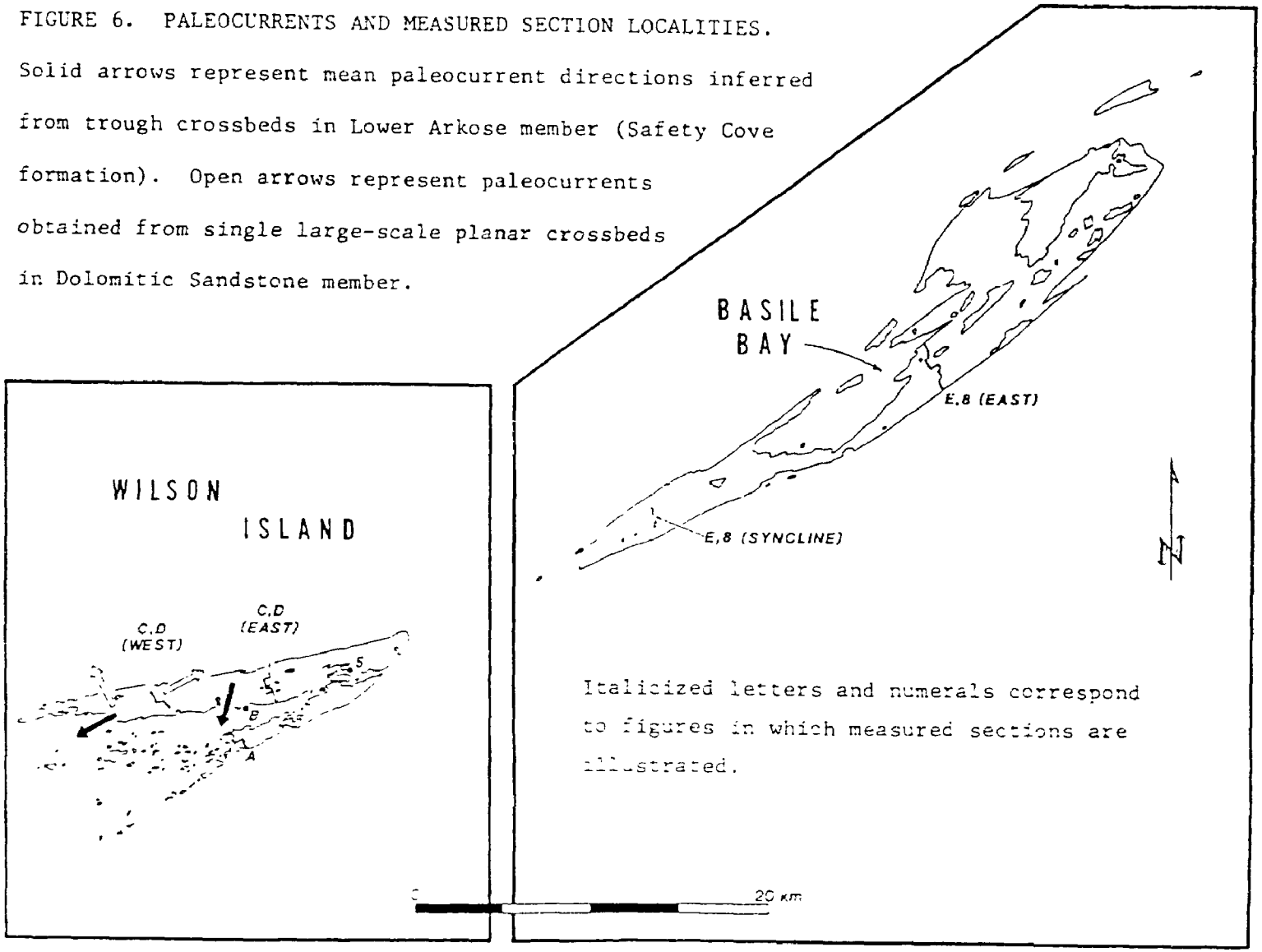
displays a rippled top.

The upper part of the Lower Arkose member consists of pink to pale orange quartzose sandstone, in which graded bedding and crossbedding are abundant. The crossbed foresets are accentuated by heavy minerals. The foresets typically appear to be inclined at a low angle to bedding, although they are only exposed in two dimensions (Fig 3H). Beds of carbonate-cemented sandstone and discontinuous thin beds of brown sandy dolostone are present in this interval.

Limited paleocurrent data were obtained from trough crossbeds. Because of low curvature of the individual troughs and the consequent difficulty of measuring trough axes, numerous measurements of the strike and dip of various parts of each foreset were recorded at each of the two stations. Stereographic corrections were made for the plunge of fold axes (indicated by local bedding-cleavage intersections) and for the inclination of strata. The resultant southwest modes are illustrated in Figure 6.

Three thin sections show that the quartzose sandstones are well-sorted to moderately sorted, and that the framework grains are subrounded to rounded (Fig. 7A). The estimated average modal composition is quartz (mostly polycrystalline megaquartz; 60-70%), feldspar (mostly orthoclase and perthite, some plagioclase; 15-25%), lithoclasts (mostly siliceous volcanic rocks, some mudstone; 10-15%), heavy minerals (2%), interstitial white mica (5%), and hematite cement (1-2%). The heavy mineral grains mostly are magnetite(?); tourmaline and zircon also are common. Detrital muscovite, biotite, and epidote occur in trace amounts.

FIGURE 6. PALEOCURRENTS AND MEASURED SECTION LOCALITIES. Solid arrows represent mean paleocurrent directions inferred from trough crossbeds in Lower Arkose member (Safety Cove formation). Open arrows represent paleocurrents obtained from single large-scale planar crossbeds in Dolomitic Sandstone member.



Italicized letters and numerals correspond to figures in which measured sections are illustrated.

DOLOMITIC SANDSTONE MEMBER

The Dolomitic Sandstone member comprises complexly intercalated quartzose sandstone and dolostone, and ranges from 550 to 650 metres in total thicknes . The basal contact, marked by the lowermost occurrence of laterally extensive beds of brown dolostone, ranges from sharp to gradational. Underlying sandstones locally have been extensively replaced by carbonate.

On a regional scale the member is characterized by a distinct westward decrease in the siliciclastic/carbonate ratio (from about 3.5:1 to 1:1), manifested by marked westward thinning of the sandstone units and complementary thickening of the intervening carbonate-rich units along Wilson Island. This relationship was documented by mapping across Wilson Island, and is illustrated by two measured sections (Figures C and D). However, precise correlation of units on opposite sides of the fault zone that extends northeastward, across about the middle of the Wilson Island map area (Map 1A) would only be possible with more detailed work, and hence the correlations shown in Figure D may be incorrect.

Each of the lithofacies of the Safety Cove formation is represented in the Dolomitic Sandstone member. Many of the facies boundaries are gradational, although sharp contacts between sandstones and dolostones also are common.

Quartzose sandstones display trough and planar crossbedding and current ripple marks. As in the Lower Arkose member, the crossbedding in some layers is accentuated by heavy minerals, and crossbed foresets commonly dip at shallow angles with respect to bedding. The tops of some ripple beds show reactivation surfaces (Fig. 7B). Interference ripple marks and diametrically opposed current-ripple marks occur lo-

cally.

Dolomitic sandstone and sandy dolostone units also exhibit both trough and planar crossbedding (Fig. 7C). Foresets are accentuated by the positive relief of laminae rich in quartz and feldspar grains, in contrast to recessive brown dolomitic laminae. Planar crossbed sets in some sandy dolostone beds are over 1 metre thick. The dips of foresets in many of these thick units is opposite to the dip direction of foresets in the crossbedded sandstones; one such unit indicates a northeast-directed paleocurrent (Fig. 6).

Dispersed pebbles and concentrated pebble lags are common in the dolomitic, carbonate-cemented, and quartzose sandstones. They also occur in sandy dolostones, and in transitional intervals between some sandstones and dolostones. Some of the pebbly layers, although discontinuous, have been traced laterally for over 3 kilometres. One such layer marks the base of a sandstone unit that overlies the eroded top of a dolostone unit, in about the middle of the western section (Figure D). The most common pebble types are vein quartz, potassium feldspar, red felsic volcanic rocks, pink to light green quartzitic sandstone, and phyllitic mudstone. Mafic(?) volcanic rock fragments are less common. The quartzitic sandstone and felsic volcanic clasts look identical to rocks lower in the succession on Wilson Island.

The argillaceous sandstone lithofacies is characterized by trough crossbedding and ball-and-pillow structures. Wavy and lenticular bedding and sandstone-mudstone couplets are common in some intervals.

Soft-sediment deformation structures also are abundant in the laminated and argillaceous dolostone lithofacies. Slump sheets locally form single recumbent folds up to 50 centimetres in wavelength (i.e.,

thickness) and several metres in amplitude (i.e., bed length). Slump folds of smaller dimensions are common. Ball-and-pillow structures in carbonate units exhibit detached segments that locally are 45 centimetres across.

The top of the Dolomitic Sandstone member is the top of the uppermost, laterally extensive, brown dolostone bed below the sandstones of the Middle Arkose member.

MIDDLE ARKOSE MEMBER

A succession of predominantly quartzose sandstone lithofacies that ranges in thickness from 260 metres in the east to about 120 metres in the west is called the Middle Arkose member. In the easternmost exposures of this unit, as typified by the eastern section of Figure C, quartzose sandstone is virtually the only lithotype present. The predominant sedimentary structures progress upward from trough crossbedding and ripple marks, to low-angle (apparent) crossbedding outlined by heavy mineral laminae, to planar lamination and graded bedding. Dark grey laminae rich in iron oxides are increasingly abundant in the uppermost 10-30 metres. Wedge-planar crossbedding is common in the upper 10 metres of the member.

In the western part of Wilson Island (western section, Figure C), the lower part of the Middle Arkose member consists mainly of carbonate-cemented sandstone; interbedded quartzose and muddy sandstones and lenses of brown dolostone are less abundant. Quartzose sandstones become predominant about 30 metres above the base of the member in this section.

SANDY DOLOSTONE MEMBER

The Sandy Dolostone member is about 170 metres thick. It consists mainly of dolostone and dolomitic sandstone, with less abundant carbonate-cemented sandstone, and minor quartzose sandstone. Intercalation of the various lithofacies is complex, and replacement of sandstone by carbonate is extensive. Contacts of the member with both the underlying and overlying sandstone units are gradational, and are marked by continuous dolomitic layers.

UPPER MIXED UNIT

The Upper Mixed unit consists mainly of quartzose and carbonate-cemented sandstones, with numerous intercalated beds of dolostone and dolomitic sandstone. The top of the unit is truncated by the McDonald-Wilson Fault.

FIVE SNARES ASSEMBLAGE

Siliciclastic and dolomitic metasedimentary rocks, exposed in a series of nearly isoclinal folds near the tip of the peninsula ("Basile point") southwest of Basile Bay, and on islands south of Keith Island, are provisionally named the "Five Snares" assemblage. "Five Snares" is the English translation of the local Chipewyan name for Basile point (anonymous resident of Snowdrift, N.W.T., written comm., 1986). The Five Snares assemblage comprises four units:

ARKOSE

The predominant rock type of the lower part of the assemblage is medium- to coarse-grained quartzo-feldspathic sandstone, similar to the quartzose sandstones of the Shelter Cove formation. Granules of feld-

FIGURE 7.

A. Photomicrograph of quartzose sandstone, Lower Arkose member. Grains of potassium feldspar and felsic volcanic lithoclasts (V) are dark relative to quartz. Plane-polarized light. Scale bar = 1 mm.

B. Reactivation surfaces (defined by dark laminae) in rippled and crossbedded sandstone. Lower Arkose member, Safety Cove formation.

C. Planar crossbedding in dolomitic sandstone lithofacies, accentuated by alternate dolomitic (dark) and quartzose laminae. Dolomitic Sandstone member, Safety Cove formation.

D. Slump-folded unit in Five Snares dolostone. Internal laminae define elliptical closures, a manifestation of smaller sheath folds within compound folded unit. Laminae near lower right corner of photograph are truncated above a soft-sediment thrust fault (arrows).

E. Photomicrograph of quartzose sandstone of Lower Quartzite member, Basile formation. Cross-polarized light. Scale bar = 1 mm.

F. Wavy and lenticular bedding in Lower Pelitic member, Basile formation.

G. Load-casted sandstone lenses in argillaceous siltstone/sandstone lithofacies. Lower Pelitic member, Basile formation.

H. Ball-and-pillow structures in sandstone (light color) intercalated with argillaceous siltstone.

spar and quartz are dispersed to concentrated in some layers. The beds typically are of medium thickness, and are separated by argillaceous partings 1-2 centimetres thick. Trough crossbeds, ripple marks, and graded bedding are rarely displayed.

A minor proportion of this unit is formed by intercalations of micaceous quartzite, phyllitic mudstone, sandy dolostone, and (on islands south of Keith Island) specular hematitic quartzite.

FELSIC PORPHYRY

Lenses of red to purple felsic porphyry either are intercalated with, or intrude, the sandstones and mudstones of the arkose unit near the tip of Basile point. Petrographic studies have shown that these rocks are composed of microcrystalline quartz and white mica, with disseminated opaque minerals. They contain up to 10 percent phenocrysts of plagioclase (albite?). Similar fine-grained quartz-sericite rocks within the assemblage are aphyric, and may not be of igneous protolith.

ARGILLACEOUS UNIT

Overlying the arkose are laminated to thin-bedded argillaceous siltstones and mudstones, intercalated with dolostone and minor sandstone. The argillaceous rocks are in part calcareous.

DOLOSTONE

A succession of dolostone, probably more than 800 metres thick, forms conspicuous high-standing outcrops. The contact of the dolostone unit with underlying argillaceous unit is characterized by intercalations of argillaceous siltstone, mudstone, sandstone, and dolostone.

Rocks of the dolostone unit include laminated, argillaceous, and sandy dolostones, which differ from similar lithotypes of the Safety Cove formation mainly in that sand- and silt-sized siliciclastic grains are much less abundant; quartz and feldspar generally occur only as dispersed grains, that show positive relief on weathered surfaces and locally define thin laminations in the sandy lithofacies. Many of the units contain thin layers and nodules of black or cream chert. Beds of massive crystalline dolostone also are common.

The rocks weather dark brown to light greyish-brown or greyish-pink, and are creamy white to pink on fresh surfaces. Beds range from a few metres to several metres thick. Primary sedimentary structures include planar to slightly undulose lamination and rare small-scale cross lamination. The most common sedimentary structures in the dolostones are slump folds, which in many cases involve entire beds several metres thick. Internally, each slump-folded unit is composed of smaller sheath folds (Fig. 7D). Soft-sediment thrust faults are locally developed within the slump sheets.

The stratigraphy and petrography of a well-exposed section of the dolostone unit were studied in detail by Wong (1986), who demonstrated that lateral facies changes within the succession are common and abrupt.

BASILE FORMATION

The thick succession of predominantly siliciclastic metasedimentary rocks that overlies the Five Snares assemblage, and which is exposed from southwest of Basile Bay to Basile Lake, is here informally named the "Basile formation." The base of the Basile formation is

marked by the first occurrence of light-coloured quartzose sandstone, with or without intercalated shale and dolostone, above the Five Snares dolostone unit. The top of the formation is nowhere exposed.

The Basile formation has been subdivided into five informal members and three other units. Four lithofacies have been named to facilitate description of the rocks:

Pink / white (quartzose) sandstone lithofacies

This lithofacies comprises well-sorted, fine- to medium-grained quartzose sandstones that range in colour from white to pink and display characteristic even bedding. The beds average 20-30 centimetres thick (minimum 15, maximum 100), typically are continuous laterally over several tens of metres, and are separated by thin argillaceous partings. The most abundant primary structures are large-scale trough and planar crossbedding (cf. Reineck and Singh, 1980). Individual crossbed sets commonly are 15-20 centimetres thick, although some trough crossbed sets are as thick as 100 centimetres. Graded bedding also is common, especially in the upper few centimetres of bedsets, where quartzose sandstone gives way upward to progressively finer-grained and more argillaceous sandstone. The argillaceous component almost invariably is sericite.

Nine thin sections of quartzose sandstones were examined. Most of the quartz grains display sutured boundaries, making textural analysis difficult. It appears that the sandstones generally are fairly well-sorted, and that the grains are subrounded to rounded except for some angular feldspars (Fig. 7E). Subarkosic and arkosic arenites seem to be equally represented. Estimated modal mineral compositions are: monocrystalline quartz (55-70%), potassium feldspar (orthoclase, microcline, and perthite) and less abundant plagioclase (collectively, 15-

35%), sericite (8-15%), and heavy minerals, including iron oxide, zircon, tourmaline, and epidote (trace to 5%). A polished thin section in reflected light showed iron oxide grains composed of magnetite with hematite alteration.

Rocks of this lithofacies are hereafter referred to as white or pink sandstones where applicable, and as quartzose sandstones otherwise.

Grey sandstone lithofacies

Rocks of the grey sandstone lithofacies are medium greyish pink to dark grey on fresh surfaces and weather light greyish-pink to purple-grey. They typically are fine-grained quartzose sandstones, although they display a continuum from medium-grained sandstone to very-fine-grained argillaceous sandstone. Grey sandstone units commonly appear massive in outcrop, except where sedimentary structures (parallel lamination, graded bedding, and rare ripple- and cross-lamination) are accentuated by thin (microscopic to 1 cm) argillaceous layers or by heavy mineral laminae. Graded bedding is displayed in thin sections of some of the rocks that otherwise appear structureless.

Petrographically, the grey sandstones are feldspathic arenites or wackes that contain abundant white mica and also biotite. Most of the mica forms thin laminae (i.e., metamorphosed clayey layers); it is not clear how much of it was matrix. Apart from the mica, the grains are well-sorted and subangular to rounded. Zircon, tourmaline, detrital muscovite, opaques (magnetite ± hematite, 10-15%, probably in part diagenetic), garnet, and epidote(?), occur in small quantities.

Argillaceous siltstone lithofacies

Units that range from very-fine-grained argillaceous sandstones to

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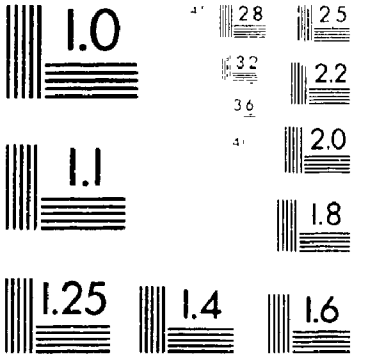


MicroD

mudstones, that typically display marked fissility, and that weather orange-brown to olive-gray or olive-brown, collectively are called argillaceous siltstones. These units typically consist of alternate thin beds and laminae of silt-rich and silt-poor mudstone, and exhibit parallel and undulose laminations and graded bedding. Petrographically, the silt-sized grains include both quartz and feldspar, and the pelitic component typically is white mica with subordinate biotite and/or chlorite. In addition, they contain about 5 percent heavy minerals, including tourmaline, zircon, and detrital biotite.

Argillaceous siltstone/sandstone lithofacies

Units of intercalated thin beds of argillaceous siltstone and grey sandstone lithofacies in subequal proportions, and less common beds of quartzose sandstone lithofacies, are for the sake of brevity classified as argillaceous siltstone/sandstone lithofacies. Typical sedimentary structures of this lithofacies include wavy and lenticular bedding, parallel, ripple and undulose laminations, graded bedding, load casts, and ball-and-pillow structures.

LOWER QUARTZITE MEMBER

In the southwestern part of the Basile Bay map area, the thick Five Snares dolostones are overlain abruptly by pink quartzose sandstone with thin intercalations of argillaceous siltstone. The contact is sharp and even, with local relief of up to 10 centimetres. Locally, sandstone immediately above the dolostone contains granules of potassium feldspar. Crossbedding in the sandstones commonly is accentuated by heavy minerals concentrated on foreset laminae. Small-scale ripple bedding, wavy bedding, mud rip-ups, local ball-and-pillow structures, and rare desiccation cracks occur in the argillaceous siltstone/sand-

stone units, which give way within a few metres up-section to quartzose sandstone.

East of longitude 111°41'W, quartzose sandstone and argillaceous siltstone/sandstone units in the basal part of the Lower Quartzite member contain intercalated thin beds or tongues of brown sandy dolostone and dolomitic sandstone (cf. lithofacies of Safety Cove fm.). The dolomitic units typically are less than one metre thick (maximum 2.5 m) and have gradational contacts with the intercalated sandstones. They display trough and planar crossbedding, which locally shows bidirectional orientations. Some beds contain abundant coarse sand- to pebble-sized (<1 cm) grains of quartz and feldspar.

This basal interval of mixed lithofacies, which typically is poorly exposed (underlying some of the nicer swamps in the area), thickens from a few metres in the southwest to over 250 metres east of Basile Bay (Figure E). Where possible, the part of this interval below the uppermost thick (50 cm. or more) and laterally continuous brown dolomitic layer was mapped as a separate unit (Wbds, Map 2A and Figure E).

The remainder of the Lower Quartzite member is a thick succession of pink to white quartzose sandstone, that forms resistant outcrops of moderate topographic relief. In the lowermost 200 metres of this succession east of Basile Bay, beds of light greyish-pink carbonate-cemented sandstone and thin dolomitic lenses form a small proportion of this interval compared to the quartzose sandstones. West of Basile Bay, carbonate-cemented sandstones are absent, and dolomitic lenses occur only locally. Sedimentary structures are rare in this part of the section.

Higher in the section, trough and planar crossbeds are abundant. The planar foresets commonly are bounded by wedge-shaped surfaces. At least 90 percent of the crossbeds indicate generally southwest-directed transport (based on visual inspection of two-dimensional exposures, allowing for estimated restoration of bedding to horizontal). In a few places near the top of the member, however, the orientation of foresets in two-dimensional exposures suggests northeast-directed flow.

Sedimentary structures are most conspicuous where they are highlighted by heavy mineral laminae, which are extensively displayed throughout an interval in the upper part of the member (Figure E). There, the fine-grained upper portions of many crossbed cosets exhibit small-scale crossbedded (or in some cases current-ripple- or parallel-laminated) intervals up to about 4 centimetres thick, which in turn grade upward into the more pelitic tops of the cosets. Crossbeds with gently curved foresets and low-angle truncations also are common. Small slump folds, some of which are recognizable as oversteepened foresets, are abundant in some layers.

The sandstone within this succession generally is pink on both fresh and weathered surfaces, except near the middle of the member, where both surfaces are white to very pale pink. Thin sections show that hematite grain coatings occu in all but the white samples, in which hematite is concentrated in a few small spots or is absent altogether. Much of the sandstone in the lowermost part of the succession is orange-pink, owing its colour to extensive hematite coatings on quartz and feldspar grains.

LOWER PELITIC MEMBER

The Lower Pelitic member consists of a succession of fine-grained

clastic rocks that are subdivided, from bottom to top, as indicated below and in Figure 8.

1. A basal "transitional" interval (pw_1), up to about 20 metres thick, consists predominantly of wavy- to lenticular-bedded argillaceous siltstone/sandstone units (Fig. 7F). Load-casted sandstone lenses (Fig. 7G) and ball-and-pillow structures (Fig. 7H) are abundant in some layers. "Pseudonodules" over 25 centimetres across locally are present. Thin to medium beds of grey sandstone (in the southwest) and/or pink sandstone (especially in the northeast) form up to 40 percent of this subdivision. The contact with the pink sandstones of the underlying Lower Quartzite is distinct, but is gradational over about 1 metre in some areas.

2. The ps_1 subdivision is several tens of metres thick, consisting of grey sandstone and minor intercalated layers of argillaceous siltstone. In the southwest, the argillaceous layers give the sandstones a blocky to platy weathering profile, whereas in the northeast such layers are rare, and the sandstones are thick-bedded and appear massive.

3. Subdivision pp comprises argillaceous siltstone/sandstone beds which gradationally overlie the ps_1 sandstones. Beds of grey sandstone up to about 1 metre thick are present in the lower and upper parts of this interval, whereas argillaceous siltstones in the middle part contain virtually no intercalated sandstone. The predominant sedimentary structures in rocks of the pp subdivision are planar and undulose laminations and graded bedding.

4. Thick beds of grey sandstone with minor intercalated argillaceous siltstone layers form the ps_2 subdivision. This unit thins

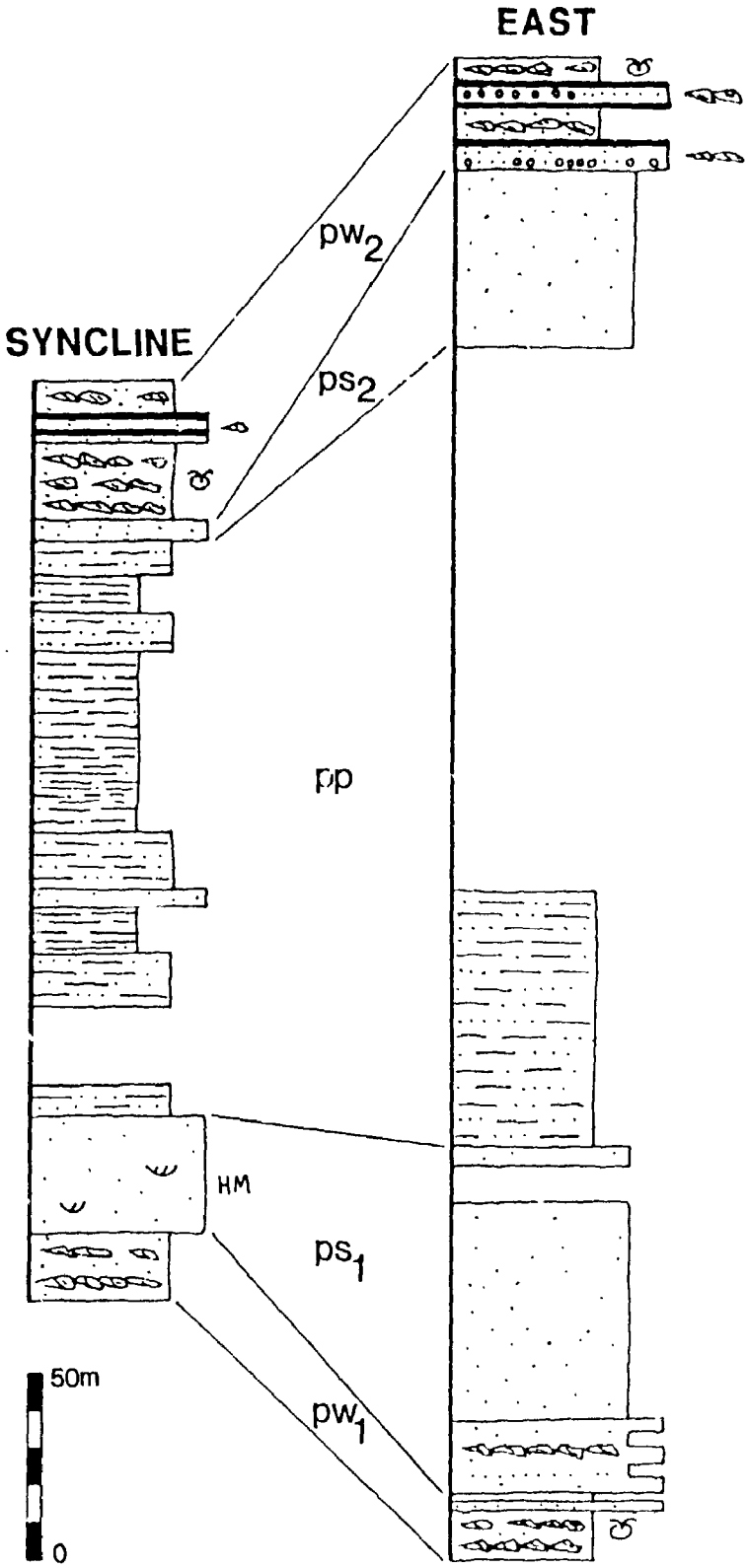


FIGURE 8. MEASURED STRATIGRAPHIC SECTIONS: LOWER PELITIC MEMBER, BASILE FORMATION.
See Figure E for explanation of symbols.

markedly from at least 50 metres in the northeast to less than 10 metres in the Basile Bay syncline (Fig. 8).

5. An upper "transitional" interval (pw_2) consists mainly of argillaceous siltstone/sandstone units and intercalated grey sandstones. Wavy bedding is the most prominent sedimentary structure. Lenticular bedding and load casts also are common.

The grey sandstones occur both as isolated thin beds and as units several metres thick. The thicker units display centimetre-scale ferruginous layers, planar laminations, and discontinuous granule laminae up to 2 centimetres thick. The granule laminae commonly form the bases of thin graded sandstone sequences that are capped by wavy-bedded layers.

A few beds of pink medium-grained sandstone that occur in the pw_2 subdivision are most numerous (and thickest) in the northeastern part of the map area.

UPPER QUARTZITE MEMBER

The Upper Quartzite member consists of a succession of white quartzose sandstone that thins southwestward from over 100 metres to about 40 metres (compare Figure E, Map 2A). The basal contact with argillaceous siltstone/sandstone units of the Lower Pelitic member is sharp and even.

Planar laminations and numerous quartz-granule layers are present throughout an interval of a few metres near the base. The granule layers are up to several centimetres thick, display gradational contacts with both underlying and overlying sandstones, and generally lack internal stratification. Within these layers the grains are poorly sorted; clast sizes range from medium sand to about 5 millimetres. The

granules are well-rounded and are composed exclusively of coarsely polycrystalline quartz. The predominant sedimentary structure is trough crossbedding. Typical cosets are 15-20 centimetres thick and show marked lateral continuity, although in some intervals the lateral extent of cosets more commonly is only a few metres. A few sandstone beds near the top of the Upper Quartzite member locally display layers rich in disseminated iron oxides.

IRONSTONE MEMBER

Thick successions of grey sandstone and subordinate argillaceous siltstone/sandstone units form the 90- to 100-metre-thick Ironstone member. A unique characteristic of the member is the presence of numerous dark iron oxide-rich layers in the sandstones.

The contact between the Ironstone member and the underlying Upper Quartzite member commonly is gradational, characterized by the intercalation of white sandstones with grey ferruginous-layered sandstones and, in some areas, wavy-bedded argillaceous siltstone/sandstone units. In other areas (e.g., Basile Bay syncline section, Figure E) the contact is marked by an abrupt up-section change from white sandstone to grey ferruginous sandstone.

The ferruginous layers range in scale from thin laminae to beds of ironstone 35 centimetres thick. They range in colour from dark purple-grey to a distinctive dark purple. Iron oxide minerals in these layers (chiefly magnetite and specular hematite, with less abundant jasper) compose up to 40 percent of the rock. Polished thin sections show that the magnetite grains commonly have rhombic outlines and rims of hematite alteration (Fig. 9A). The non-ferruginous component of these

rocks is a mixture of granular fine-grained quartz, sparry carbonate, and minor chlorite.

Apart from the ferruginous layers, the sandstones of the Ironstone member weather greyish-pink and contain about 10 percent disseminated iron oxides, typical of the grey sandstone lithofacies. The succession contains some intervals of sandstone several metres thick which lack ferruginous layers. Some of these units instead display thin, discontinuous, recessive-weathering calcareous laminations that are pink on fresh surfaces.

Grey sandstones (both with and without ferruginous layers) are intercalated with argillaceous siltstones in the upper part of the member. These rocks typically form sharp-based graded sequences, 10-30 centimetres thick, that show complete compositional grading from sandstone to mudstone. Argillaceous siltstones, which form the middle and upper portions of graded sequences, display undulose and planar laminations defined by alternate silty and pelitic laminae. Isolated sandstone lenses locally are present within the argillaceous siltstone layers.

The top of the Ironstone member is marked by a discontinuous dark purple bed of pisolitic ironstone that typically is 20-50 centimetres thick. Calcareous pisoliths (mainly composed of white to pink dolomite) are densely concentrated in the basal part of the bed. Isolated and dispersed pisoliths also occur above this basal zone, but generally are absent near the top. The pisoliths weather recessively and display a brown weathered surface. Where they are concentrated in the basal part of the bed, the rock weathers dark brown and has a rubbly outcrop expression; where they are dispersed, the pisoliths weather out to form cavities in the enclosing purple ironstone.

The pisoliths generally are flattened parallel to the regional cleavage, exhibiting oblate-spheroidal to subspherical shapes. They commonly are about 5-6 millimetres in maximum dimension, although pisoliths as large as 20-30 millimetres (rarely up to 45 mm) across typically occur in about the middle of the bed. In some areas it appears that, above the basal zone of highly concentrated pisoliths, the overall upward decrease in the concentration of pisoliths is accompanied by an increase in the size of the largest pisoliths. However, some large pisoliths also occur in the basal concentrated zone.

The pisoliths commonly display a concentric internal structure, accentuated by laminae of hematite or jasper grains. These hematite grains are elliptical in outline and, in thin section, display cores of carbonate (Fig. 9B). Some of the grains also display concentric laminae of hematite and of material that remains dark in reflected light. Slabs cut from one sample of the pisolitic unit show, in the basal concentrated zone, pisoliths that contain nuclear flat clasts of quartzose siltstone up to 25 millimetres across (Fig. 9C). Ovoid hematite/carbonate grains form a thin veneer over the top of the pisolite zone, and occupy the gaps between adjacent pisoliths (Fig. 9D).

The pisolitic ironstone bed is exposed almost continuously from the core of the Basile Bay syncline to about 2 kilometres southwest of Basile Bay (a restored distance of over 6 km along strike), and at numerous localities between there and Basile Lake (where outcrops of this stratigraphic level are relatively limited). On Map 2A, a solid line denotes the upper contact of the Ironstone member only where this bed is exposed.

GRADED MEMBER

The Graded member comprises a succession of interbedded grey sandstone and argillaceous siltstone about 650 metres thick. Throughout most of the map area, the Graded member sharply overlies the pisolitic ironstone bed. Where this bed is absent, the contact is considered to extend across the same stratigraphic horizon, but its position could only be estimated to within a few metres in the field.

The rocks that make up the Graded member typically are grey to greenish-grey on fresh surfaces. The sandstones are fine-grained to very fine-grained and weather pink to light grey, whereas the argillaceous components weather orange-brown to olive-grey or olive-brown. These rocks typically form compositionally and texturally graded sequences (i.e., upward increase in the ratio of pelitic components to quartz and feldspar, accompanied by an overall decrease in the size of quartz and feldspar grains) that display a corresponding gradation of colour on the weathered surface. For example, sequences that grade from fine-grained sandstone to argillaceous siltstone to mudstone typically show an upward change in colour from pink to orange-brown to olive-grey; sequences that grade from very-fine-grained argillaceous sandstone to argillaceous siltstone commonly show an upward colour change from light grey to light orange-brown. Typical graded units are 15-80 centimetres thick.

The sandstone portions of graded sequences are 10-20 centimetres thick throughout most of the succession. However, sandstone units 30-50 centimetres thick are common in some intervals, especially in the upper half of the member, and some sandstone beds are 150 centimetres thick. The sandstones generally have sharp, even bases (rarely load-casted) and gradational tops. Thin sections of these sandstones dis-

play thin, planar, argillaceous (sericitic) laminae between the sandy laminae.

The argillaceous siltstones exhibit planar and slightly undulose laminations, defined by alternate pelitic and silty or, locally, calcareous laminae. Discontinuous sandy laminae occur in some of the argillaceous units, and in many cases form the basal portions of recurrent graded subunits. Less commonly these units display wavy and lenticular bedding, and thin granule laminae.

Sandstone becomes progressively less abundant near the top of the Graded member, giving way up-section to argillaceous siltstone and mudstone.

MUDSTONE

An unknown thickness of green to dark grey laminated mudstone gradationally overlies the Graded member. Planar lamination is defined by alternate laminae of contrasting grain size, each up to a few millimetres thick. The coarser-grained layers generally are light-coloured and silty. Dark hematitic laminae are abundant in some intervals. Mudstones exposed on the northeast shore of the large island in Basile Bay contain strained carbonate concretions up to over 50 centimetres across.

Petrographic studies indicate that representative samples of the mudstones are composed of 60-70 percent phyllosilicate minerals (sericite, chlorite, and/or biotite), up to 25 percent quartzose silt, up to 15 percent disseminated opaque minerals (mainly pyrite) and traces of heavy minerals.

BASALT

Discontinuous layers of plagioclase-phyric basalt or andesite are intercalated with mudstones on the large island in Basile Bay (Map 2A), and are well-exposed along the island's eastern shore. On the southern part of this shoreline, the basalts display large elliptical structures, up to 2 metres in maximum dimension, that resemble pillows. Breccias in the northeastern part of the island consist of angular fragments of vesicular-porphyritic basalt, a few centimetres to several metres across, surrounded by dark grey mudstone. Because the outcrop permits only a two-dimensional view, it is not clear whether the largest blocks of basalt are completely isolated within, or are simply infilled by, the mudstone matrix.

Lath-shaped plagioclase phenocrysts locally form over 20 percent of the basalt. A thin section shows that the groundmass is composed of iron-rich chlorite, epidote, biotite, tremolite-actinolite, white mica, and microlitic plagioclase.

At the northeast end of Basile Lake, an amygdaloidal basalt unit is intercalated with rocks of the Graded member (Map 2A). This basalt contains phenocrysts of plagioclase, and of chlorite pseudomorphous after pyroxene(?).

INTRUSIVE ROCKS

Two small intrusions occur in the Lower Quartzite and Lower Pelitic members, near the southwest end of Basile Bay. The larger of these is a medium-grained metadiorite, composed of highly altered plagioclase (andesine?), chlorite (Mg-, Al-rich), and minor biotite and iron oxides. The smaller body is composed mainly of biotite, with unidentified phenocrysts that have been almost completely replaced by musco-

vite. This rock also contains microcrystalline quartz ± feldspar, carbonate, and disseminated titanite. Both of these units exhibit a prominent foliation parallel to the regional cleavage. Together they are named the "Rabesca intrusive suite," for the family name made conspicuous by part-time residents of a cabin 2 kilometres east of the smaller intrusion.

Another pluton intrudes the Mudstone unit along the Basile Bay fault. This rock, a coarse-grained, locally foliated sodic granite, consists of intensely kinked albite plagioclase (and possibly orthoclase; 75%), granulated interstitial quartz (20-25%), and disseminated pyrite.

"TURBIDITE" UNIT

The "turbidite" unit, which crops out on islands northeast of Wilson Island, comprises intercalated feldspathic wackes, silty mudstones, quartzose sandstones, and unstratified, green-weathering schists that resemble tuffs. These schists contain grains of coarse sand-sized quartz, dispersed in a fine-grained matrix of biotite, quartz, and feldspar. The feldspathic wackes are poorly sorted. The "turbidite" unit was not mapped in detail, and hence the genetic name used in previous descriptions (Hoffman *et al.*, 1977; Hoffman, 1977) has been used here.

OUTPOST ISLANDS

A southward-facing succession of rocks in the northeastern archipelago of the Outpost Islands (Fig. 10) has been subdivided into six lithostratigraphic units:

UNIT 1

Quartzites and micaceous quartzites of Unit 1 are extensively trough crossbedded and, in the lower part of the exposed succession, contain abundant rounded pebbles of quartz. Subparallel to bedding in about the middle of the unit are zones rich in chalcopyrite and pyrite, that also contain gold, tungsten, and tin (Lord, 1951). Near the top of the unit, quartzites that display sets of crossbeds and current-ripples alternate with thin layers of quartz-biotite schist, and contain angular pebbles of the same.

UNIT 2

Unit 2 comprises subequal proportions of intercalated micaceous quartzite and metagreywacke, with local thin layers of biotite schist. Metagreywackes at the base of Unit 2 lie with erosional contact on crossbedded quartzites of Unit 1. The metagreywackes are quartzo-feldspathic biotite-muscovite schists. They contain lenses and (in the lower part of the unit) boulders of quartzite. Andalusite porphyroblasts that resemble golf balls in both size and shape are locally abundant in the schists (Fig. 9E). The micaceous quartzites display trough crossbedding.

UNIT 3

Thick, monotonous, white quartzites at four different stratigraphic levels are collectively called Unit 3. Except for rare trough crossbedding, sedimentary structures in these quartzites are absent or have been obscured by deformation.

UNIT 4

Metagreywacke, with lenses and thin beds of white quartzite and porphyroblastic mica schist, forms Unit 4. Much of the metagreywacke is paraconglomeratic, containing pebbles, cobbles, and boulders of white quartzite. The thin quartzite beds commonly are boudinaged and locally form rootless intrafolial folds. Unit 4 is interrupted by one of the quartzite layers of Unit 3.

UNIT 5

White quartzite with intercalated thin layers of andalusite-biotite-muscovite schist overlies Unit 4 in the southwestern part of the archipelago. This unit (Unit 5) grades into a thick Unit 3 quartzite.

UNIT 6

Unit 6 comprises intercalated metagreywacke and schist. The schists contain porphyroblasts of staurolite and andalusite, in a matrix of biotite, muscovite, and quartz. Chlorite commonly occurs as a retrograde product of the biotite.

FIGURE 9.

A. Photomicrograph of ironstone, showing rhombic magnetite grains rimmed with with hematite alteration (white). Reflected light. Scale bar = 0.5 mm.

B. Photomicrograph showing ovoid hematite/carbonate grains within a calcareous pisolith. Note internal lamination of large grain in lower right. Plane-polarized light. Scale bar = 0.5 mm.

C. Polished slab of pisolitic ironstone. Note concentric internal lamination of pisoliths. Pisoliths in basal concentrated zone contain siltstone nuclei. Scale bar (top) = 1 cm.

D. Thin section of two calcareous pisoliths (left half and lower right quarter of photograph). Ovoid hematite/carbonate grains fill gap between, and define lamination in, the pisoliths. In upper right, small opaque grains of magnetite are dispersed in matrix of quartz (white) and carbonate (grey). Plane-polarized light. Scale bar = 0.5 mm.

E. Intercalated quartzite and schist of Unit 2, Outpost Islands. Andalusite porphyroblasts (bumps) and quartzite beds (ridges) display positive relief. Hammer in centre of photo for scale.

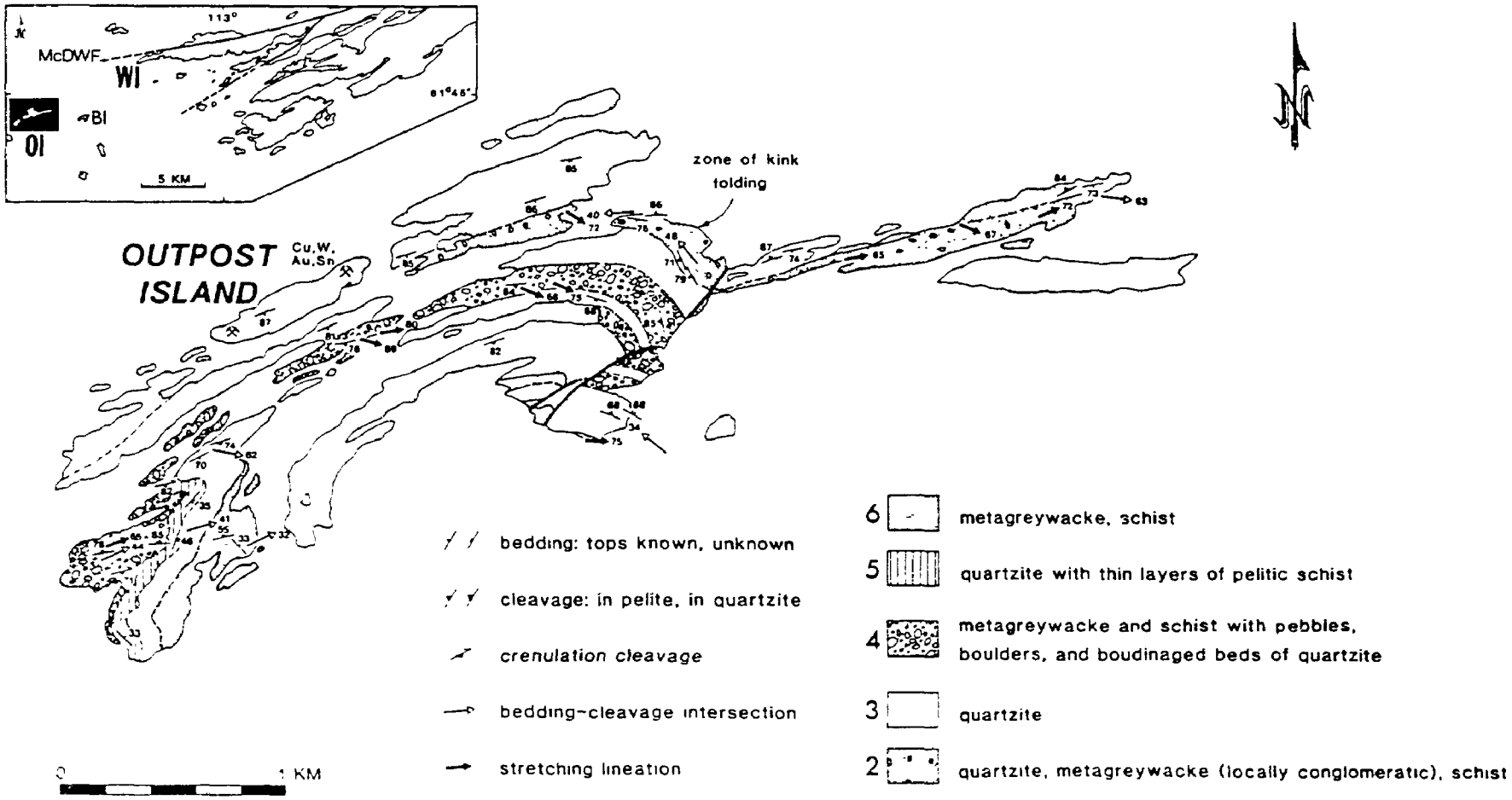
F. Tight fold in intercalated quartzite and pelite.

G. Incipient transposition of bedding in impure quartzite, Reinhardt formation.

H. Photomicrograph of felsic volcanic rock showing lenticular aggregate of feldspar phenocrysts, chlorite (dark grey), and fine-grained feldspar and quartz. Plane-polarized light. Scale bar = 1 mm.

FIGURE 10.

Geology of the northeastern archipelago of the Outpost Islands. Rock units on southwesternmost island display an asymmetric S-shaped fold pattern, which is overprinted by a large Z-shaped kink fold. Location shown in inset. BI: Butte Island; OI: Outpost Islands; McDWF: McDonald-Wilson Fault; WI: Wilson Island.


OUTPOST ISLAND
Cu,W, Au,Sn
zone of kink folding
McDWF
WI
BI
OI
113°
61°45'
5 KM
0
1 KM
bedding: tops known, unknown
cleavage: in pelite, in quartzite
crenulation cleavage
bedding-cleavage intersection
stretching lineation
fault: defined, approximate
6 metagreywacke, schist
5 quartzite with thin layers of pelitic schist
4 metagreywacke and schist with pebbles, boulders, and boudinaged beds of quartzite
3 quartzite
2 quartzite, metagreywacke (locally conglomeratic), schist
1 quartzite, conglomeratic quartzite

3. STRUCTURAL GEOLOGY

Tectonic structures in the Wilson Island Group are grouped into two general categories: (1) "main phase" structures (folds with associated foliation and extension lineation, defined by metamorphic minerals), and (2) "post-main phase" structures (folds and fabrics that deform the main phase structures)

Folds

Stratigraphic units of the Wilson Island Group typically strike northeastward and dip steeply to the southeast. The rocks of Wilson Island form the overturned northwest-facing limb of a major anticlinal flexure, or breached anticlinorium, that closes toward the east end of Blind bay and that here is termed the "Blind Bay anticline" (Maps 1A, 1B). The otherwise homoclinal character of the strata is disrupted locally by open, eastward-plunging inclined folds with a Z-shaped asymmetry. In contrast, open, eastward-plunging inclined folds with an S-shaped asymmetry occur in the Outpost Islands, where the general stratigraphic facing is to the southeast (Fig. 10).

Large-scale folds of Wilson Island Group rocks are best illustrated by a series of tight, overturned folds on the peninsula southwest of Basile Bay. A nearly isoclinal syncline with an amplitude of about 3 km and a wavelength of approximately 2 km is the largest and best exposed of these folds (hereinafter referred to as the "Basile Bay syncline"). This and other synclines have lobate forms, in contrast with the cuspate forms of the intervening anticlines in which carbonate rocks of the Five Snares assemblage are exposed.

For the most part, stratigraphic units have been thickened in the hinge areas of the folds, so that their geometry approximates that of similar folds. This is particularly evident in the relatively ductile pelitic units (e.g., Wbp, Map 2A) and carbonates (Wfd). Even the more competent quartzitic units show thickened hinges and/or thinned limbs (e.g., Wbiq in small syncline northwest of Basile Bay syncline, Map 2A), but in general these units display a more parallel fold geometry (especially Wbuq). Intercalated quartzitic and pelitic units in tight minor folds display disharmonic folds.

Axial surfaces of the folds strike north-northeast and dip 60-80 degrees southeast. The slightly arcuate nature of the axial surface traces (Map 2B) implies that the surfaces are curviplanar.

Fold axis orientations (Fig. 11) were determined from bedding-cleavage intersections. In most cases the intersections were calculated by plotting field measurements of bedding and cleavage on a stereonet. Bedding-cleavage intersection lineations were measured in the field where possible, such as around the Basile Bay syncline and in the Mudstone unit. Both calculated and measured intersection lineations are shown in Figure 11.

Most of the lineations trend eastward and plunge between 25 and 50 degrees. This is particularly true where the rocks are essentially homoclinal, as on the main part of Wilson Island and in the region between Basile Bay and Basile Lake. Near the hinges of the large and tight folds (especially the near-isoclinal Basile Bay syncline and the major anticlinal flexure on Wilson Island), the lineations plunge steeply and more southeasterly, parallel to the dip direction of the axial surfaces. These folds are therefore reclined (cf. Hobbs _et al._, 1976).

Southwest-plunging bedding-cleavage intersection lineations have been measured at the northeast end of Basile Lake, and mesoscopic folds with S-shaped geometries and gentle to moderate southwest plunges occur on islands south of Wilson Island. In localities where demonstrable refolding has occurred (discussed in a later section), the lineations display an almost random distribution.

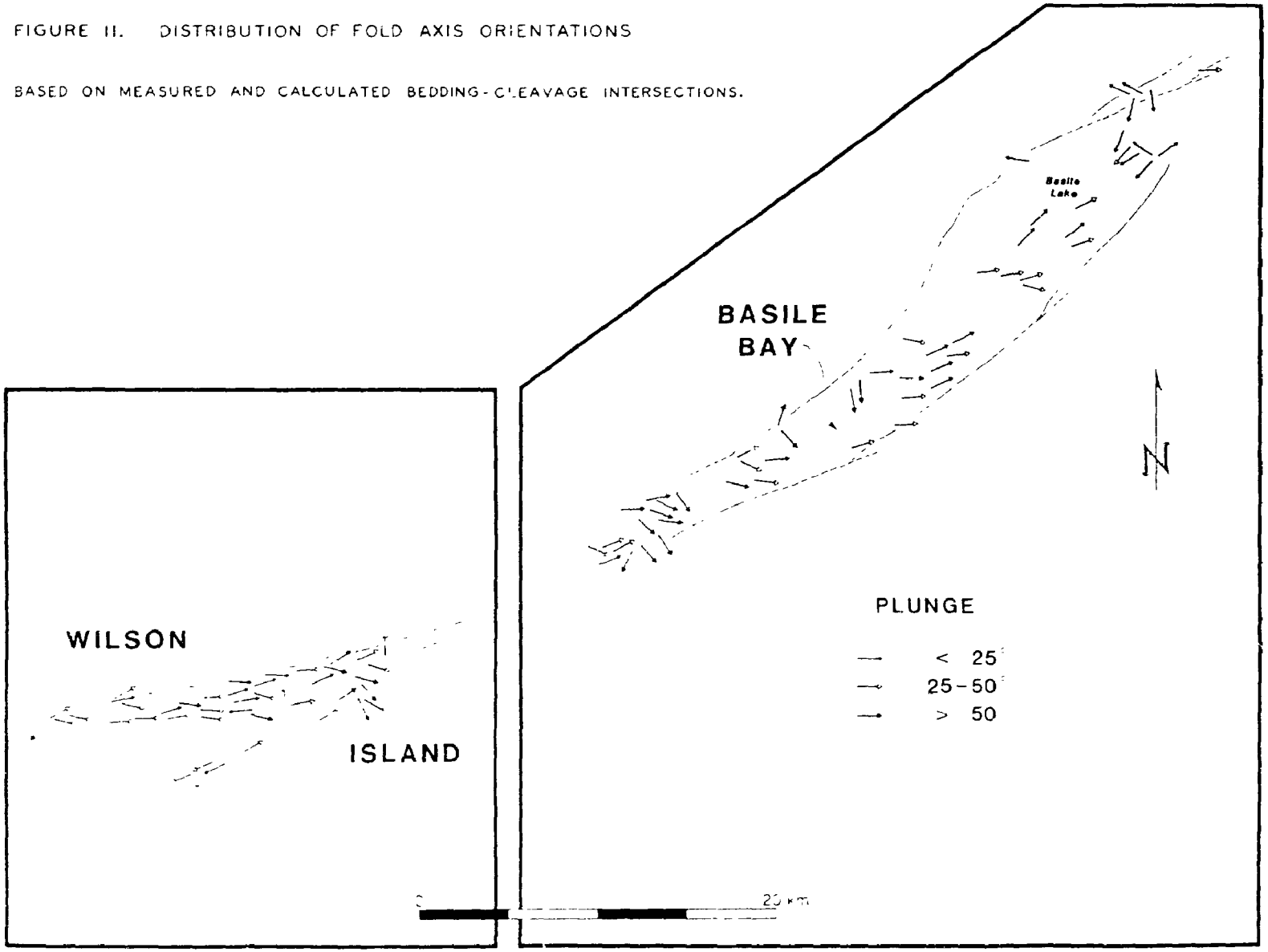
A spectacular tight fold is exposed on a small island northeast of Wilson Island, adjacent to Inconnu Channel (Fig. 9F). This fold displays the Z-shaped asymmetry typical of the other folds in northwest-facing successions. However, bedding along this fold is everywhere overturned, and the fold axis plunges gently to the southwest. Downward-facing folds also occur in the Dolomitic Sandstone member of the Safety Cove formation, near the west end of Wilson Island.

Foliation

Most rocks of the Wilson Island Group exhibit a southeast-dipping tectonic foliation (Maps 1B, 2B). In the greenschist-facies metapelitic rocks the foliation is an ubiquitous slaty cleavage that, in the homoclinal panels, consistently strikes northeast. This cleavage forms divergent fans in the thick pelitic units (e.g., Wbp) and in thin pelitic layers in the hinges of tight minor folds. In the hinge of the Basile Bay syncline, platy minerals in thin argillaceous layers of both the basal Lower Quartzite member and the basal Lower Pelitic member define a bedding-parallel cleavage. In general, however, the orientation of slaty cleavage in the hinge areas of the major folds is not well documented, because of the difficulty in distinguishing it from primary layering.

FIGURE 11. DISTRIBUTION OF FOLD AXIS ORIENTATIONS

BASED ON MEASURED AND CALCULATED BEDDING-CLEAVAGE INTERSECTIONS.



A few measurements in the Blind Bay anticline (and in parasitic folds on its northern limb) suggest that the slaty cleavage in that fold displays an approximately constant angle and vergence with respect to bedding, irrespective of the spatial relationship to the hinges of the folds. The same kind of relationship is exhibited by folded pelitic rocks of the Five Snares assemblage, and amphibolite-facies metapelites at the Outpost Islands display a strong schistosity that also "wraps around" fold hinges.

Quartzites in hinge areas of the main-phase folds near Basile Bay display a strong axial plane cleavage, commonly accentuated by evenly spaced micaceous films up to about 3 millimetres apart. Axial plane cleavage also is well developed in fine-grained micaceous sandstones that occupy mesoscopic folds of the Reinhardt islands. Bedding has been incipiently transposed by axial plane cleavage in tight, outcrop-scale folds (Fig 9G). Away from the folds, cleavage in quartzites is only locally well developed.

In conglomeratic units (especially the Blind Bay conglomerate member and the lapilli-ash tuffs of the Reinhardt formation and paraconglomerates of the Outpost Islands), the main foliation is manifested as a strong preferred orientation of flattened clasts, and of phyllosilicates which typically are abundant as matrix. Cleavage in the conglomerate and in quartzites displays convergent fanning (Wrc, Maps 1A, 1B; Wblq, Maps 2A, 2B). Cleavage refracts where it crosses contacts between quartzitic and pelitic layers.

The nature of the foliation is dependent on lithology: In silt-poor mudstones the slaty cleavage is essentially planar, and spaced on a microscopic scale; in argillaceous siltstones and micaceous quartzites, phyllosilicate films anastomose around parallel lenticular

quartzose domains that range in thickness from less than 1 millimetre to about 3 millimetres; in non-micaceous quartzites the cleavage is defined by preferred orientation of inequant quartz and feldspar grains, and is spaced on the scale of the grains. In thin section the quartz grains commonly are lenticular and have sutured contacts. Thin sections of dolomitic rocks show cleavage defined by preferred orientation of tabular to lenticular grains and polycrystalline aggregates of carbonate and quartz.

Tectonic foliation in the metavolcanic rocks is locally well-developed, especially near the margins of the felsic units. This foliation is distinguished from primary flow banding on the basis of three criteria: (1) where flow banding displays outcrop-scale folds and undulations (e.g., Fig. 3E), it is cross-cut by the tectonic fabric, (2) where the tectonic foliation is well developed, the rock is fissile, and (3) the tectonic foliation commonly is characterized by alternate light- and dark-coloured layers, the latter exhibiting lenticular aggregates of feldspar phenocrysts (commonly pulled apart along the foliation) and chlorite. Thin sections show that these aggregates (Fig. 9H) also contain recrystallized fine-grained feldspar and quartz, biotite/chlorite, epidote, and in some instances white mica, opaques (magnetite or pyrite?), and carbonate (Fig. 9H). In addition to the phenocryst aggregates, alternate parallel domains of (1) phyllosilicates, (2) microcrystalline quartz $\pm$ feldspar, and (3) "fine-grained" (ca 0.1 mm.) quartz $\pm$ feldspar, define the foliation. The "fine-grained" domains and the phenocryst aggregates typically are lenticular in section.

Schistosity in the amphibolite-facies units is characterized by

alternate parallel quartzose domains and domains of preferentially oriented micas. The micas wrap around porphyroblasts of andalusite and staurolite, which contain sigmoidal inclusion trails (Fig. 12A).

Stretching Lineation

Foliation surfaces commonly display a strong down-dip stretching lineation, which is expressed in the field by preferred orientation of inequant minerals, of streaks and lenticular domains of mineral aggregates, of elongate clasts, and of ovoid strained amygdules. The lineation trend ranges from south-southwest to east, and with rare exceptions, plunges between 50 and 90 degrees (Maps 1B, 2B). It typically is best developed where main-phase folds are prominent, and where the foliation is intense. For example:

(a) In the core of the Basile Bay syncline, spindle-shaped deformed quartz granules in the Upper Quartzite member of the Basile formation have aspect ratios (X:Z) of at least 10:1.

(b) In the easternmost Reinhardt islands (around longitude 113°W), stretched amygdules in basalts and elongate clasts both in basalt breccias and in sandy breccia-tuffs show a strong preferred orientation. Outcrop-scale folds of inconsistent vergence indicate the existence of larger folds in this area.

(c) Strongly foliated felsic volcanic rocks and micaceous quartzites typically display parallel streaks of mineral aggregates on foliation surfaces.

(d) In the Outpost Islands, the paraconglomeratic metagreywackes exhibit spectacular elongate boulders of quartzite, and elongate pressure shadows bordering ellipsoidal andalusite porphyroblasts define the stretching lineation in the schists (Fig. 12B).

On a microscopic scale the stretching lineation typically is expressed by pressure shadows and pressure fringes marginal to grains, and by pulled-apart grains. The pressure shadows on the margins of andalusite porphyroblasts are composed of quartz, biotite, and muscovite. Chlorite and sericite occur as reaction rims on the andalusite. Quartz grains in thin sections of lineated, greenschist-facies, micaceous quartzites show recrystallized margins that grade into pressure shadows composed of microcrystalline quartz and sericite (Fig. 12C). Pressure shadows marginal to feldspar grains are composed of microcrystalline feldspar, chlorite (or biotite in some instances), and sericite. Feldspar grains also exhibit composite pressure shadows, consisting of an internal core of chlorite or biotite and an external sheath of sericite $\pm$ quartz.

Euhedral crystals of metallic minerals with pressure fringes of quartz fibres are displayed in thin sections of strongly lineated felsic porphyries and phyllitic mudstones from Basile point, and of magnetite ironstone from east of Basile Bay. Pressure fringes in two samples are asymmetrical when viewed parallel to the stretching lineation: looking northeast, the asymmetry is S-shaped in one sample, and Z-shaped in the other. Numerous samples show symmetrical pressure fringes (Fig. 12D). The metallic grains commonly have been flattened, fragmented, and strung out along the stretching lineation. Quartz fibres have grown in the spaces between the pulled-apart fragments. Similar pull-apart structures are exhibited by feldspar phenocrysts in felsic volcanic rocks of the Reinhardt formation; thin sections show that the spaces between the feldspar fragments contain quartz and chlorite or carbonate.

Mylonitic Fabrics

In the Petitot Islands (Fig. 2A), a fault-bounded sliver of quartzites and biotite-chlorite schists display a strong northeast-striking foliation, and a gently plunging stretching lineation that ranges through the horizontal from southwest to northeast in trend. Shear bands indicate dextral transcurrent shear. In thin section, spindle-shaped quartz granules within the quartzite display about 100:1 reduction of grain size by extensive development of subgrains (Fig. 12E). Reinhardt (1969) suggested that these rocks belong to the Wilson Island Group.

Post-Main Phase Structures

On virtually all scales throughout the map area, the main foliation is deformed by kink folds and crenulations, and locally by crenulation cleavage.

The mudstones of the Basile formation are extensively deformed by kink folds, which range from centimetre-scale to map-scale. Axes of the kink folds range from subhorizontal to subvertical in orientation. The pattern of foliations in the Mudstone unit defines a major kink structure that plunges gently to the northeast and has an S-shaped asymmetry. Most of the outcrop-scale kinks in the Basile Bay area also have this geometry (Fig. 12F).

Phyllites and micaceous granulestones of the Five Snares assemblage also display extensive kink folds. Gently plunging kinks range from centimetre-scale to many metres in wavelength and, as in the mudstones, typically have an S-shaped geometry (viewed to the northeast); steeply plunging small-scale kinks commonly occur in conjugate

FIGURE 12.

A. Photomicrograph showing inclusion trails in staurolite porphyroblast. Plane-polarized light. Scale bar = 1 mm.

B. Foliation surface of andalusite-muscovite-biotite schist, Outpost Islands. Elongate pressure shadows on andalusite porphyroblasts define steep down-dip stretching lineation.

C. Photomicrograph of micaceous quartzite. Sericite in pressure shadows is attached to margins of quartz grains. Plane-polarized light. Scale bar = 0.5 mm.

D. Photomicrograph showing euhedral pyrite grain with symmetrical pressure fringes composed of quartz. Cross-polarized light. Pyrite grain is 0.25 mm across.

E. Photomicrograph of mylonitic quartzite, Petitot Islands. Stretched quartz granules exhibit numerous subgrains. Cross-polarized light. Scale bar = 0.5 mm.

F. Kink folds in fine-grained sedimentary units. Folds plunge to SE (left). Graded member of Basile formation, near Hornby Channel.

G. Photomicrograph showing crenulation cleavage in schist of Outpost Islands. Plane-polarized light. Scale bar = 1 mm.

H. Widely spaced incipient crenulation cleavage in intercalated granulestone and silty mudstone, Wilson Island.

FIGURE 3

sets.

In the Outpost Island , rock units have been folded by a large, steeply plunging, Z-shaped kink, producing a flattened horseshoe-shaped interference pattern (Fig. 10). Intercalated thin beds of quartzite and schistose metagreywacke (Unit 2) on the short limb of the kink display abundant outcrop-scale kink folds.

In the hinge of the Basile Bay syncline, argillaceous layers within the basal Basile formation (Lower Quartzite member) display chevron-shaped crenulations, with distinct subhorizontal axial planes spaced less than one millimetre apart.

Crenulations with about 3-15 centimetre wavelengths occur in feldspathic granulestones of the uppermost Reinhardt formation (Blind Bay conglomerate), in the hinge of a small anticline on the shore of Wilson Island (61°48'N, 112°57'30"W). Axial planes of the crenulations dip steeply to the southeast. Dextral slip has occurred along attenuated limbs of some of the crenulations. Chlorite and white mica have formed along these limbs, giving rise to a widely spaced, incipient, crenulation cleavage (Fig. 12H). At the same locality, phyllitic rocks of both the Reinhardt and the overlying Safety Cove formations display chevron kink folds up to a few centimetres in wavelength. Lineations formed by intersections of these kinks with the main cleavage plunge gently to the northeast. Intercalated phyllitic and granule units form metre-scale cuspate anticlines and lobate synclines that also plunge gently to the northeast.

In the Outpost Islands, a steeply dipping, east-northeast striking, crenulation cleavage disrupts the main schistosity in the hinges of folds (southeasternmost island of the archipelago, Fig. 10). In thin section, the crenulation cleavage is defined by parallel muscovite

and biotite that cut across kinked micas of the older schistosity (Fig. 12G). Porphyroblasts such as andalusite, also locally bent, are confined to the lithons between crenulation cleavage foliae. The andalusite porphyroblasts have grown over the older fabric.

Some pelitic rocks, both of the Basile and the Reinhardt formations, display a weak second cleavage. Thin sections of these rocks show a preferred orientation of scattered "oversized" plates of chlorite, or of biotite sheaths around cores of white mica, that cut across the main cleavage.

Faults

The McDonald-Wilson Fault (Hoffman *et al.*, 1977) is a major high-angle fault that extends east-northeastward, from along the north shore of Wilson Island to beyond the north end of McDonald Lake, where it merges with the McDonald Fault (Figs. 1, 2A). An estimated 65-80 kilometres of dextral transcurrent movement has taken place along the McDonald-Wilson Fault (Hoffman, 1981), resulting in separation of the Wilson Island and Basile Bay segments of the Wilson Island Terrane (Hoffman, 1985), and juxtaposing rocks of the Wilson Island Group against conglomerates of the younger Murky Formation (Et-Then Group). South of the Basile Bay area, the Murky Formation forms a thin veneer, resting unconformably on Archean crystalline rocks of the Simpson Islands Terrane (Hoffman, 1985). The Murky Formation contains boulders derived from the Wilson Island Group.

Two northeast-trending faults separate the Wilson Island Terrane from rocks of the Great Slave Supergroup. One of these faults, called the Inconnu Thrust (Hoffman, 1981), is expressed surficially as a

valley that separates rocks of the Reinhardt formation (Wilson Island Group) from sandstones of the Hornby Channel Formation (Sosan Group). The Hornby Channel Formation is underlain to the south by Archean crystalline rocks of the Simpson Islands Terrane (Hoffman et al., 1977; Hoffman, 1985). The other main bounding fault, which extends from Keith Island to Murky Channel and juxtaposes rocks of the Wilson Island and Kahochella Groups, is here called the "Basile Bay fault." On the north side of Basile point (Map 2B), the Basile Bay fault dips steeply to the southeast. The Wilson Island Terrane has internally been dissected by high-angle faults that delineate two distinct orientations: (1) northeast to north-northeast, and (2) northwest to west.

The first set includes numerous faults in the Basile point area, and a narrow system of faults that extends northeastward for about 8 kilometres across the main part of Wilson Island (straddling longitude 113°W). One of these faults on Wilson Island dips 60 degrees to the southeast and is exposed for over 100 metres along strike. Most of the faults that belong to the first set display right-lateral separation of strata, up to a maximum of 800 metres. A notable exception to this relationship is a pair of faults that cut across the northwest flank of the Blind Bay anticline on Wilson Island. These faults display an estimated net separation of 1500 metres in a sinistral sense, and they apparently eliminate part of the Reinhardt section. The fault with the greatest separation, which is called here the "Blind Bay fault," probably extends southwestward along Blind bay (Maps 1A, 1B). A few minor faults of north-northeasterly orientation also show left-lateral separation.

A series of faults that strike northwest to west across Wilson Island represents the second set. On the main part of the island these

faults are exposed in cross section where they cut through hills formed by the basal part of the Wsds unit, Map 1A); they dip steeply to the southwest and separate strata in an apparent dextral sense. Both dextral and sinistral apparent offsets have been produced by faults of this set on the southern part of Wilson Island and around the foot-shaped bay at the east end of Basile Lake. There are three different ways in which the northwest-trending faults terminate: (1) truncation by faults of the northeast-trending set, (2) by merging and splaying, and (3) by dying out into folds and numerous small faults, each with little displacement. All three kinds of terminations are represented among faults on the main part of Wilson Island.

Quartz stockworks and associated hematite commonly occur in quartzites proximal to faults. Along the McDonald-Wilson Fault, mixed carbonate and sandstone assemblages commonly have been brecciated and altered; carbonate rocks have locally been silicified, and sandstones have been partly replaced by carbonate. Adjacent to the Basile Bay fault, rocks of the Basile formation have locally been brecciated. Along a major west-trending fault in the hinge area of the Blind Bay anticline (here named the "Scimitar fault," Map 1B), felsic volcanic rocks have been brecciated and chloritized, and sandstones of the Wro unit have a bleached (silicified) appearance.

4. DEPOSITION OF THE WILSON ISLAND GROUP

REINHARDT FORMATION

Extensive mafic volcanism is recorded by the basalts of the Lower Igneous Complex, the oldest exposed unit of the Wilson Island Group. Whether the basaltic lavas were erupted from fissures or from shield volcanoes is unknown. Physical features of the basalts give no conclusive information about the environment of deposition. However, because of the intercalated mudstones, the flows are presumed to have been deposited subaqueously or periodically inundated by water.

The silty mudstones probably were derived in part from mafic or intermediate ash, as suggested by their high iron content (abundant biotite) and by their association with lapilli tuffs. Deposition of the ash along with fine-grained epiclastic sediment probably took place in lakes or a low energy sea, producing the planar-laminated sedimentary units.

The thick units of felsic porphyry are interpreted as extrusive domes and flows. The lateral tapering of these units is consistent with the morphology that would be expected for domes or clusters of domes formed by the extrusion of highly viscous siliceous lava. Flow banding and abundant spherulites in many of the rhyolitic rocks are indicative of rapid cooling. In many places, the felsic domes are adjacent to trough-crossbedded pebbly arkoses that probably are of fluvial origin (discussed below). These domes are therefore likely to have formed at least in part subaerially.

Although no geochemical data are available, field and thin section observations strongly suggest that a significant depositional gap

exists between the felsic and mafic igneous rocks. In contrast to the basalts, the felsic rocks probably range from rhyolite to dacite in composition. The only rocks of possible intermediate composition are the dark grey plagioclase-phyric intrusions, the black "mafic" dikes of the Lower Igneous Complex, and the "mafic" border phases of zoned dikes, which all contain modal quartz. However, these rocks collectively form only about 1 percent of the total igneous component of the Reinhardt formation (e.g., Figure A).

In the Lower Igneous Complex, the occurrence of enclaves, both of felsic porphyry in the black dikes and vice versa, is suggestive of mechanical mixing of mafic and felsic liquids. Mafic inclusions in felsic porphyries throughout the Reinhardt assemblage probably represent quenched blobs of mafic magma, analogous to those in Cenozoic rocks described by Bacon (1986) and by Blake et al. (1965).

Possible coevality of two magmas of contrasting composition is also suggested by field relations between felsic porphyries and the dark grey plagioclase-phyric dikes: the more mafic magma in some cases may have first intruded the sedimentary country rocks, in turn to be intruded by the more siliceous magma (possibly while the former was still molten), thus forming a mafic "sheath" between the siliceous dike and the country rocks. This process has been proposed by Blake et al. (1965) for the origin of composite Tertiary dikes in Scotland and Ireland. Sharp contacts may have been produced by chilling of the mafic material against the silicic magma as the latter was emplaced; where contacts are gradational, some mixing of the two phases apparently took place before cooling.

Composite units that show only one chilled mafic margin per component subunit (Fig. 4) may be parts of a feeder conduit that was repeat-

edly injected with magma. The compositional zonation of these units and of dikes like the one shown in Figure 3G may be either an artifact of coexisting bimodal magmas or the result of magmatic differentiation.

With rare exceptions, compositionally bimodal volcanic suites are generated in areas underlain either by extending continental crust or by thick oceanic crust (Hildreth, 1981). Granitoid pebbles, coarse polycrystalline quartz of probable plutonic origin, and abundant fresh potassium feldspar in the arenaceous rocks of the Reinhardt formation indicate a continental provenance. Proximity of the depositional basin to the source is implied by the low maturity of the arenites and conglomerates. The Reinhardt formation therefore must have been deposited on, or adjacent to, continental crust, and hence an extensional tectonic regime probably was in effect during deposition.

The sedimentary units exhibit many characteristics which collectively are suggestive, although not diagnostic, of deposition by a braided fluvial system. Fining-upward cycles dominated by trough-crossbedded sandstones have been well documented in ancient fluvial successions (e.g., Allen, 1965; Cant and Walker, 1976). The scoured bases of fining-upward cycles in the Reinhardt formation are interpreted as the floors of channels, croded during high stages of floods. Thin pebble lags, dispersed pebbles, and both normal and reverse graded bedding in the arenites may reflect fluctuations in the energy of the transporting medium. Trough crossbedding is attributed to the migration of dunes. Planar- and ripple-laminated sandstone and argillaceous siltstone at the tops of fining-upward sequences are similar to the Fl lithofacies of Miall (1977), and to facies F of Cant and Walker (1976). These fine-grained units probably represent sedimentation during the

falling stages of floods.

This assemblage of features is characteristic of distal sandy braided rivers and alluvial plains (Rust, 1978). The preservation of fining-upward cycles could be due to the aggradation of river beds during the waning stages of floods. Before the Late Paleozoic, alluvial plains probably were extensive and subject to frequent flooding (Schumm, 1968; Rust, 1978). The lateral continuity of epiclastic units of the Reinhardt formation may therefore favour an alluvial plain origin, as depicted in Figure 13A.

Poorly sorted massive conglomerate units, such as the basal Pebbly Sandstone member and layers of conglomerate and granulestone within the Blind Bay conglomerate, are inferred to be analogous to the Gm lithofacies of Miall (1977). Such conglomerates are produced by the vertical accretion of longitudinal bars, a process which is especially common in the gravelly proximal reaches of braided rivers and alluvial plains (Rust, 1978; Rust and Koster, 1984). The association of such gravelly deposits with trough-crossbedded sandstones is consistent with the model presented by Miall (1978) for cyclic deposits of sandy braided rivers (South Saskatchewan type).

A fluvial origin is difficult to demonstrate without adequate paleocurrent data, particularly because of the ambiguity of many of the sedimentary structures. For example, trough crossbedding and graded bedding may form in a variety of sedimentary environments, as summarized by Reineck and Singh (1980). Reactivation surfaces, common in the ripple-laminated units, may develop in a fluvial regime during falling stage (Collinson, 1970), although they also are common in tidal sediments (Klein, 1970b). Some of the fine-grained units display wavy and lenticular bedding, which are characteristic of tidal settings (Reineck

basaltic and rhyolitic glass.

Grains of sand and fragments of sandstone and basalt could have been incorporated into eruptions as accidental ejecta, if the strata around the vent and feeder pipe included layers of sandstone (in part poorly consolidated) and basalt. Alternatively, volcaniclastic debris flows could have entrained detrital sand (and mud) as they passed through the depositional basin from steep slopes of volcanoes, or mixed pyroclastic-epiclastic debris flows could have been initiated by catastrophic flooding within a sandy alluvial basin that was choked with unconsolidated pyroclastic debris.

Clasts of sandy lapilli-ash tuff within layers of the same are interpreted as chunks of older, dissected, debris-flow deposits that were entrained by subsequent flows. The complex intercalation of lapilli-ash tuffs with arenites in Submember 1 of the Lower Clastic member is inferred to represent dissection of the debris-flow deposits by stream channels, and subsequent deposition of sand within these channels. The lenticular arenite units may be channels that were filled by vertical accretion, and then buried by the next debris flow.

Convex-downward deformed enclaves of arenite within tuffaceous units probably are isolated ball-and-pillow structures, that formed after sand was deposited over hydroplastic debris-flow units. The abundance of soft-sediment deformation structures in clastic units throughout the Reinhardt succession is indicative of shock from recurrent tectonic activity (including volcanism and probable syndepositional faulting) and is suggestive of rapid deposition.

and Wunderlich, 1968), although similar units have been reported to form the upper parts of fluvial fining-upward successions (Cant and Walker, 1978). Fining-upward cycles have also been reported in tidal sandstones (Klein, 1970a). There is thus a possibility that fluvial environments, which probably were responsible for most of the sedimentation, interacted with either marginal marine (e.g., tidal flat or deltaic) or lacustrine depositional systems to produce the sedimentary units of the Reinhardt formation.

Laterally extensive vertical facies changes from sandstone to conglomerate characterize the boundaries between certain stratigraphic units (i.e., Lower Clastic member - Pebbly Sandstone member; Unnamed Sandstone unit - Upper Pebbly unit) and are displayed repeatedly throughout the Blind Bay conglomerate. Such facies boundaries may be the manifestation of abrupt temporal changes from sand-dominated to gravel-dominated sedimentation, which in the proposed alluvial plain setting probably were caused by periodic high-energy floods. Alternatively, these changes may represent progradation of the distal portions of gravelly alluvial fans out onto the alluvial plain. However, other lithofacies that are characteristic of alluvial fan assemblages, such as planar-crossbedded arenites and debris-flow paraconglomerates (Rust and Koster, 1984) are not present in the Reinhardt section. In any case, widespread deposition of coarse detritus must have been a response to periodic rejuvenation of the system, due to uplift of the source area or to subsidence of the basin.

Syndepositional uplift marginal to the active locus of deposition is implied by the occurrence of intrabasinal clasts, which are especially abundant in the Blind Bay conglomerate. Although the rhyolite

clasts may have been derived by the erosion of subaerial domes, pebbles of quartzitic arkose and subarkose presumably were derived from lower parts of the succession. Introduction of these clasts into the depositional system would have required erosional unroofing of lower stratigraphic levels. Normal faulting is a likely mechanism for such unroofing, especially within the proposed extensional stress regime.

The lapilli-ash tuffs are interpreted as pyroclastic or mixed pyroclastic-epiclastic debris flows, perhaps triggered by explosive magmatic or phreatomagmatic eruptions. They are poorly sorted, are matrix-supported, and lack internal stratification, all characteristics typical of pyroclastic debris flows as outlined by Fisher (1982) and by Smith (1986).

The irregular-shaped fragments of biotite schist morphologically resemble pyroclastic fiamme. In composition they are similar to the mafic or intermediate inclusions that are common in many of the felsic dikes. They probably represent blobs of mafic or intermediate magma that were ejected as tephra, and consequently cooled to form glass. The fine-grained matrix is inferred to have been mainly ash of fairly siliceous composition (probably dacite).

The coarse sand-sized quartz and feldspar grains may be a mixture of epiclastic detritus and broken phenocrysts. Most of the quartz is coarsely polycrystalline, indicating a non-volcanic source, and is thus either epiclastic or accidental. Siliceous and mafic pyroclastic material therefore apparently erupted simultaneously. The juvenile products of such eruptions may have been analogous to modern pyroclastic deposits from Askja, Iceland, described by Sigurdsson and Sparks (1981); deposits of predominantly rhyolite pumice from Askja contain clasts of dark-coloured pumice that is composed of intimately mixed

basaltic and rhyolitic glass.

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SAFETY COVE FORMATION

The Safety Cove formation exemplifies a dilemma familiar to Precambrian sedimentologists: the lack of fossil control and the lack of sedimentary structures diagnostic of a specific depositional environment.

Yeo (1976a, b) proposed that the unimodal paleocurrent patterns, and the abundance of trough crossbeds from which he obtained them, favoured a braided alluvial origin for the sandstones. Certainly these and other characteristics of the sandstones are compatable with this contention. The trough- and planar-crossbedded units suggest deposition by migration of, respectively, dunes and small-scale transverse bars in fluvial channels, analogous to (respectively) the St and Sp lithofacies of Miall (1977). Sandstones that contain apparent low-angle crossbedding may be analogous to facies G of Cant and Walker (1976), which they inferred to represent vertical accretion of sand in non-channellized (overbank) areas during flood stage. Low-angle crossbedding in braided alluvial sediments has also been attributed to the formation of crevasse splays (Miall, 1978) and to the infilling of low-relief scours under conditions of shallow, high-velocity flow (Rust, 1978). Such scours could have formed extensively on flood plains by erosion of low-amplitude bedforms during falling stage (cf. Collinson, 1970). The planar-laminated sandstones also suggest possible upper flow regime sedimentation in shallow water.

However, similar assemblages of sedimentary structures have been described from shallow-marine sandstones (e.g., Anderton, 1976). Moreover, unimodal paleocurrent patterns can be produced in shallow-marine settings if one phase of the tidal cycle is dominant (Klein, 1970b),

especially when processes related to a dominant ebb tide are enhanced by storm surge (Anderton, 1976).

Lenticular bedding, which is common in the upper half of the Dolomitic Sandstone member, indicates alternate periods of current and still water (Reineck, 1960), for which tidal environments are favoured (Reineck and Wunderlich, 1968). The sandstone-mudstone couplets resemble the "tidal bedding" (cf. Wunderlich, 1970), ascribed by Reineck and Wunderlich (1967) to cyclic tidal sedimentation. However, storms have been credited for the formation of both lenticular bedding (Anderton, 1976) and sandstone-mudstone rhythmites (Reineck and Singh, 1972), and the stratigraphic relationships of these units in the Dolomitic Sandstone do not preclude this mode of origin. Units consisting of thin intercalations of sand and mud can also form in alluvial successions (Cant and Walker, 1978), but such units cap fining-upward cycles.

Limited paleocurrent data from trough crossbeds in the Lower Arkose member are in agreement with the southwesterly mode reported by Yeo (1976a). In the Dolomitic Sandstone member, although one predominant paleocurrent direction is indicated by crossbedding in the sandstones, a few examples of interference ripples and of diametrically opposed current-ripple marks indicate that sediment was reworked by other currents. Of greater importance are thick sets of planar crossbeds in sandy dolostones that have yielded a northeasterly paleocurrent direction, in approximate diametric opposition to the regional mode. If the regional mode represents a southwest paleoslope, which seems reasonable, then some other factor is required to explain these other crossbeds. They may have formed as a result of the migration of large-scale bedforms, such as bars or large sandwaves, in a marine environment due to ocean currents. Interpretation of paleocurrent data for

the Wilson Island Group is risky, however, because the data are scarce and the post-depositional deformation history is poorly constrained (Chapter 5).

All that can be inferred with some degree of certainty is that a body of water that was shallow and warm enough to permit carbonate production, and probably large enough that its sedimentary processes could be affected by storms or tides, served as a depocentre for part of the Safety Cove formation. A possible depositional setting would include a coastal alluvial braidplain system, connected by distributaries or tidal channels to a shallow marine platform, a shallow epeiric sea, or a playa lake. Accepting the limited evidence for marine processes outlined above, this interpretation is presented in terms of a coastal plain / shallow marine platform model, as illustrated in Figure 13B.

The Lower Arkose member is interpreted as a distal braidplain succession, because (1) the attributes of the quartzose sandstones that make up the member can be readily explained in terms of alluvial processes, (2) the sandstones show no features that are strongly suggestive of marine deposition, and (3) they resemble sandstones of the conformably subjacent Blind Bay conglomerate, which is of inferred alluvial braidplain origin. The sandstones are somewhat more mature than typical sandstones of the Reinhardt formation, and therefore were transported farther before deposition.

The Dolomitic Sandstone member testifies to the complex interaction of siliciclastic and carbonate depositional processes on, and marginal to, the shallow platform. The superposition of the basal carbonate units on sandstones of the Lower Arkose member is inferred to

record transgression of the sea over the alluvial plain. Carbonate could have formed on tidal flats that were isolated from terrigenous depocentres, and on distal parts of the offshore platform. During times of no significant siliciclastic sedimentation, carbonate production would accelerate both marginal to the sea and offshore, perhaps covering vast portions of the platform (Fig. 13B). The locus of siliciclastic sedimentation would prograde during times of voluminous terrigenous influx. Consequently, nearshore carbonate production would cease, and the belt of active offshore carbonate sedimentation either would be forced to shift seaward or would become entirely inactive. The facies belts would also shift laterally in response to fluctuations in sea level, climate, and the rate of subsidence.

The quartzose and carbonate-cemented sandstones of the Dolomitic Sandstone member (and of the similar Upper Mixed unit) may represent a combination of tidal channels, distal reaches of the inferred alluvial braidplain, and lower intertidal sand flats or subtidal bars. Argillaceous sandstone units may represent intertidal flats or subtidal storm deposits. The laminated dolostones resemble cryptalgal laminites, and may have originated as carbonate detritus that accumulated on upper intertidal flats, with the aid of algal entrapment.

Periodic storms or spring tides could have redistributed siliciclastic mud and silt into areas of active carbonate sedimentation, thus forming the argillaceous dolostones. Sandy quartzose layers that characterize the sandy dolostones may also be of storm origin. Numerous ancient examples of interlaminated carbonate and siliciclastic rocks have been interpreted as storm deposits (Fairchild, 1980; Kreisa, 1981; Tucker, 1982).

The dolomitic sandstone units (commonly crossbedded) probably were

deposited in tidal channels or shallow subtidal zones, where migrating dunes and bars were stable bedforms. Mixing of siliciclastic and carbonate material, in part reworked from upper tidal flats, could have readily occurred in these areas during floods or storms. Other Precambrian sandstones that consist of intimately mixed carbonate and quartz grains have been attributed to deposition in subtidal to lower intertidal areas, adjacent to carbonate tidal flats. Examples are Member 3 of the Bonahaven Formation (Dalradian Supergroup) of Islay, Scotland (Fairchild, 1980), and the upper Black Reef Formation (Transvaal Supergroup) of northern Cape Province, South Africa (Beukes, 1977). In the Dalradian example, mixing was enhanced by fluvial input and by storms (Fairchild, op cit.).

High-energy processes such as floods or storms could also account for (1) the common occurrence of pebbles in the dolomitic sandstone lithofacies, (2) pebbly layers in other units, and (3) the erosional surfaces on which some of the pebbly units were deposited. The resemblance of pebbles to rock types in lower parts of the Wilson Island Group suggests that lower stratigraphic levels were uplifted and unroofed, and hence a tectonic control of the terrigenous sediment supply (and of the basin morphology) appears likely. The abundance of soft-sediment deformation structures in the Dolomitic Sandstone member is consistent with this possibility.

The Middle Arkose member may represent progradation of an alluvial system, wherein the distal reaches were characterized by deposition in relatively deep rivers (thus creating trough crossbedding), and the more proximal facies were deposited by shallow, unconfined, sheet floods (hence low-angle crossbedding and planar lamination). The over-

lying Sandy Dolostone member is probably of shallow-marine origin, implying that the sea transgressed following deposition of the Middle Arkose member. Alternatively, the Middle Arkose member may be interpreted as an emergent subtidal to lower-intertidal sand flat, that was in turn buried by a prograding upper-intertidal carbonate flat (Sandy Dolostone member). Similar interpretations have been offered for siliciclastic to carbonate facies transitions in the Early Proterozoic Houtenbek Formation (Pretoria Supergroup) of the Transvaal, South Africa (Button and Vos, 1977), and for smaller siliclastic-carbonate cycles in the Black Reef Formation (Beukes, 1977).

FIVE SNARES ASSEMBLAGE

For reasons outlined previously, the Five Snares assemblage at Basile point is probably in part correlative with the Safety Cove formation, and hence the same general paleogeographic scenario is applicable to both units.

In keeping with Walther's Law, the argillaceous unit of the Five Snares assemblage is inferred to represent the distal, subtidal, fringe of the coastal siliciclastic belt; this would have separated areas of nearshore quartzose sand deposition (represented by the arkose unit) from a subtidal carbonate belt (dolostone unit). The assemblage is thus interpreted as an overall transgressive succession.

A possible modern analogue is the Arlington Reef Complex, located in the north-central part of the Great Barrier Reef. As reported by Maxwell and Swinchatt (1970), a thin belt of siliciclastic sands along the Queensland coast grades into a seaward-thinning veneer of terrigenous muds that intermingle with offshore carbonate sands and muds of the Arlington Complex. Relic quartzose sands lie beneath and among these

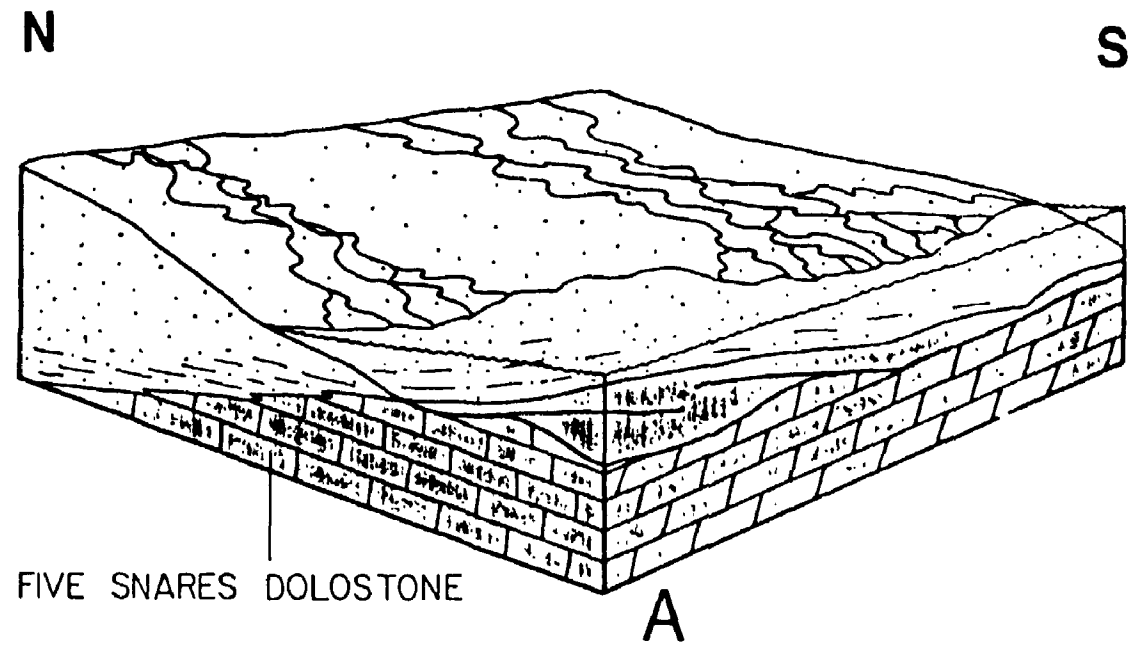
modern sediments. Carbonate sands of the Arlington Complex locally contain abundant quartz grains derived from reworked relic sands.

Carbonates of the Five Snares assemblage may have formed banks or other buildups, with sloping sides on which frequent slumping could occur. However, slumping may occur on slopes as low as one degree if triggered by strong earthquakes (Lewis, 1971). The complex internal deformation displayed within the slump sheets indicates that differential strain rates existed within each moving sheet.

BASILE FORMATION

The Lower Quartzite member probably represents a siliciclastic wedge that prograded over the platform (Fig. 14A) in response to an increased supply of terrigenous sediment, or to lowering of the sea level. The complex intercalation of siliciclastic and carbonate rocks in the northeastern part of the Basile Bay map area indicates that a gradual change in sedimentation conditions took place there at the beginning of Lower Quartzite deposition. The sharp contact between the Five Snares dolostone and the basal Lower Quartzite member in the southwest indicates that the change in this area was more abrupt. This relationship suggests that progradation was diachronous; the locus of initial siliciclastic influx probably was closer to the southwest part of the map area, and it later spread laterally.

The rocks of the basal Lower Quartzite member resemble rocks of the Dolomitic Sandstone member (Safety Cove formation), and by analogy may represent similar environments. The presence of rare desiccation cracks and mud rip-ups implies at least temporary subaerial exposure. Bidirectional crossbed orientations in some beds of sandy dolostone



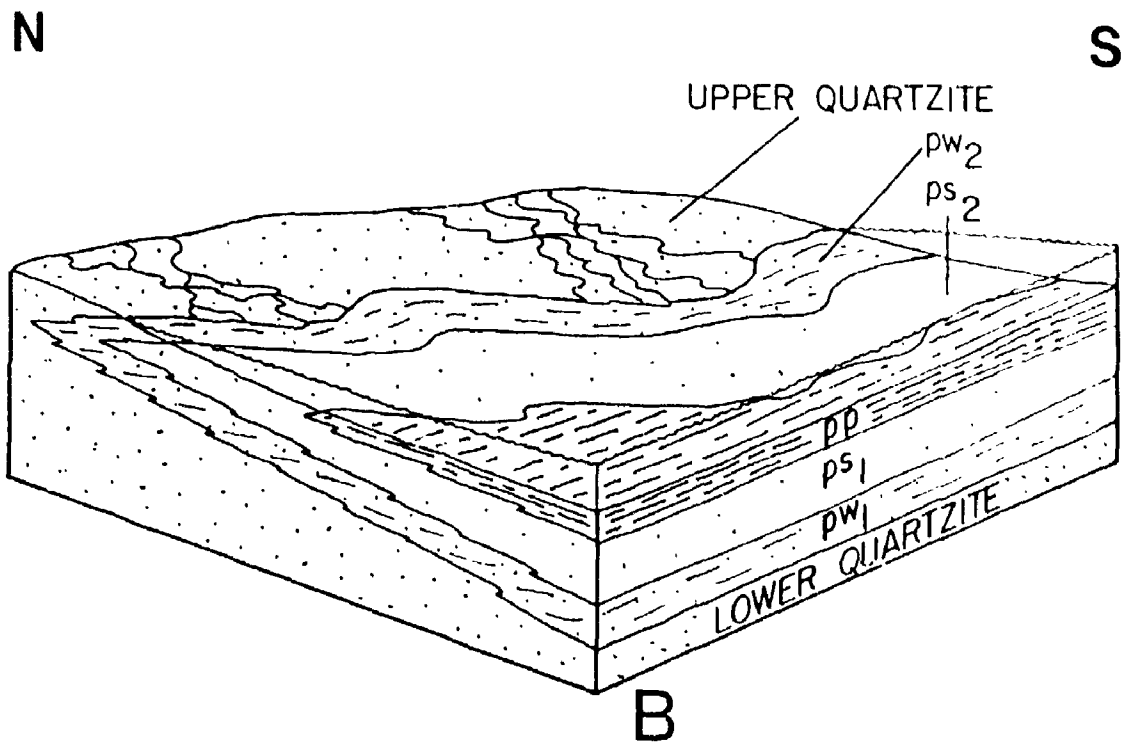


FIGURE 14. PALEOGEOGRAPHIC MODELS: BASILE FORMATION.

(A) Deposition of the Lower Quartzite member. (B) Beginning of regressive phase, Lower Pelitic and Upper Quartzite members.

suggest a tidal influence; this possibility is strengthened by the presence of wavy bedded units. Initial deposition of the Lower Quartzite member therefore probably occurred in an intertidal setting.

It is postulated that the thick sandstones which form the bulk of the Lower Quartzite member mainly were deposited on a prograding alluvial braidplain, although possibly they are at least in part of shallow-marine origin. Abundant trough and planar crossbedding are characteristic (although not diagnostic) of cyclic successions deposited by sand-dominated braided river systems (Miall, 1978), especially in the distal reaches of such systems (Rust, 1978). A modern example of this type of alluvial system is the South Saskatchewan River, in which stacked cosets of trough and planar cross strata have formed by the migration of dunes and sandbars, respectively (Cant and Walker, 1978). The South Saskatchewan River also contains cross-channel bars, which build laterally to form cosets of planar crossbeds that dip at high angles to the downstream direction (Cant and Walker, op cit.). In two-dimensional exposures, this kind of crossbedding could appear to dip in diametric opposition to the mode. Therefore, although the apparent northeast-dipping foresets of the Lower Quartzite member could have formed in an intertidal setting, they may represent cross-channel bars in a fluvial environment.

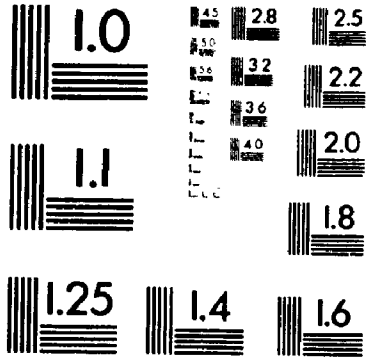
The tops of crossbedded units are ornamented with planar and small-ripple laminations and graded bedding, suggesting that periods of sedimentation began in water that was deep enough for the migration of dunes and bars, and ended with the water depth and competence decreasing. This relationship is consistent with flood cycles in an alluvial regime.

Not all characteristics of the Lower Quartzite member can be explained by analogy to a modern alluvial system. The absence of terrestrial vegetation in the Precambrian must have severely reduced the tendency of river banks to become stable, and thus material transported in Precambrian alluvial systems could have been largely unconfined by channels (Schumm, 1968). Given these conditions, a Precambrian alluvial system could have deposited extensive tabular sand bodies. Moreover, if the system were competent to erode fine- or medium-grained sand but not clay (i.e., clay layers that cap the deposits of the previous flood cycle and thus form the substrate), then the bases of such units could be essentially non-erosional.

The Lower Pelitic member displays an almost symmetrical arrangement of lithofacies, with argillaceous siltstone in the middle. The basal and upper parts of the member (subdivisions pw_1 and pw_2) consist mainly of wavy and lenticular bedded units. Thick cosets of these types of bedding (pw_1 is about 20 m thick) are characteristic of tidal settings (Reineck and Singh, 1980, p. 117). It is inferred that these transitional subdivisions of the member were deposited in intertidal mixed-flat settings, where both sand and mud were available. The grey sandstones (ps_1 and ps_2) and argillaceous siltstones (pp) probably were deposited in lower intertidal and subtidal zones, respectively. Thus, the Lower Pelitic member is interpreted as a transgressive-to-regressive oscillation (Fig. 14B).

The Upper Quartzite member is interpreted as a progradational alluvial succession, implying continued regression. Planar laminations and poorly sorted quartz granule layers near the base of the member suggest that sand was deposited by shallow-water sheet floods as it gradually buried former tidal flats. Trough crossbedding in overlying

3



layers would reflect deeper-water deposition.

Above the Upper Quartzite member, the Basile formation displays an overall upward-fining succession which is inferred to record a final(?) transgression.

Processes of carbonate deposition and diagenesis may have played significant roles in formation of the ironstones. The small, ovoid, carbonate-hematite grains probably are ooids, intraclasts, and/or small oncoliths, that were formed in somewhat agitated water. If these grains are indeed detrital, then the calcareous pisoliths must have formed above the sediment-water interface, because the carbonate-hematite grains form concentric laminae within the pisoliths and a graded veneer (geopetal lag?) on top of the pisolite layer. The pisoliths may therefore have originated by accretion and mechanical deposition in agitated water. Alternatively, they may have formed by in situ growth in the vadose zone. The rhombic magnetite grains probably are pseudomorphs after euhedral carbonate crystals. Diagenetic oxidation of primary ferruginous carbonate (e.g., siderite, ankerite) has been credited with the formation of rhombic magnetite pseudomorphs in Proterozoic iron formations of the Lake Superior region (Lougheed, 1983).

Lougheed (1983) has suggested that Proterozoic cherty iron formations were deposited as carbonates in tidal environments; the carbonate was extensively replaced by silica during diagenesis, and iron oxides were formed by diagenesis in intertidal settings, where oxygen was produced by algal photosynthesis. Perhaps the ironstones of the Basile formation were formed in a similar manner, although the original sediments may have been mixed carbonate-siliciclastic sands.

The grey feldspathic sandstones that constitute most of the Ironstone member may have been derived from relic terrigenous sands that

were submerged during this phase of shallow-sea deposition. The grey sandstones probably were deposited in lower intertidal or subtidal environments.

The graded units which typify the Graded member resemble partial and complete ABDE or BDE Bouma sequences. They are interpreted as fine-grained turbidites, perhaps deposited in a prodelta setting. This delta may have been the inferred "clastic wedge" that was built during earlier progradational phases of the Basile formation, and much of the underlying Basile strata may have been deposited in deltaic subenvironments. Transgressions coupled with decreased clastic influx would have led to submergence of the delta and would have allowed tidal processes to dominate (e.g., Lower Pelitic member?). The Graded member could thus represent a transgressive facies shift, from intertidal delta-front to subtidal prodelta and delta-foreslope sedimentation, in response to a rise of sea level or to subsidence of the basin.

Continued transgression is indicated by gradation of the member into the overlying laminated sulfidic Mudstone unit, which probably was deposited in deep water with poor circulation. Similar mudstones are intercalated with the "turbidite" unit to the southwest (Hoffman, 1977), which raises an alternative possibility that the Mudstone unit and the Graded member (also of probable turbidite origin) are distal equivalents of the "turbidite" unit. This would imply that deepening of the basin was accompanied by influx of a different depositional system from the southwest, but paleocurrent data to support this is lacking. The massive green-weathering schists of the "turbidite" unit probably are of volcaniclastic origin, and may be related to the subaqueous mafic volcanism indicated by pillowed basalts in Basile Bay.

5. DEFORMATION

The main tectonic foliation, the extension lineation, and associated folds that deform the Wilson Island Group collectively are manifestations of what is here referred to as the "main phase" of deformation. Both the foliation and the lineation are defined by preferentially oriented metamorphic minerals, and therefore the main phase of deformation was synchronous with metamorphism.

In the Wilson Island and Basile Bay areas, the rocks mainly were metamorphosed to the biotite zone of greenschist facies. The higher metamorphic grade of rocks at the Outpost Islands suggests that they belong to a deeper structural level. This is consistent with the predominant easterly plunge of bedding-cleavage intersections. Porphyroblasts of staurolite and andalusite at the Outpost Islands have in part overgrown the foliation, and in turn are enveloped by the foliation, indicating that they grew during deformation.

Structural aspects of the main phase of deformation are most consistent where the rocks form steeply dipping homoclinal panels, such as on the main part of Wilson Island and between Basile Bay and Basile Lake. In these areas the cleavage consistently dips steeply to the southeast, and bedding-cleavage intersections plunge eastward at moderate angles.

Where folds are prominent, the stretching lineation is well developed and shows very little deviation from its steep southerly plunge. The other structural elements, however, have more complex relationships to the folds. For example, the main foliation in pelitic rocks "wraps around" the hinges of: (1) the eastward-plunging folds at the Outpost

Islands, (2) the outer arc of the Blind Bay anticline (i.e., northeast of the Scimitar fault), (3) asymmetric folds on the north limb of, and probably parasitic to, the Blind Bay anticline, and (4) rocks of the Five Snares assemblage. At the Outpost Islands, the foliation is cut by a crenulation cleavage. In the core of a small anticline on Wilson Island, the main cleavage is deformed by kinks and crenulations that are coaxial with northeast-plunging minor folds. Except at the Outpost Islands, bedding-cleavage intersections deviate considerably from the regional eastward trend in each of these examples. These relationships clearly indicate that the rocks were folded after the formation of cleavage in the pelites.

In contrast, the cleavage in quartzitic and conglomeratic units fans about the hinges of folds in such a way that the angle it makes with bedding displays opposite vergence on opposite limbs of the folds. Development of this cleavage was therefore clearly synchronous with folding. The diachrous cleavage development and the variation in fold hinge orientations illustrates that the main phase of deformation may have been two separate events, rather than an episode of simple, cylindrical folding.

One possibility is that the main foliation in the pelitic rocks (slaty cleavage and schistosity) was created during a separate, earlier period of deformation, which perhaps also caused folding or tilting of the strata. A pre-existing tilt would have controlled the orientation of later folds. This could explain why the Basile Bay syncline plunges steeply and why some minor folds actually face downward, whereas most folds have gentle or moderate plunges. The possibility of two unrelated phases of deformation is difficult to evaluate, because cross-cutting structures only are locally developed.

A second possibility is that the main phase of deformation was a single progressive orogenic event. In this case, foliation in the pelitic rocks must have been created early in the deformational history, and then folded during later stages, perhaps in response to migration of the hinges of folds perpendicular to their axes. This seems reasonable, considering that some folds which deform the main cleavage are nearly coaxial with the regional easterly trend. The asymmetric folds in the Outpost Islands and the small anticline on Wilson Island are in this category.

Some property of the quartzite and conglomerate units prevented them from developing a cleavage until late in the folding history. In environments of low-grade metamorphism, deformation of porous rocks can be accomplished by grain-boundary sliding (particulate flow) without significant deformation of the grains. High pore fluid pressures can enhance the degree to which particulate flow occurs (Borradaile, 1981). Because the sandstone and conglomerate protoliths of these rocks probably were relatively porous, particulate flow could have played a major role in their deformation during early stages. Late-stage dewatering would have made conditions favourable for cleavage formation. The prevalence of sutured contacts between adjacent quartz grains, and of pressure shadows containing sericite and quartz that are fused to the recrystallized margins of quartz grains, indicates that diffusional mass transfer (pressure solution) was important in the development of the cleavage.

Although the main-phase folds have apparently deformed the slaty cleavage, formation of layer-parallel cleavage is possible, even in the hinges of folds. For example, the bedding-parallel cleavage in thin

pelite layers of the Lower Quartzite member could be analogous to the arcuate hinge-cleavage of Roberts (1971). Model experiments by Roberts and Strömgård (1972) have shown that such cleavage tends to form only on the inner arcs of incompetent layers in folded multilayered sequences. They speculate, however, that the effects of layer-parallel compression on relatively thin incompetent units may be minimized in successions which are dominated by thick competent layers, hence favouring the development of arcuate hinge-cleavage.

Alternatively, if during folding the quartzite members (Wblq and Wbuq) behaved as stiff bars that maintained an essentially constant thickness, then these members might have deformed by tangential longitudinal strain (cf. Ramsay, 1967, p. 397). The distribution of strain in folds of this type (Ramsay, 1967; Hobbs, 1971) is such that a neutral surface of no finite longitudinal strain separates domains of layer-parallel shortening (expressed by convergent cleavage fanning) in the inner arc from domains of layer-parallel extension (hence bedding-parallel cleavage) in the outer arc.

Slaty cleavage in the Five Snares assemblage clearly has been folded around the hinges of anticlines beneath the Basile Bay syncline, because the bedding-cleavage vergence is constant. There is no obvious reason to believe that these anticlines are not genetically related to the syncline. Therefore, since only one cleavage is conspicuous in this area, it is doubtful that its development was concomitant with folding. This does not preclude the possibility that the mechanism of tangential longitudinal strain was important in the folding history; any pre-existing cleavage in the outer arc that was inclined at a small angle to bedding probably would rotate toward the bedding in response to tangential stresses.

Tangential longitudinal strain can operate together with flexural mechanisms (Ramsay, 1967). Donath and Parker (1964) have distinguished between flexural folds that form by movement of material along discrete surfaces (flexural slip) and those which deform on a granular scale (flexural flow). In both cases, movement occurs parallel to layering. In contrast, they define passive folds as those which form by slip or flow of material across layering. Passive folds form in rock types that are characterized by high mean ductility and low ductility contrast (such as slate, marble, or gneiss) and typically have similar fold geometries, whereas flexural slip and flexural flow mechanisms occur in less ductile rocks and give rise to parallel, or nearly parallel, folds (Donath and Parker, 1964).

In the Basile Bay syncline, the Lower Pelitic member approximates a similar fold profile on the scale of the member itself. Although sandstone layers within the member show no separation across layering, flow or slip across layering (and hence passive folding) possibly has occurred in the more ductile pelites. In contrast, the more concentric fold profiles displayed by the quartzite members are characteristic of flexural folds, and the well-bedded nature of these rocks seems aptly suited to flexural mechanisms. During the course of folding, some combination of flexural slip, flexural flow, and tangential longitudinal strain may have operated in the quartzites, while the thick pelites were deformed by flexural flow or by passive mechanisms. Each of these processes may have operated in different parts of the folding pile at different times.

In ideal situations, layers folded by tangential longitudinal strain exhibit no dimensional changes parallel to the fold axis, so

that the surface areas of layers in the inner arc are decreased, whereas the surface areas of the outer layers are increased (Hobbs, et al., 1976, p. 187). In natural situations, shortening perpendicular to the fold axis in the inner arc conceivably could be compensated by extension parallel to the fold axis. This could be considered a possible explanation for the extension lineation in the inner arc of the Basile Bay syncline (i.e., strained quartz granules in the Upper Quartzite member), which is collinear with the fold axis as defined by bedding-cleavage intersections. However, the extension lineation in the Wilson Island Group plunges steeply even where associated with folds of gentle plunge.

Within the context of progressive deformation, the range in orientations of fold hinges suggests that they were rotated during deformation. This could have happened on a regional scale (i.e., the hinges are curved), or it may be that the hinges of map-scale folds have rotated independently of a larger structure (i.e., the folds are incongruous; cf. Ramsay and Sturt, 1973).

Ramsay (1979) has shown that the hinges of folds which initiate even slightly oblique to the Y axis of the regional strain ellipsoid (i.e., regional fold axis) tend to rotate within the XY plane of strain during orogenic deformation. The largest angular deviations from the regional trend that result from such rotation are associated with the tightest folds. A comparable relationship could exist in the Wilson Island Group. The hinge of the near-isoclinal Basile Bay syncline deviates markedly from the regional moderate plunge. In contrast, folds that plunge parallel to the regional norm are more open, as exemplified by typical folds on the main part of Wilson Island, in the

inner arc of the Blind Bay anticline, and at the Outpost Islands. Hinges of folds like the Basile Bay syncline could have rotated into their present orientation if they initially formed somewhat oblique to the regional trend. Why the hinge would initiate at an angle to the regional trend is not clear, but a pre-existing tilt imposed by earlier deformation could have been a controlling factor.

The most striking case for curved fold hinges is near the southern tip of Wilson Island, where a structural culmination in the Reinhardt formation separates a domain of southwest-plunging folds in the Reinhardt islands from a domain of northeast-plunging folds in the core of the Blind Bay anticline. Although such a culmination could have been produced by superposed upright folding about north- or northwest-trending axes, no mesoscopic structures indicative of such folds were observed; the noncylindricality of main phase-folds most likely is a function of main-phase deformation. Parasitic main-phase folds in the inner arc of the Blind Bay anticline (at about the level of the Pebbly Sandstone member) are disharmonic relative to the outer arc of the fold. Ramsay and Sturt (1973) have pointed out that noncylindrical fold styles can evolve as a result of non-uniform extension within the folding layers, and that disharmonic patterns in noncylindrical folds may reflect an added flattening component of strain.

Flattening with a subvertical component of extension, imposed at a late stage in the main phase of deformation, could also have been responsible for the development of cleavage in the quartzites, of crenulation cleavage in pelites, and of the stretching lineation. Such flattening may have been non-uniform or inhomogeneous on a regional scale, such that its overprint was strongest in areas where the section

was thickened by folds. This could explain why the stretching lineation and late cleavages are strongly expressed in the folds, and less so in the homoclinal panels.

Attempts to document the kinematics associated with development of the stretching lineation revealed that symmetrical pressure fringes are common in rocks of the Five Snares assemblage. Symmetrical pressure fringes imply irrotational strain (Ramsay and Huber, 1983) in the XZ plane. This is consistent with the flattening hypothesis, and does not preclude the possibility that noncylindrical folds developed by rotation of hinges in the XY plane of strain. The two samples that display asymmetrical pressure fringes are considered to reflect local rotations in XZ that have little significance for the regional kinematic history, especially since they have opposing senses of asymmetry.

Mylonitic quartzites at the Petitot Islands were dextrally sheared in the biotite zone of greenschist facies. This style of deformation is characteristic of the youngest portions of the Great Slave Lake Shear Zone (Hanmer and Lucas, 1985), east of the Petitot Islands. If these quartzites are part of the Wilson Island Group (Reinhardt, 1969), then a component of transcurrent shearing should also have affected other rocks of the Group during the main phase of deformation. For example, transcurrent shearing (transpression?) could have induced the migration of fold hinges and rotation of early formed cleavage.

It can be said summarily that the main phase of deformation was characterized by north-south compression with a subvertical component of extension. Deformation may have occurred in two (or more) separate episodes, but all structures were formed under metamorphic conditions. The total strain is suggestive of buckling with superposed flattening.

Some of the kink folds and crenulations that deform the main

foliation may have formed during late stages of the main phase of deformation, as previously discussed (i.e., crenulation cleavage in Outpost Islands; crenulations on Wilson Island). Others have geometries that are incompatible with the main phase structures, and thus were most likely related to a separate tectonic event. The best example of the latter is the Z-shaped kink that deforms S-shaped folds in the Outpost Islands. This structure post-dates even the crenulation cleavage, which is axial planar to the S-shaped folds. Gently dipping crenulations like the ones in the hinge of the Basile Bay syncline indicate vertical compression. These also must be unrelated to the main phase, which was characterized instead by subvertical extension. It is speculated that much of the kink folding was related to thrusting during emplacement of the East Arm nappes. Some of the steeply plunging kinks could have formed in response to later transcurrent faulting.

Based on their orientation and sense of stratigraphic separation, the northeast-trending faults are inferred to be dextral transcurrent faults of the McDonald fault system. They are thus related to an extensive system of transcurrent faults that formed in response to a probable terminal collision outboard of Wopmay Orogen (Hoffman, 1980b), between 1.84 and 1.81 Ga (Hoffman and Bowring, 1984). Later dip-slip reactivation of these faults (Hoffman, 1980a) could be responsible for the apparent sinistral separation exhibited across some of them.

The Blind Bay fault could be a thrust fault akin to the Inconnu Thrust, which parallels the Blind Bay fault to the south. The southern "lower jaw" of Wilson Island may therefore be a tectonic horse that was torn from the back of the Wilson Island Terrane by the Simpson Islands

Terrane during overthrust deformation of Athapuscow Aulacogen. The marked difference between the mean northeast strike of the Reinhardt formation and the east-northeast strike of the Safety Cove formation is better explained by a thrust fault than by rotation due to transcurrent faulting. The southeast dip of the fault that straddles longitude 113°W on Wilson Island suggests that it, too, may have initially formed as a thrust fault.

The northwest-trending faults are interpreted as normal faults that formed concurrently with the McDonald fault system. This is consistent with their orientation, angle of dip, and sense of stratigraphic separation. Near the west end of Wilson Island, faults of this set have displaced blocks of Safety Cove formation in a west-side-down, stepped fashion. Both west-side-down and east-side-down faults cut the Reinhardt formation. The foot-shaped bay at the east end of Basile Lake occupies a triangular graben.

Because bodies of rock bounded by dextral transcurrent faults tend to rotate counterclockwise during deformation (Freund, 1970), the faults with more westerly orientations may be somewhat older than those which strike northwest. The separations displayed by many of the faults near the south end of Wilson Island are probably best explained by oblique slip. Such faults probably are normal faults that rotated into orientations favouring strike-slip movement.

6. SYNTHESIS - TECTONICS OF THE WILSON ISLAND GROUP

The Wilson Island Group is inferred to have been deposited mainly in shallow water. An early stage of volcanism and alluvial sedimentation (Reinhardt formation) was succeeded by a period of mixed carbonate and siliciclastic sedimentation (Safety Cove formation and Five Snares assemblage) that appears to have involved interaction between an alluvial system and a shallow sea. Soft-sediment deformation structures, intrabasinal coarse detritus, and vertical facies oscillations collectively provide evidence that deposition in these early stages was tectonically controlled. If the interpretation presented herein is correct, the upper half of the succession (Basile formation) records the progradation of an alluvial system, followed by a major transgressional stage. Based on the aggregate thickness of the Reinhardt, Safety Cove, and Basile formations, the Wilson Island Group is at least 8 kilometres thick.

Folds and penetrative fabrics formed in response to north-south compression with a subvertical component of extension, at metamorphic grades ranging from greenschist to amphibolite facies. Deformed clasts of Wilson Island Group in the lowermost Great Slave Supergroup indicate that this tectonic event happened before the Great Slave Supergroup was deposited. The Wilson Island Group probably was involved in later thrusting that carried nappes of basinal facies of the Great Slave Supergroup northward over their autochthonous platformal equivalents (Hoffman et al., 1977; Hoffman, 1981, 1985), between about 1860 and 1885 Ma (Hoffman and Bowring, 1984).

Interpretation of the tectonic setting of the Wilson Island Group

is complicated by its structural isolation relative to other rocks, but an appropriate tectonic model for these rocks must satisfy certain constraints. From a regional standpoint, one of the most important constraints is the age. Bowring et al. (1984) interpreted the U-Pb zircon date of 1928 ± 11 Ma as a minimum age of the Wilson Island Group, because they considered the felsic porphyry from which the date was obtained to be intrusive into the clastic and volcanic strata of the Reinhardt formation. Based on the present investigation, the northern margin of this felsic unit is reinterpreted as a fault (the Scimitar fault, Maps 1A, 1B); the southern margin appears to be concordant with the underlying Unnamed Sandstone unit, which pinches out to the west. It is therefore likely that this felsic unit is an extrusive dome or flow, and that the zircon date thus represents the age of Reinhardt strata.

Secondly, the depositional basin must have been either on, or adjacent to, continental crust, because detrital compositions clearly indicate a continental provenance. The bimodal volcanic suite of the Reinhardt formation suggests that the crust was in a tensional stress regime during initial deposition.

Previous workers have suggested that the Wilson Island Group was deposited in a fault-bounded trough (Yeo, 1976a, b) or rift (Hoffman et al., 1977). These hypotheses are consistent with extensional tectonics, bimodal volcanism, the composition of the sandstones, and the overall upward-fining sedimentary stratigraphy. The Wilson Island Group could have formed in a continental rift, in either the Slave craton or in a rifted fragment thereof.

More recently, Hoffman et al. (1986) have outlined evidence for the collision of two cratonic terranes, now represented by the Slave

and Churchill Provinces, around 1.89 Ga. The effects of oblique east-dipping subduction prior to collision are manifested along the northwest margin of the Churchill Province, where magmatism and concomitant dextral shearing occurred between about 1.98 and 1.90 Ga. A batholithic belt that delineates the Churchill margin, and of which the Great Slave Lake Shear Zone (Hanmer and Lucas, 1985; Fig. 2A) is a part, is thus inferred to be the deeply eroded remnant of a magmatic arc. Based on its age and position relative to this arc, the Wilson Island Group is a possible candidate for a forearc basin (Hoffman et al., 1986).

Although quartzo-feldspathic sandstones typically reflect an uplifted continental basement provenance, dissected magmatic arcs can supply quartzo-feldspathic sediment of mixed volcanic and plutonic origin to forearc basins (Dickinson and Suczek, 1979). An example of the latter is the late Mesozoic Great Valley Group of California, which records a progressive change in the dominant souce terranes from volcanic and sedimentary to plutonic, as a function of progressive unroofing of the Sierra Nevada Batholith (Ingersoll, 1983). Similarly, volcanic and sedimentary sources contributed detritus to the Reinhardt formation and later became less important. Quartzo-feldspathic material of probable plutonic origin is abundant throughout the Wilson Island Group, which means that if this sediment were derived from a magmatic arc, the plutonic "roots" of the arc had already been unroofed by about 1930 Ma.

Forearc slivers bounded on one side by a trench and on the other by arc-parallel strike-slip faults are characteristic of two-thirds of the modern subduction systems that have upper plates of continental crust (Jarrard, 1986). The Wilson Island Terrane (Hoffman, 1985) could have been such a forearc sliver, separated from the Churchill plate by the dextral Great Slave Lake Shear Zone and upper-crustal brittle

equivalents that have since been eroded. Reinhardt (1969) has suggested that the earliest strike-slip activity of the dextral McDonald fault system dates back to pre-Wilson Island Group times. Yeo (1976b) and Badham (1978) have contended that strike-slip faulting controlled the deposition and the deformation of the Wilson Island Group. If so, then the basin of deposition could be tectonically analogous to the Progreso Basin of Ecuador, which probably formed in response to strike-slip movements on the Dolores-Guayaquil megashear (Moberly et al., 1982). However, basin evolution of the Wilson Island Group may also be explained by extension unrelated to strike-slip faulting, even in terms of a forearc setting.

Dewey (1980) has shown that strain within arcs can be compressional, extensional, or neutral. The determining mechanical parameters are the component of motion of the overriding plate toward the trench and the rate at which the hinge of the subducting plate rolls back from the trench as the plate sinks. Compressional, extensional, and neutral arc tectonics are predicted, respectively, when the rate of upper-plate convergence is greater than, less than, or equal to the roll-back rate. The values of these parameters may change during the complex evolution of an arc system (Dewey, 1980).

Shelved forearc basins (Dickinson and Seely, 1979) can localize thick accumulations of shallow-water sediments, and such basins can be characterized by extensional tectonics even where the overall strain regime of the arc is compressional. For example, the Manabi basin of Ecuador contains up to 10 kilometres of shallow marine and terrigenous sediment, and is underlain by block-faulted oceanic crust (Lonsdale, 1978), even though the central Andean subduction system, on a regional scale, is a classic example of a compressional arc (Dewey, 1980).

Shelved forearc basins that rest on the cratonic arc massif off the coast of central Peru (the Sechura and Salaverry Basins) also are characterized by extension, while adjacent slope basins are undergoing compression (Thornburg and Kulm, 1981; Kulm et al., 1982).

The Sechura Basin rests in part on coastal Peru, and extends offshore toward the Salaverry Basin, which is entirely submarine. Both basins are separated from the slope by a topographically positive, linear, outer shelf high that parallels the coast (Thornburg and Kulm, 1981). The Cenozoic history of these basins has included periods of marine sedimentation and of subaerial exposure (Kulm et al., 1982). The physiography of the Sechura Basin is a suitable forearc model for the Wilson Island Group, as depicted in Figure 15A. An outer shelf high, parallel to the continental margin, could have controlled the prevalent westerly paleocurrents. The paleocurrents are subparallel to the Churchill margin, and, unless they have been modified by deformation, are consistent with the axial paleocurrent patterns that typify forearc basins (see Dickinson and Seely, 1979). Terrigenous clastic sedimentation could have taken place in the subaerial upper end of the basin, while marine sedimentation occurred offshore.

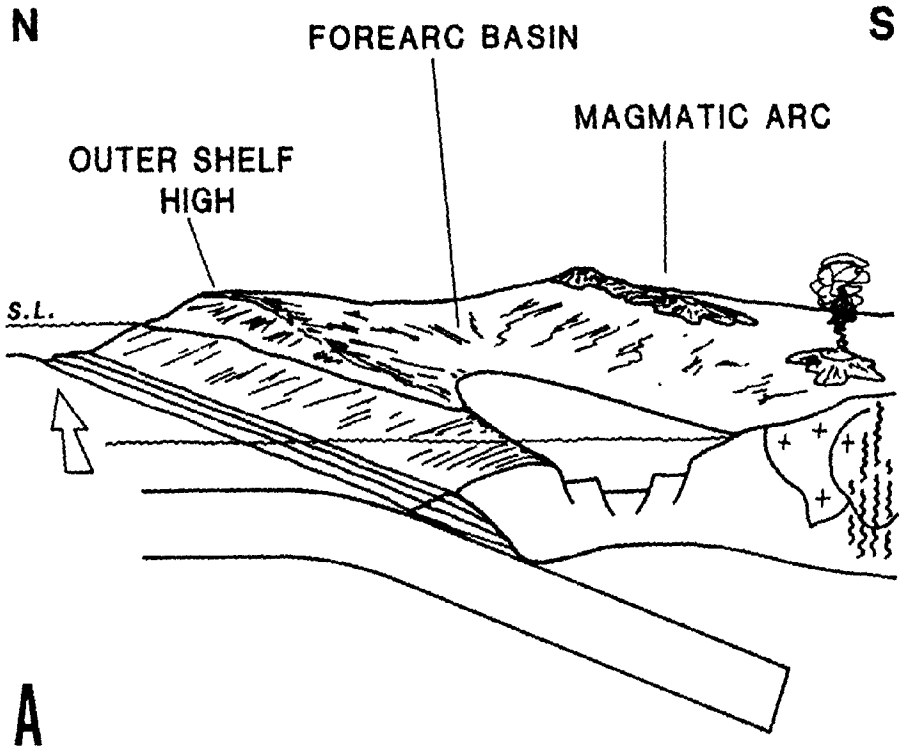
Transgression during late-stage sedimentation of the Basile formation may have been caused by deepening of the basin. This is consistent with foundering of the shelf in response to tectonic erosion of the inner trench wall, which is a process characteristic of compressional arcs (Dewey, 1980). The turbidite and mudstone units may therefore be foredeep deposits.

An arc can become more compressional as progressively younger and more buoyant oceanic crust is subducted. Compressional deformation of

FIGURE 15.

A. Forearc basin model for deposition of the Wilson Island Group. Forearc basin is separated from slope and trench by an outer shelf high, underlain by either continental or transitional crust. Arrow indicates oblique motion of subducting plate.

B. Cartoon illustrating compressional deformation of the Wilson Island Group in hypothesized forearc setting.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FOREARC BASIN
MAGMATIC ARC
OUTER SHELF
HIGH
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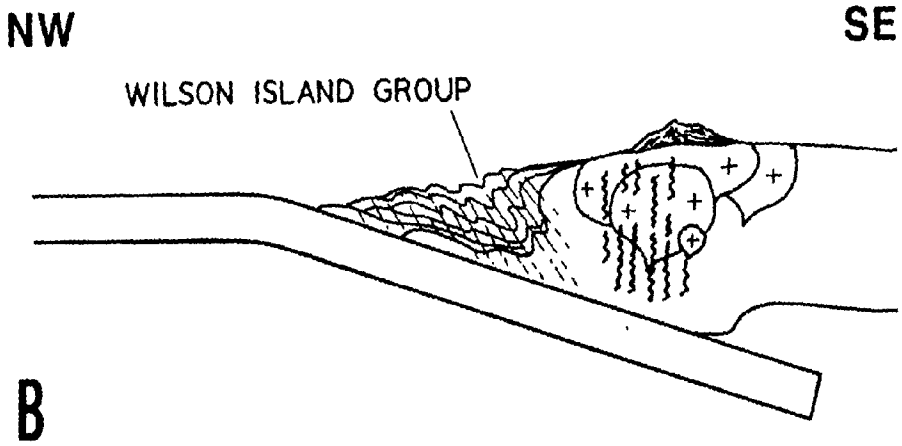
NW
SE
WILSON ISLAND GROUP
B

the Wilson Island Group could have been a consequence of this process (Fig. 15B). The polyphase, or diachronous, history of deformation indicated by the "main phase" structures may have been related to changes in the coupling force between the forearc sliver and the subducting plate. Such coupling along oblique convergent margins leads to partitioning of strain in forearc slivers into a steeply dipping component of underthrusting and a trench-parallel strike-slip component (Jarrard, 1986). Changes in the coupling characteristics could be expected to result in alteration of the relative amounts of these two components. Consequently, structures that were formed by underthrusting could have been later modified by strike slip, and vice versa.

In conclusion, the tectonic history of the Wilson Island Group probably initiated with an early stage of crustal extension, and closed with a late stage of compressional deformation. The tectonic setting is poorly constrained. Although the Wilson Island Group could have been deposited in a continental rift, comparisons of its characteristics with modern and ancient forearc basins have revealed numerous possible analogies.

7. CONCLUDING REMARKS

This study has emphasized refinement of the stratigraphy of the Wilson Island Group, and documentation of both the distribution of, and relationships among, the tectonic structures. These data have served to tighten constraints on the depositional and deformational history, and the age, of these rocks.

Before this project, the U-Pb zircon date of 1928 ± 11 Ma was regarded as the minimum age of the Wilson Island Group (Bowring *et al.*, 1984). The current mapping has confirmed that felsic porphyritic rocks are part of the stratigraphic succession, rather than strictly intrusive, and that the unit from which the date was obtained is concordant with sedimentary strata and therefore probably a stratigraphic unit. This allows for reinterpretation of the date as the age of Wilson Island strata. Evidence has been presented for alluvial and shallow-sea deposition of the sedimentary rocks, and for a tectonic control of sedimentation.

Geometrical aspects of the main-phase folds have been outlined, and it has been shown that the development of cleavage was diachronous. Attention has been drawn to the persistent steep stretching lineation, which could prove to be one of the most significant aspects of the Wilson Island Group. Steep stretching lineations have been observed in rocks 100 kilometres northeast of Basile Bay, and work has been scheduled to delineate what appears to be an extensive narrow belt of steep lineations (S. K. Hanmer, pers. comm., 1986).

The results of this study have spotlighted additional problems for future work. Chemical analyses of the volcanic rocks could reveal

information about the tectonic setting. The significance of the deformation of the Wilson Island Group will be better understood if some outstanding stratigraphic problems can be resolved. For example, were the clasts in the Hornby Channel Formation really derived from the Wilson Island Group? If not, the main phase of deformation of the Wilson Island Group may not pre-date deposition of the Great Slave Supergroup; it could in fact be related to emplacement of the thrust nappes in the East Arm. Do the quartzites of the Petitot Islands belong to the Wilson Island Group? If so, what is the relationship between ductile transcurrent shearing of these rocks to the main phase of deformation? What is the relationship between the Wilson Island and Union Island Groups? Perhaps these problems can be addressed in further investigations by means of comparative studies of heavy mineral suites in the quartzites, and of the geochemistry of volcanic rocks from the Wilson Island and Union Island Groups.

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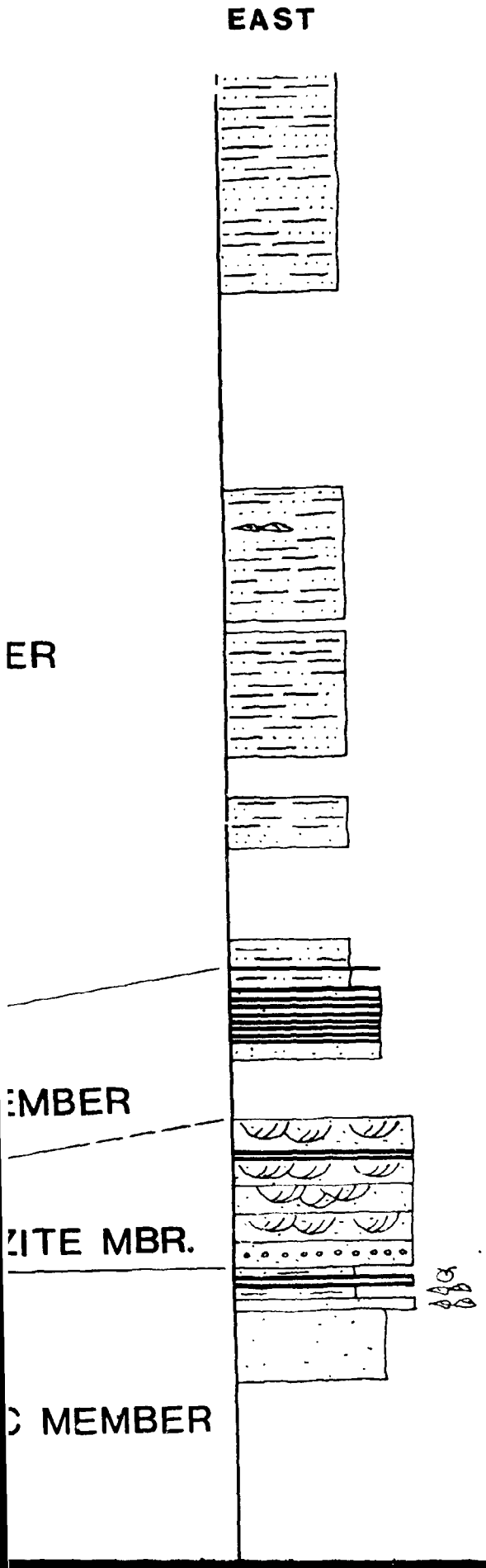
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TABLE 3: STRATIGRAPHIC SUBDIVISIONS OF THE WILSON ISLAND GROUP

BASILE FORMATION

Basalt or Andesite (intercalated with Mudstone unit)

Mudstone unit

Graded member

Ironstone member

Upper Quartzite member

Lower Pelitic member

Lower Quartzite member

FIVE SNARES ASSEMBLAGE

Dolostone

Argillaceous unit

Arkose

---------- McDonald-Wilson Fault ----------

SAFETY COVE FORMATION

Upper Mixed unit

Sandy Dolostone member

Middle Arkose member

Dolomitic Sandstone member

Lower Arkose member

REINHARDT FORMATION

Blind Bay conglomerate

Upper Pebbly unit

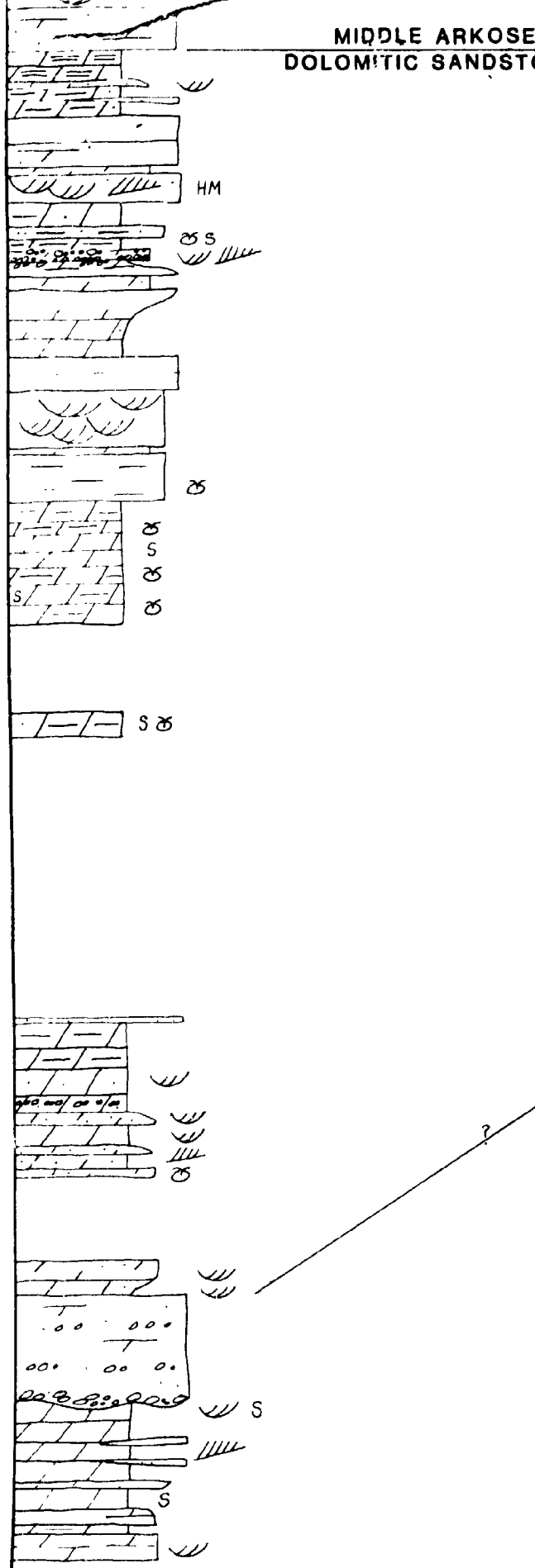
Unnamed Sandstone unit

Felsic porphyritic rocks (occur at numerous stratigraphic levels)

Pebbly Sandstone member

Lower Clastic member

Lower Igneous Complex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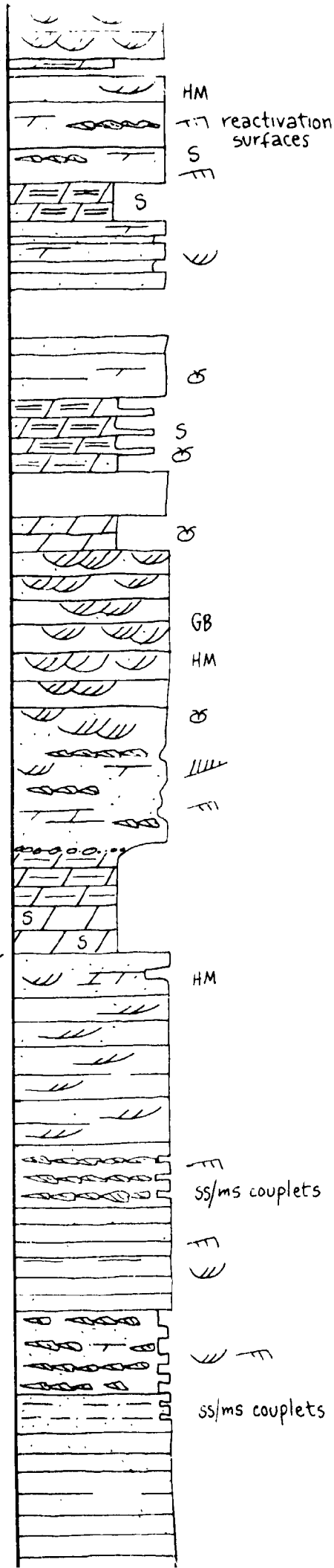
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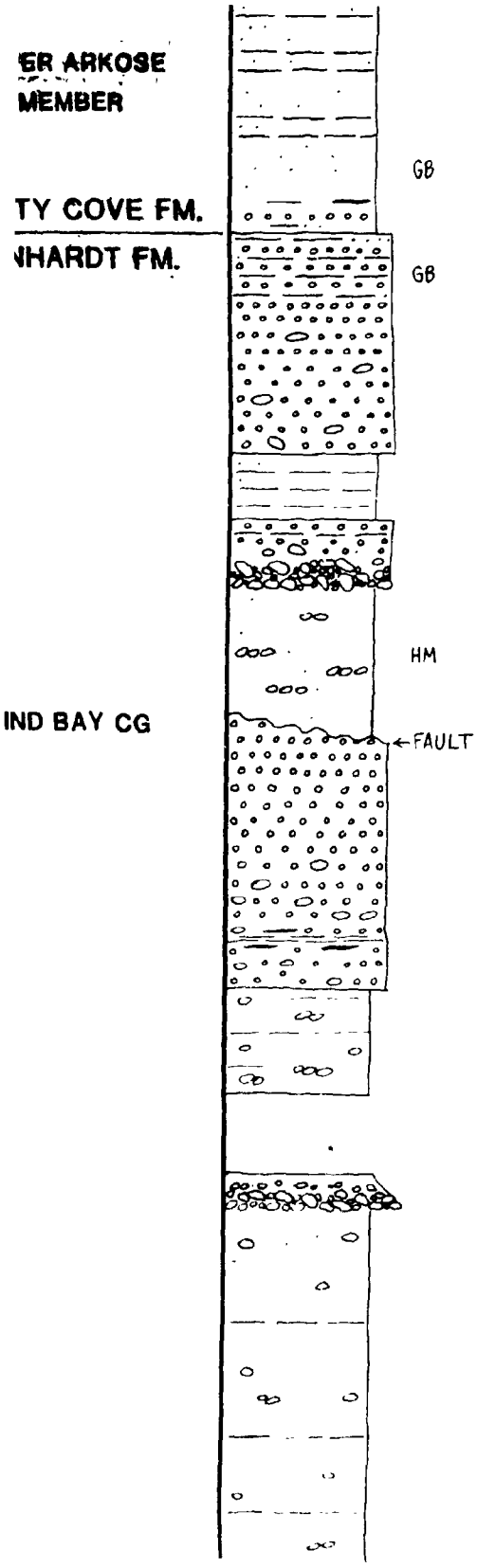

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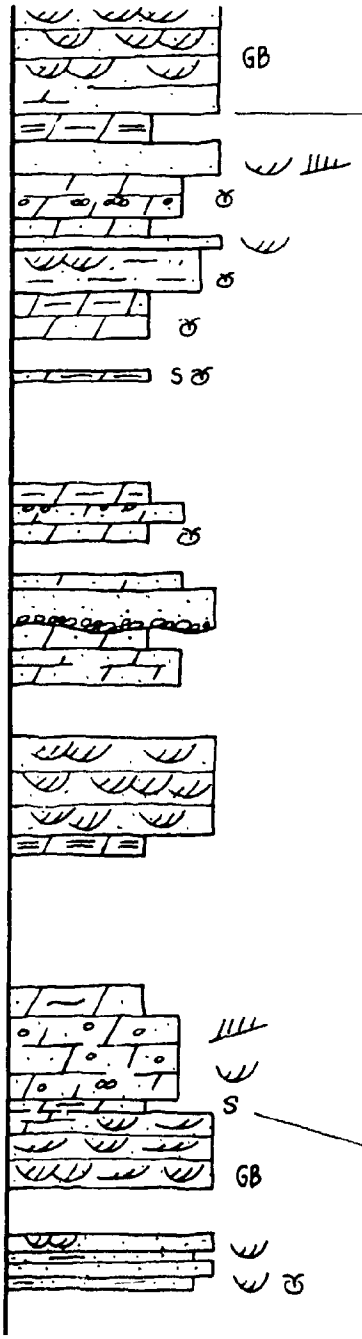

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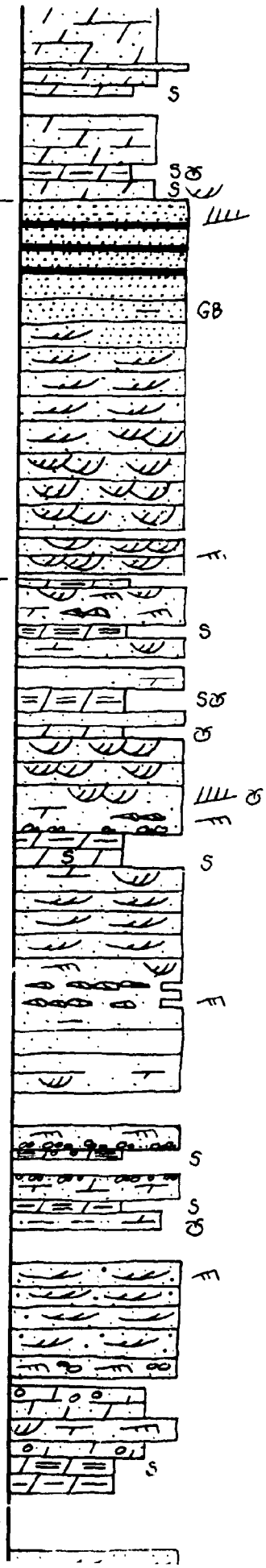

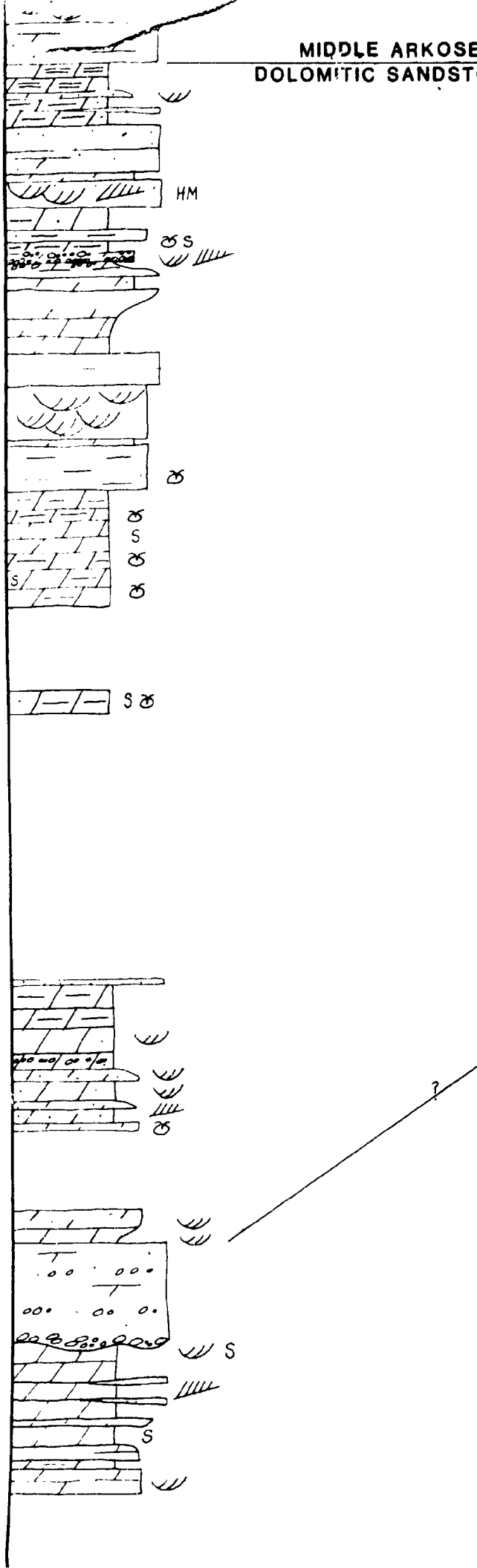
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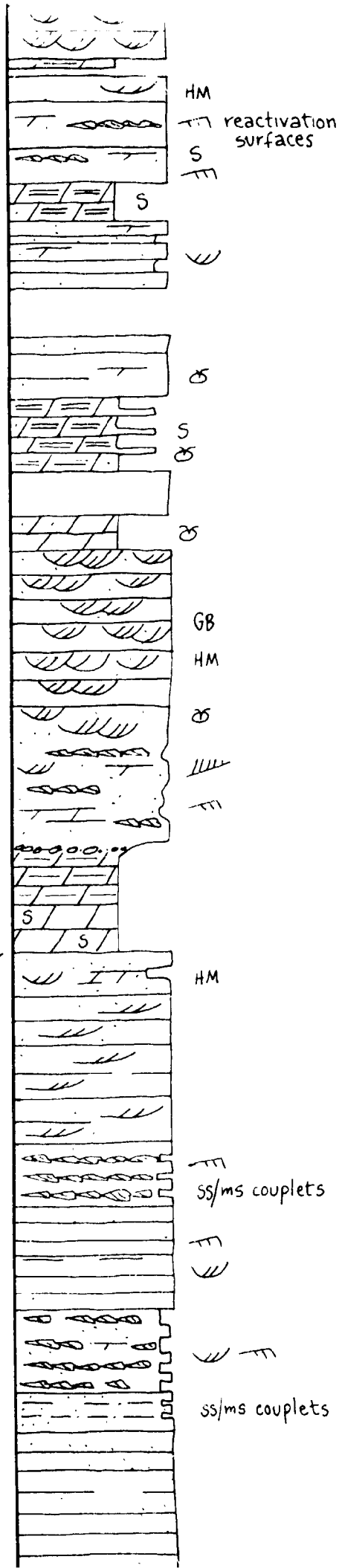

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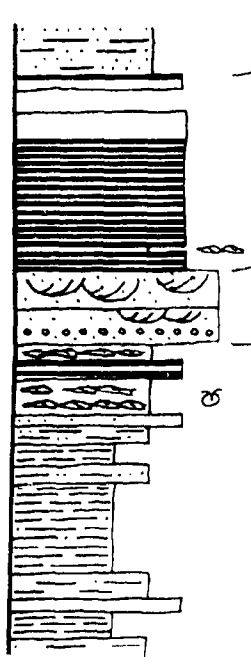

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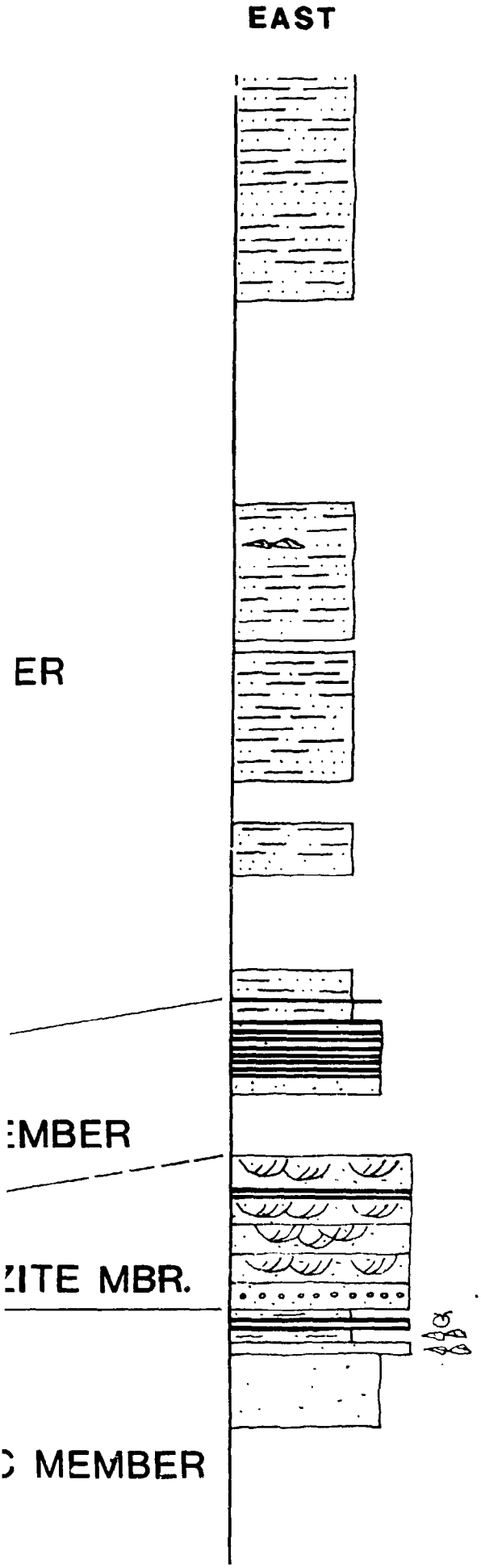
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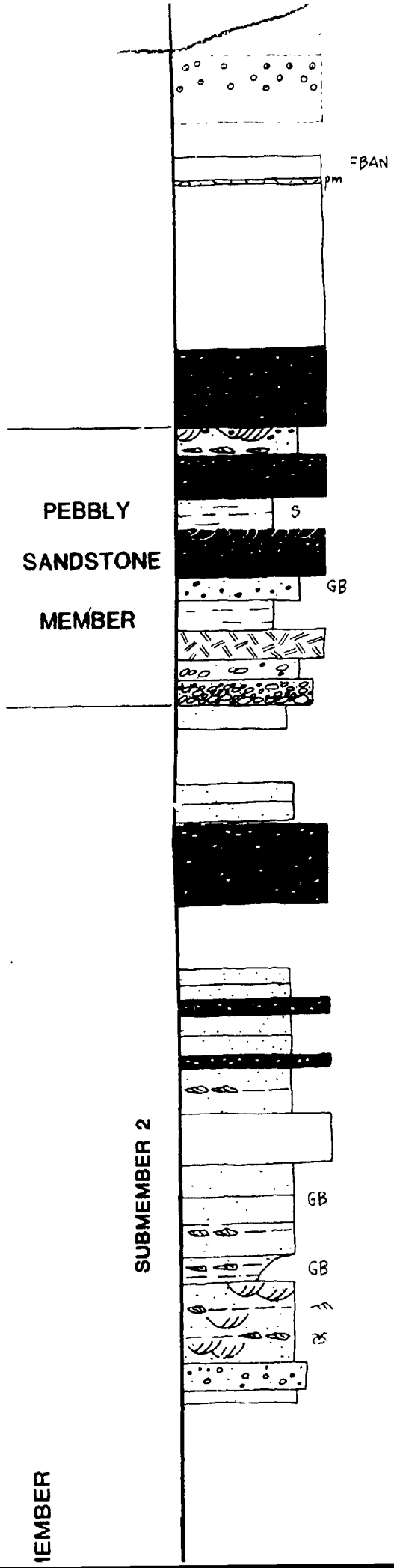
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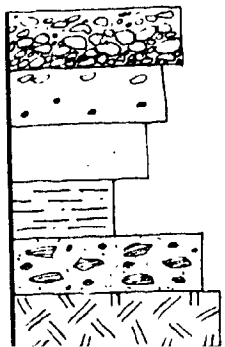
EXPLANATION

S	slump structures
	ball-and-pillow
GB	graded bedding
	lenticular bedding
	ripple lamination
	trough crossbedding



CONGLOMERATE

SANDSTONE { pebbly, granular }

SILTSTONE, MUDSTONE

LAPILLI-ASH TUFF

GABBRO

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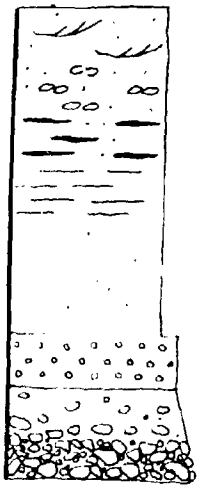
EXPLANATION



HM heavy mineral laminae

GB graded bedding

trough crossbedding

thin pebbly layers

mud rip-ups

argillaceous layers

planar lamination

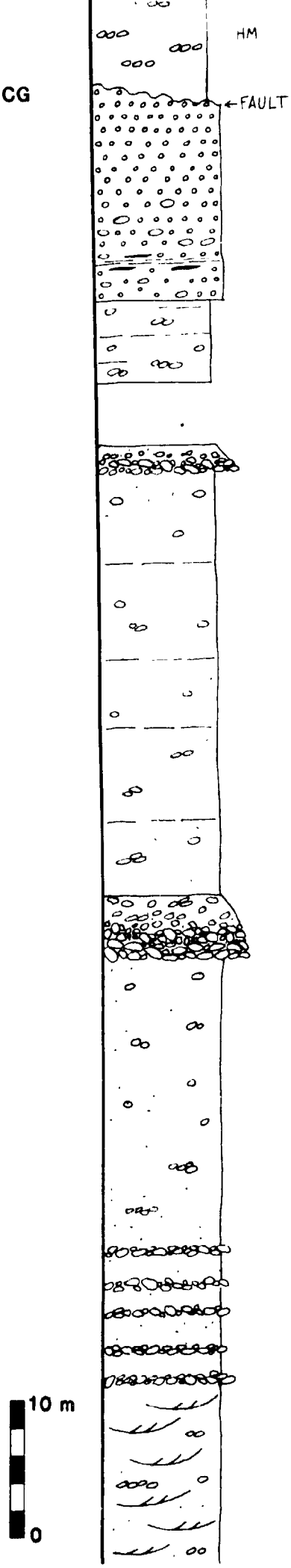
SANDSTONE

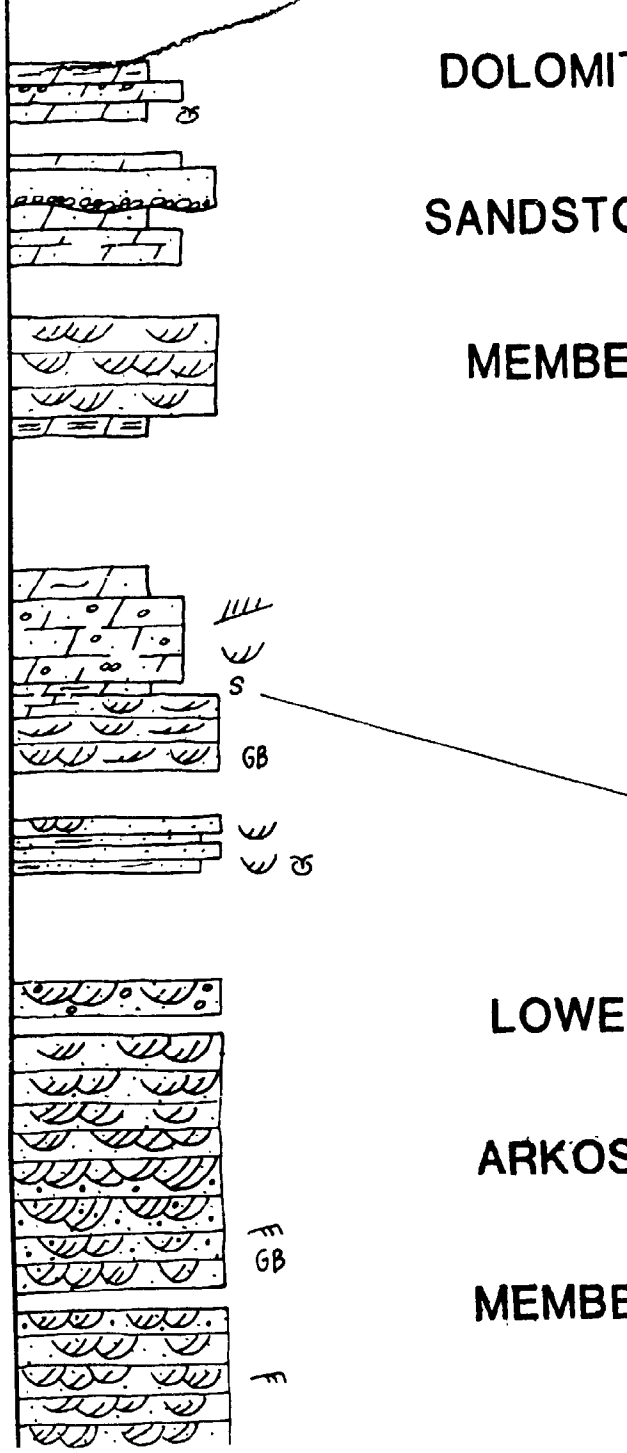
GRANULESTONE

PEBBLY SANDSTONE

CONGLOMERATE

ND BAY CG





EXPLANATION

HM	heavy mineral laminae
GB	graded bedding
	dolomitic lenses/layers
S	slump structures
	ball-and-pillow
	lenticular bedding
	ripple lamination

TIC

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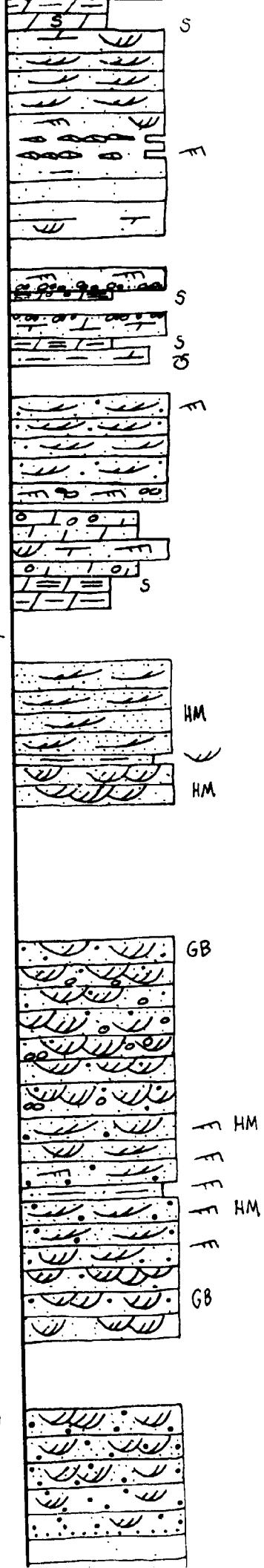

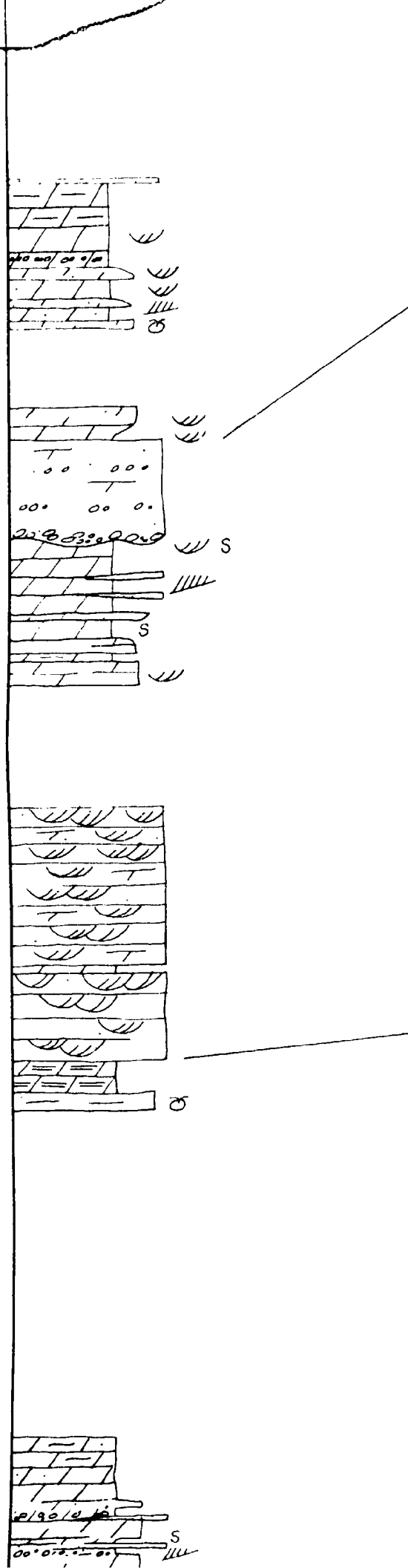
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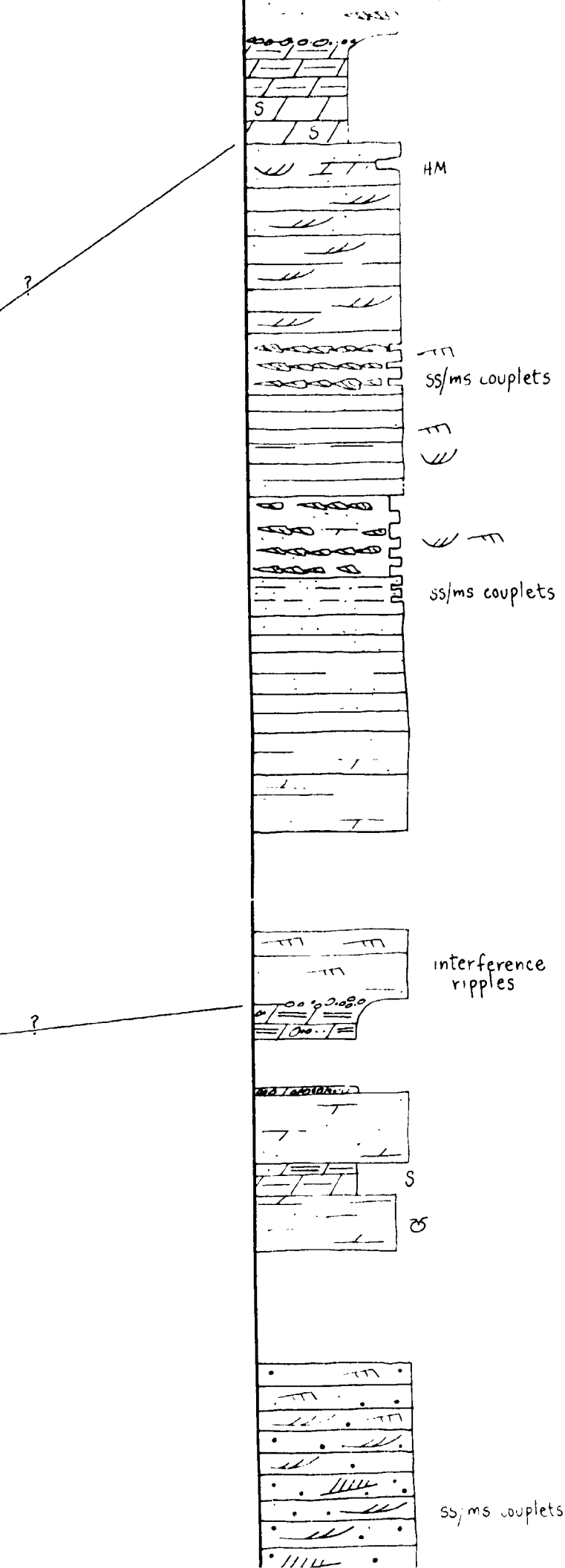
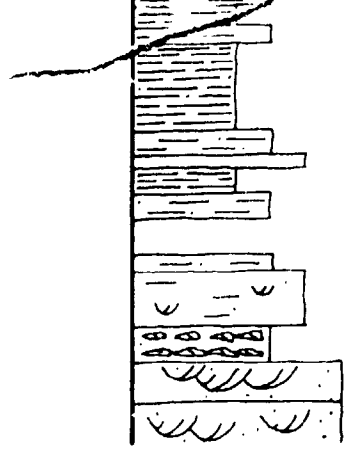

S
S
HM
?
ss/ms couplets
ss/ms couplets
interference ripples
?
S
ss/ms couplets



LOWER

LOWER
MEI

EXPLANATION

granule laminae

HM heavy mineral l

S slump structure

ball-and-pillow

wavy/lenticular

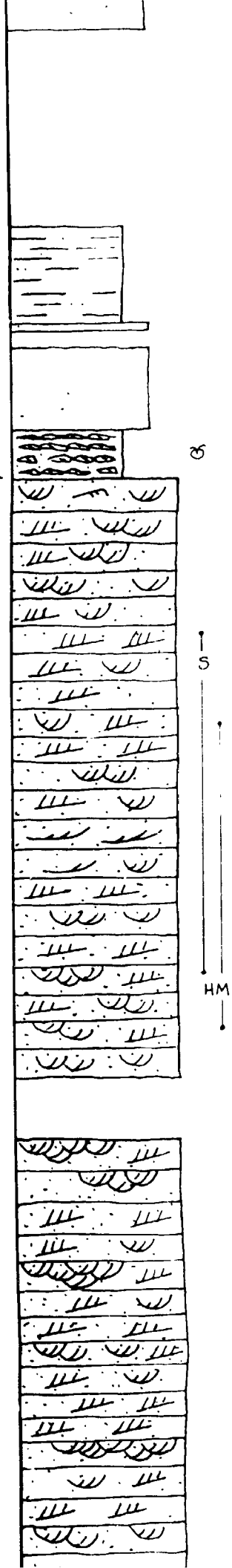
ripple laminatio

planar crossbe

trough crossbe

PELITIC MEMBER

QUARTZITE
MBER



aminae

s

bedding

n

lding

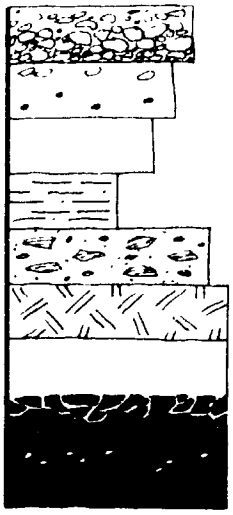
dding

S ∂MEMBE
GB
GB
LOWER CLASTIC MEMBER
GB
SUBMEMBER 1
GB
LOWER
IGNEOUS
COMPLEX

lenticular bedding

ripple lamination

trough crossbedding



CONGLOMERATE

SANDSTONE – pebbly, granular

SILTSTONE, MUDSTONE

LAPILLI-ASH TUFF

GABBRO

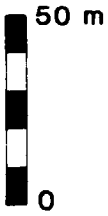
FELSIC PORPHYRY

BASALT – brecciated, vesicular

pm DARK GREY PORPHYRY

FBAN flow banding

spherulitic



EXPLANATION

HM heavy mineral laminae

GB graded bedding

dolomitic lenses/layers

S slump structures

ball-and-pillow

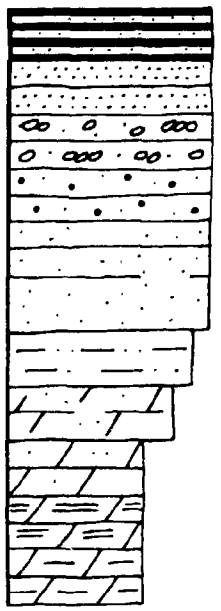
lenticular bedding

ripple lamination

planar crossbedding

low-angle crossbedding

trough crossbedding



iron oxide laminae

planar laminae

pebbly

granular

QUARTZOSE SANDSTONE (SS)

CARBONATE-CEMENTED SS

ARGILLACEOUS SS

DOLOMITIC SS

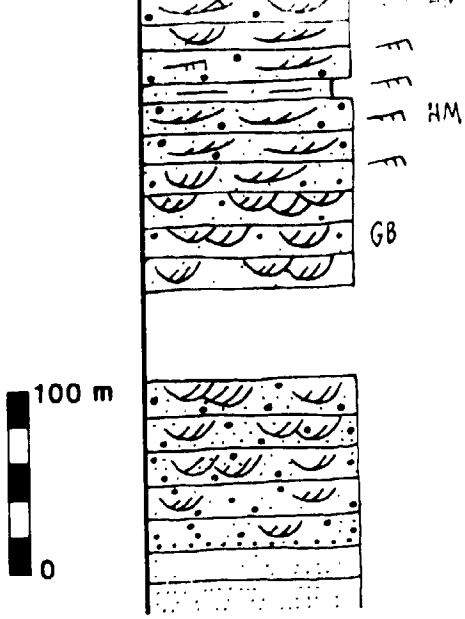
SANDY DOLOSTONE (DOL)

ARGILLACEOUS DOL

LAMINATED DOL

FIGURE C: MEASURED

SAFETY COVE FORMATION,



SECTIONS,

WILSON ISLAND.

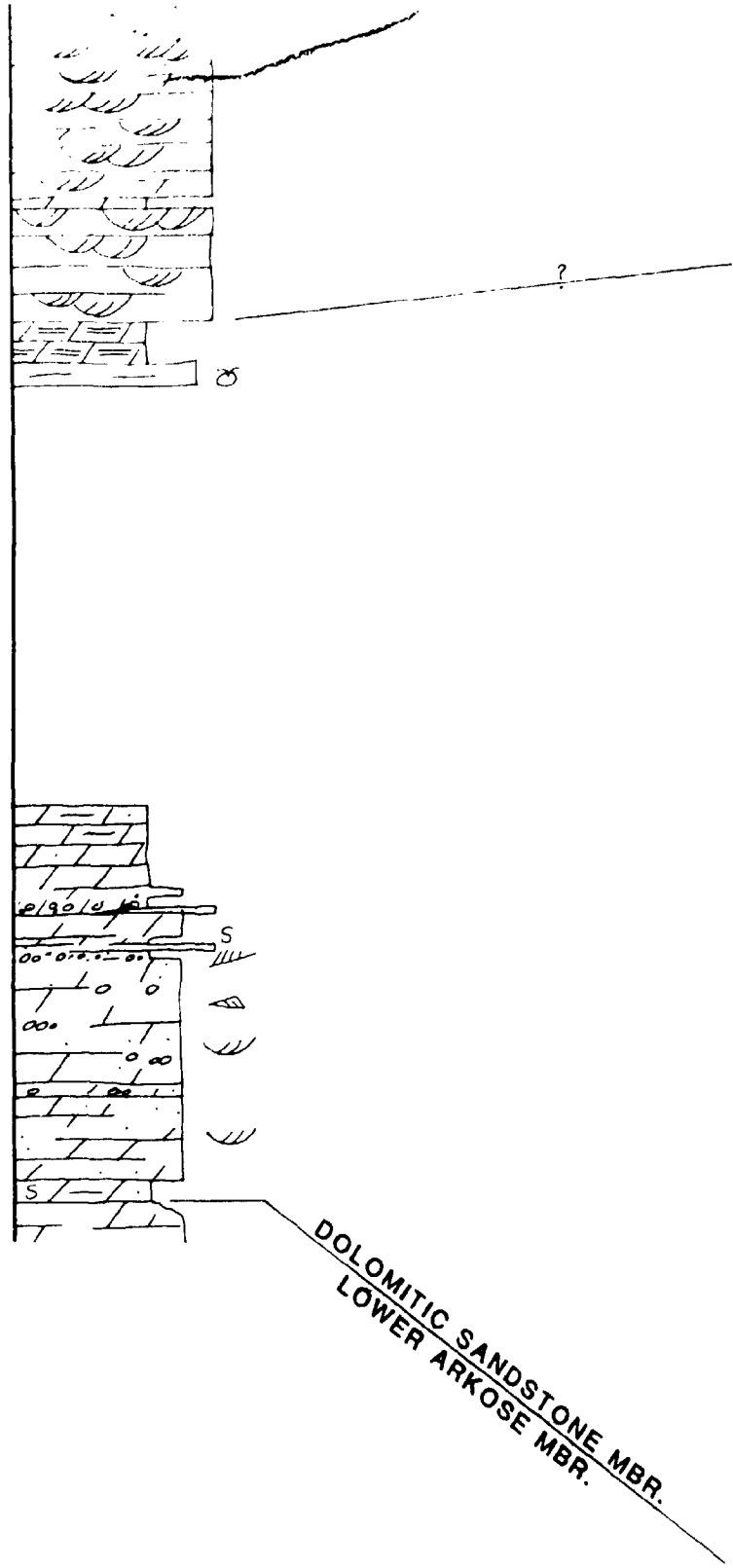
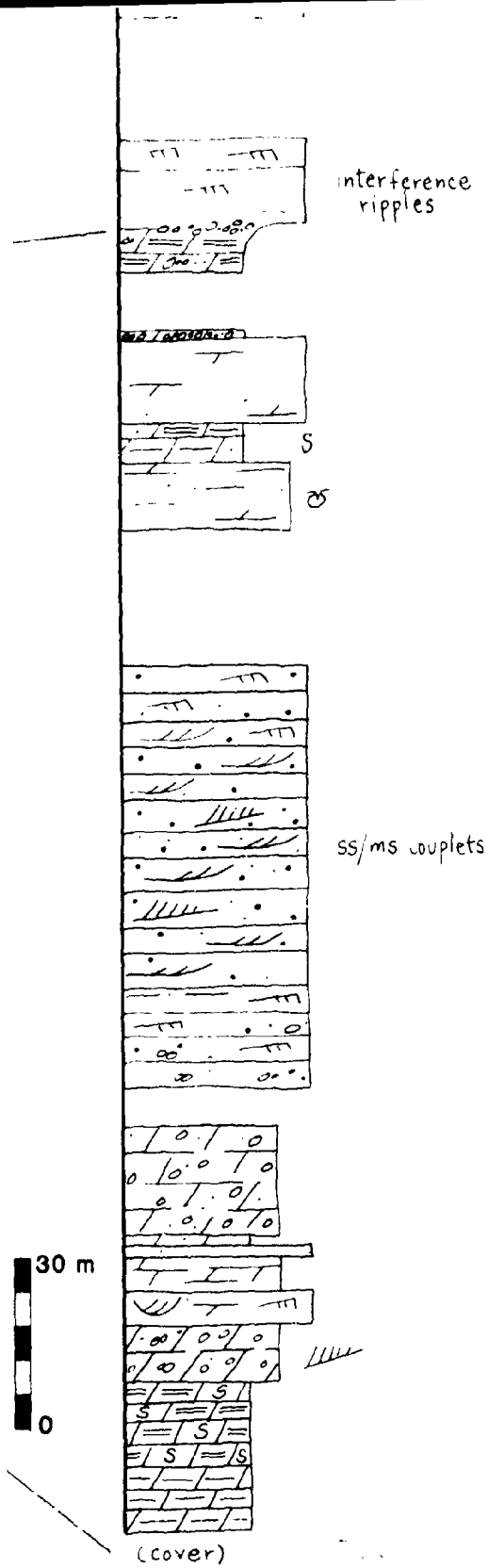


FIGURE D: MEASURED

DOLOMITIC SANDSTONE MEMBER OF

WILSON ISLAN

See Figure C for explanatio

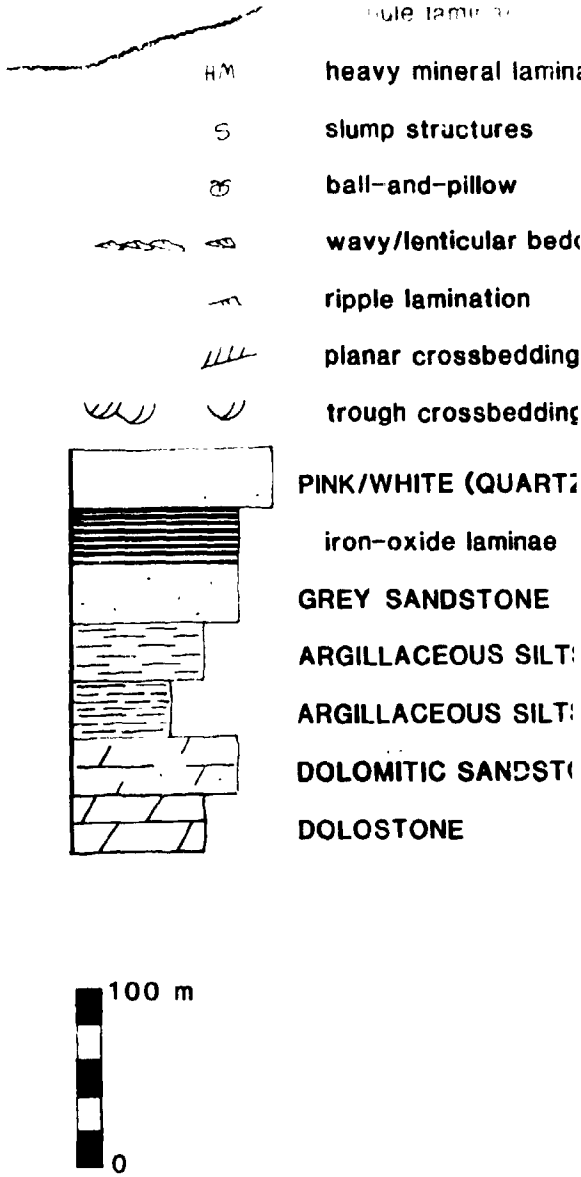


SECTIONS,

SAFETY COVE FORMATION,

ND.

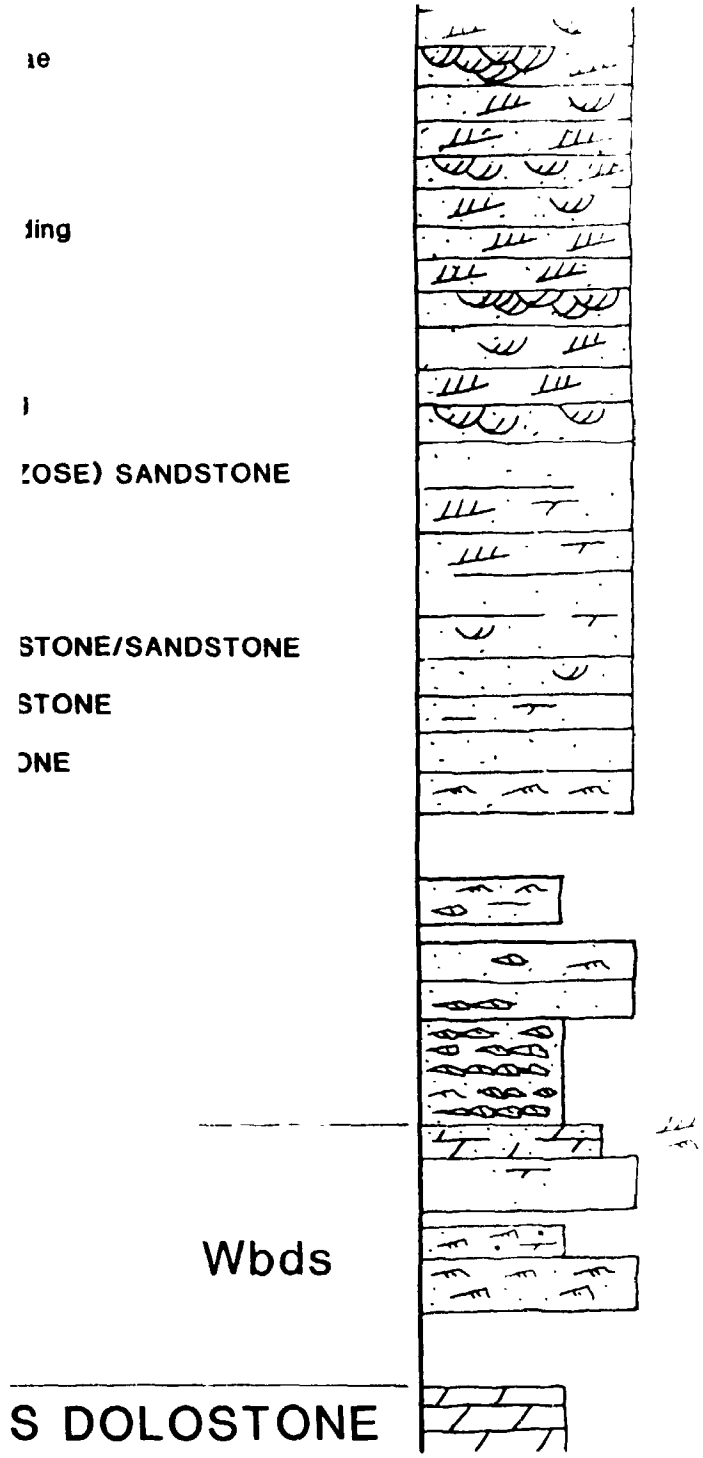
n of symbols.



FIVE SNARE

FIG. E. MEASURED SECT

Basile Bay syncline (L)



IONS: BASILE FORMATION

and east of Basile Bay (R).

		E1-THEN GROUP
	HELIKIAN	Em MURKY FORMATION: c
	--------	----------
		SOSAN GROUP
		S undifferentiated,

		WILSON ISLAND GROUP
		g' diabase dikes
		Wt "TURBIDITE UNIT:"
P		SAFETY COVE FORMATION
R		Wsu Upper Mixed unit:
O		Wssd Sandy Dolostone me
T		Wsm Middle Arkose memt
E		Wsds Dolomitic Sandston
R		Wsa Lower Arkose membe
O	**APHEBIAN**	REINHARDT FORMATION
Z		g gabbro sills and d
O		Wrs sandstone of unkno
I		Wrc Blind Bay conglome
C		Wru Upper Pebbly unit:
		Wro Unnamed Sandstone
		Wrp Pebbly Sandstone m
		Wr_2 Wr_1 Lower Clastic memb
		m mafic or intermedi
		p felsic flows, dome
		k K-feldspar porphyr
		b basalt
		b' basalt, mudstone, a
?	--------	----------
	ARCHEAN?	?

ADDENDUM: suggested colours for map units

Numbers correspond to Prismacolor pencils

MAP 1A

g' maroon (925)

Wt light green (912)

SAFETY COVE FORMATION

Wsu light blue (919)

Wssd dark blue (933)

Wsm yellow (915)

Wsds aquamarine (905)

Wsa sandy yellow (940)

REINHARDT FORMATION

g maroon (925)

Wrs yellow-brown (942)

Wrc orange (918)

Wru light brown (943)

Wro light green (913)

Wrp medium brown (945)

Wr_2 light green (920)

Wr_1 light green (910)

m magenta (930)

p red (924)

k flesh (927)

b purple (932)

b' "

MAP 2A

gr pink (928)

mp light purple (956)

di "

BASILE FORMATION

Wbv violet (931)

Wbm olive green (911)

Wbb maroon (925)

Wbh medium grey (936)

Wbf dark brown (946)

Wbuq cream (914)

Wbp slate grey (936)

Wblq yellow (916)

Wbds aquamarine (905)

FIVE SNARES ASSEMBLAGE

Wfd blue (903)

Wfa dark grey (965)

Wfp red (926)

Wfq golden yellow (917)

	HELIKIAN	**El-THEN GROUP**
		Em MURKY FORMATION: co
	--------	----------
		SOSAN GROUP
		S undifferentiated, m

		WILSON ISLAND GROUP
		g' diabase dikes
		Wt "TURBIDITE UNIT:" f
P		SAFETY COVE FORMATION
R		Wsu Upper Mixed unit: f
O		Wssd Sandy Dolostone mem
T		Wsm Middle Arkose membe
E		Wsds Dolomitic Sandstone
R		Wsa Lower Arkose member
O	APHEBIAN	REINHARDT FORMATION
Z		g gabbro sills and di
O		Wrs sandstone of unknow
I		Wrc Blind Bay conglomer
C		Wru Upper Pebbly unit:
		Wro Unnamed Sandstone u
		Wrp Pebbly Sandstone me
		Wr_2 Wr_1 Lower Clastic membe
		m mafic or intermedia
		p felsic flows, domes
		k K-feldspar porphyry
		b basalt
		b' basalt, mudstone, ar
?	--------	----------
	ARCHEAN?	?

Jackson

EXPLANATION

onglomerate

---------- unconformity ----------

ainly HORNBY CHANNEL FORMATION: sandstone

---------- inferred unconformity ----------

eldspathic wacke, silty mudstone, quartzose sandstone, tuff

feldspathic sandstone, dolostone

ıber: dolostone, feldspathic sandstone

er: feldspathic sandstone

e member: feldspathic and dolomitic sandstone; laminated, argillaceous, and sandy dolostone

r: feldspathic sandstone

kes

n stratigraphic position

rate: conglomerate, sandstone, feldspathic granulestone

feldspathic granulestone, pebbly sandstone, mudstone, basalt.

nit: feldspathic sandstone, siltstone, basalt

ember: pebbly feldspatic sandstone, conglomerate, siltstone, basalt.

er (submembers 1 and 2): feldspathic sandstone, siltstone, lapilli-ash tuff, basalt

ite porphyritic dikes

, and intrusions

nd sandstone of Lower Igneous Complex

---------- inferred unconformity ----------

05'

113°00'

EXPLANATION

Flaceous, and sandy dolostone

alt.

-ash tuff, basalt

Geological contact (defined, approx

Bedding (inclined, overturned)

Cleavage

Fault (defined, approximate, assume

MAP

GEOLOGICAL MAP OF THE

WILSON ISLA

GREAT SLAVE

By B. J. JOHN

Geology mapped by B

ximate, assumed)

ed)

55'

MAP IA

GEOLOGICAL MAP OF THE WIL

WILSON ISLAND A

GREAT SLAVE LAKE,

By B. J. JOHNSON, 1987

Geology mapped by B J Johnson (1984, 198

50′

112°45′

SON ISLAND GROUP,

REA,

N. W. T.

p felsic flows, dome

k K-feldspar porphyr

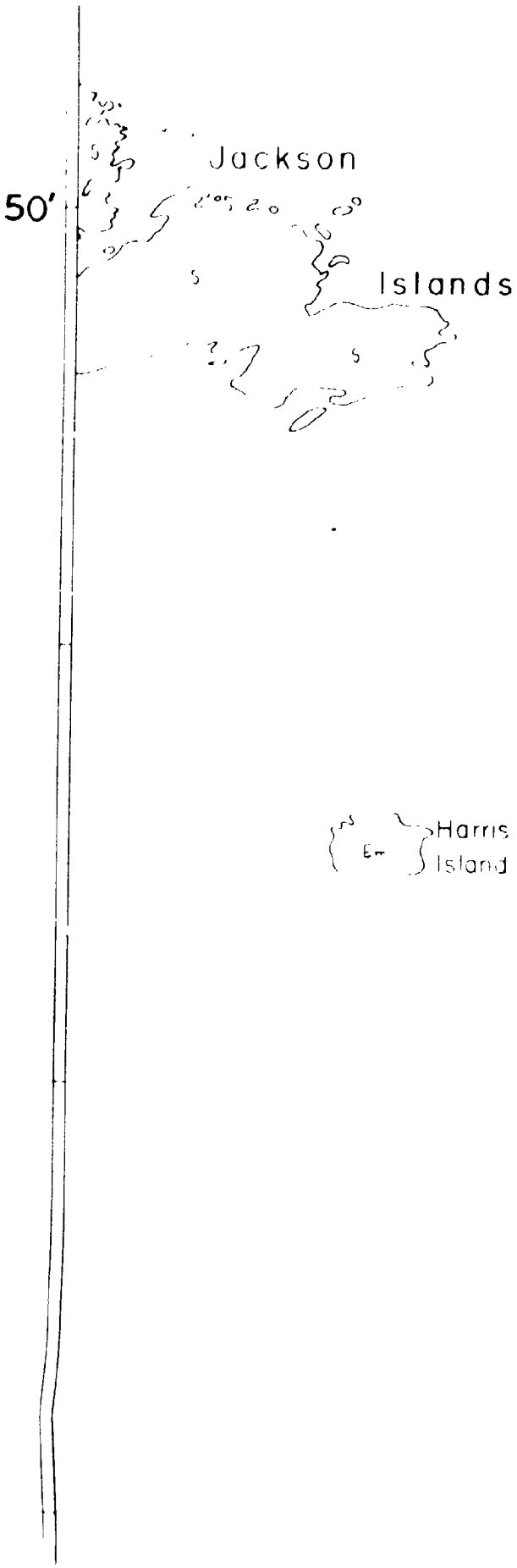
b basalt

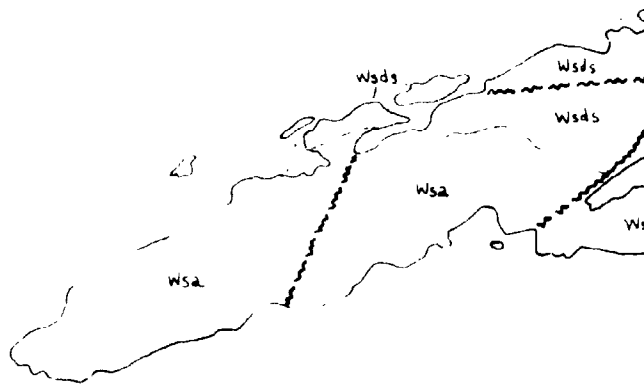
b· basalt, mudstone, a

? -------- ----------

ARCHEAN? A undifferentiated g

* Proterozoic rocks are desc





s, and intrusions

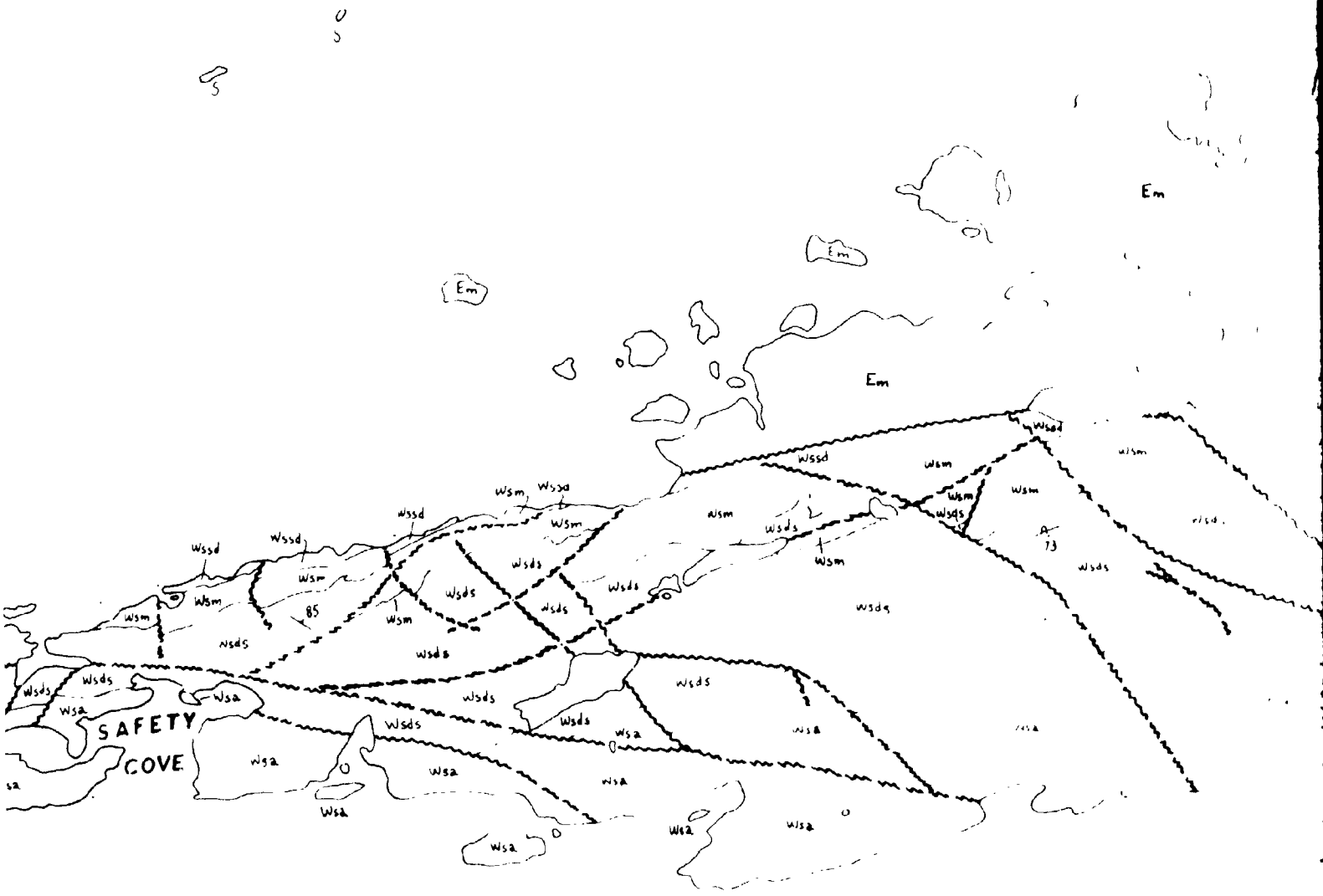
y

and sandstone of Lower Igneous Complex

---------- inferred unconformity ----------

ranitic and gneissic rocks

ribed in terms of their unmetamorphosed equivalents




WILSON
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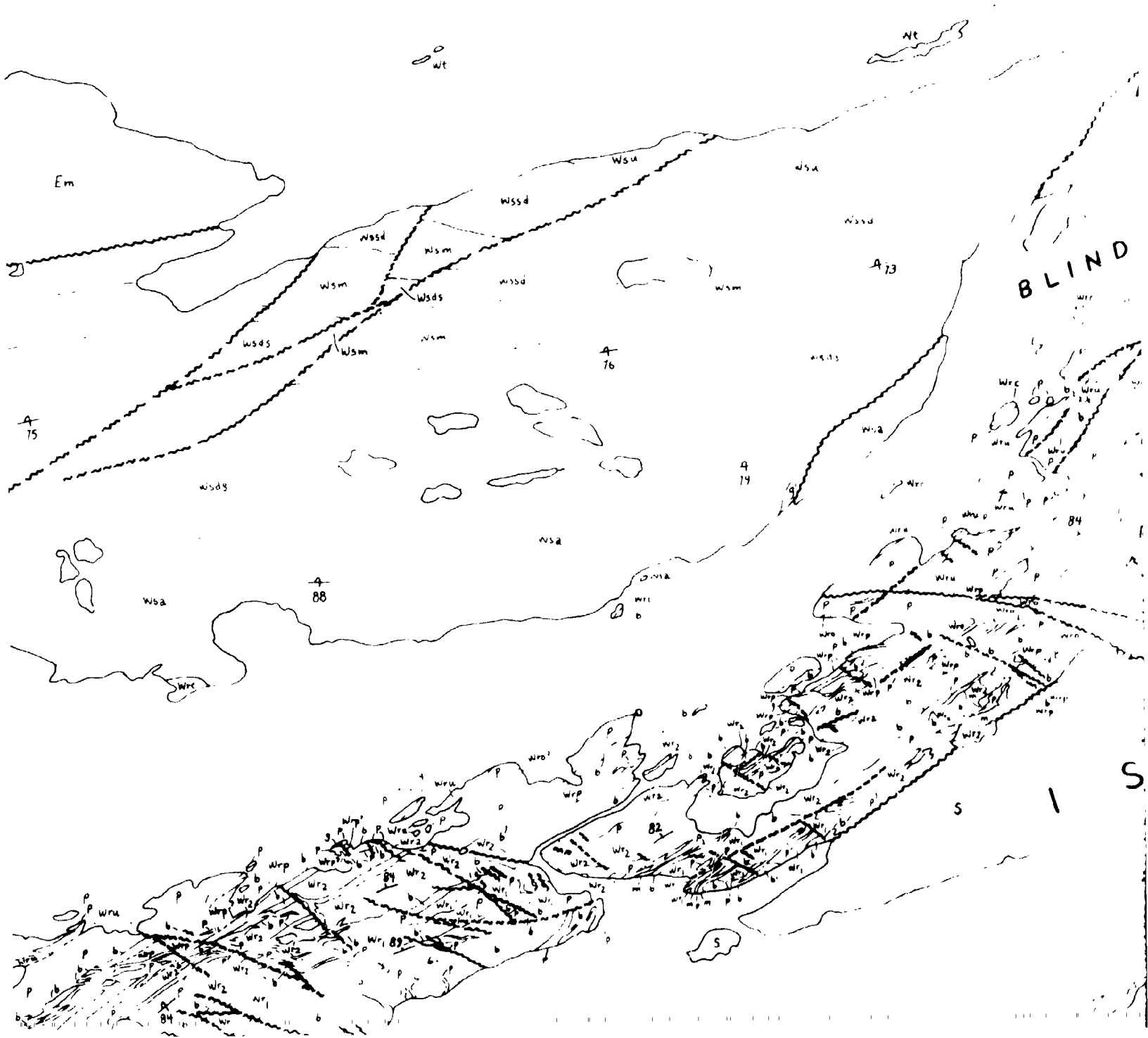
Em
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Wsm
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Wr2
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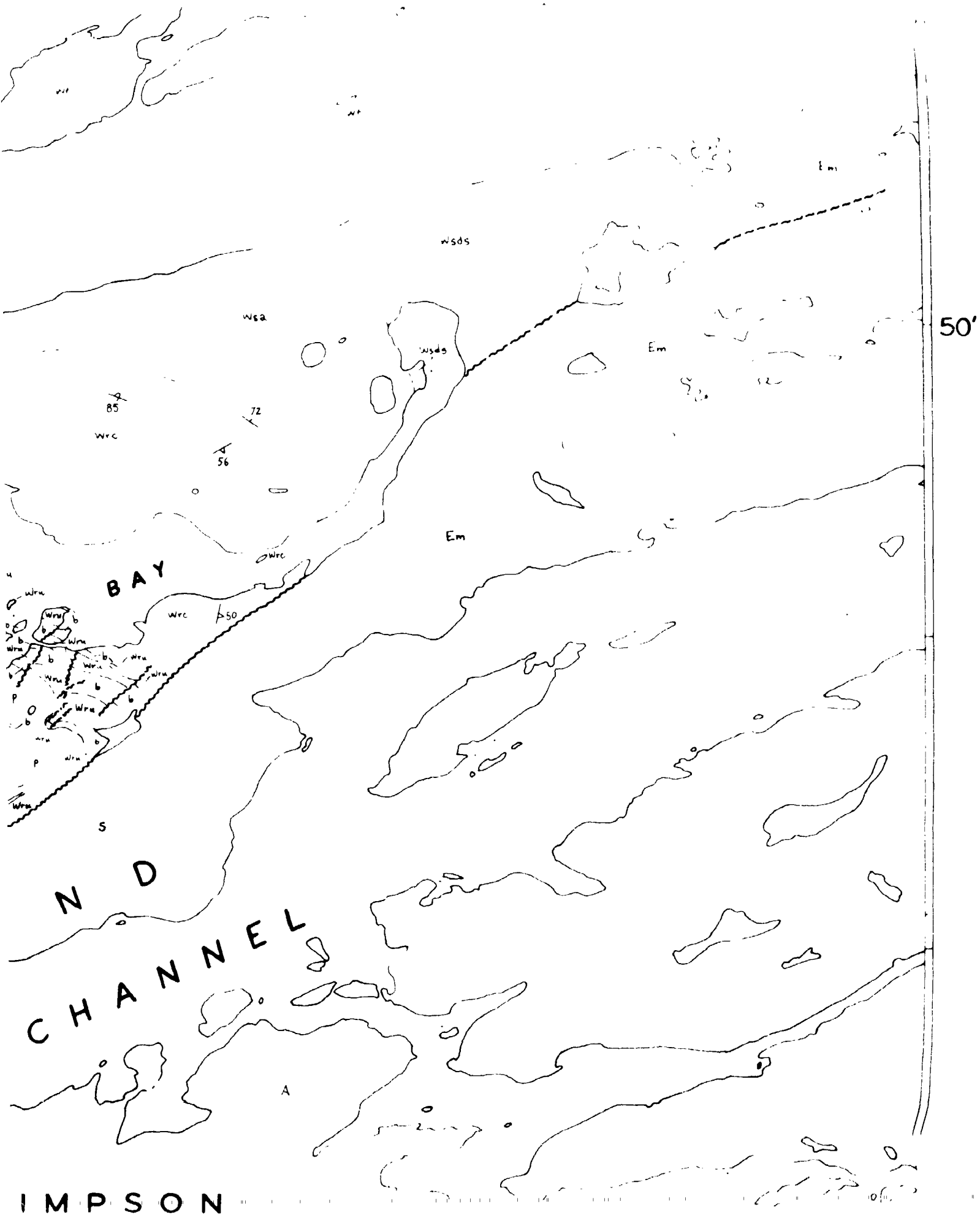
BLIND
ISLA


Em
wsds
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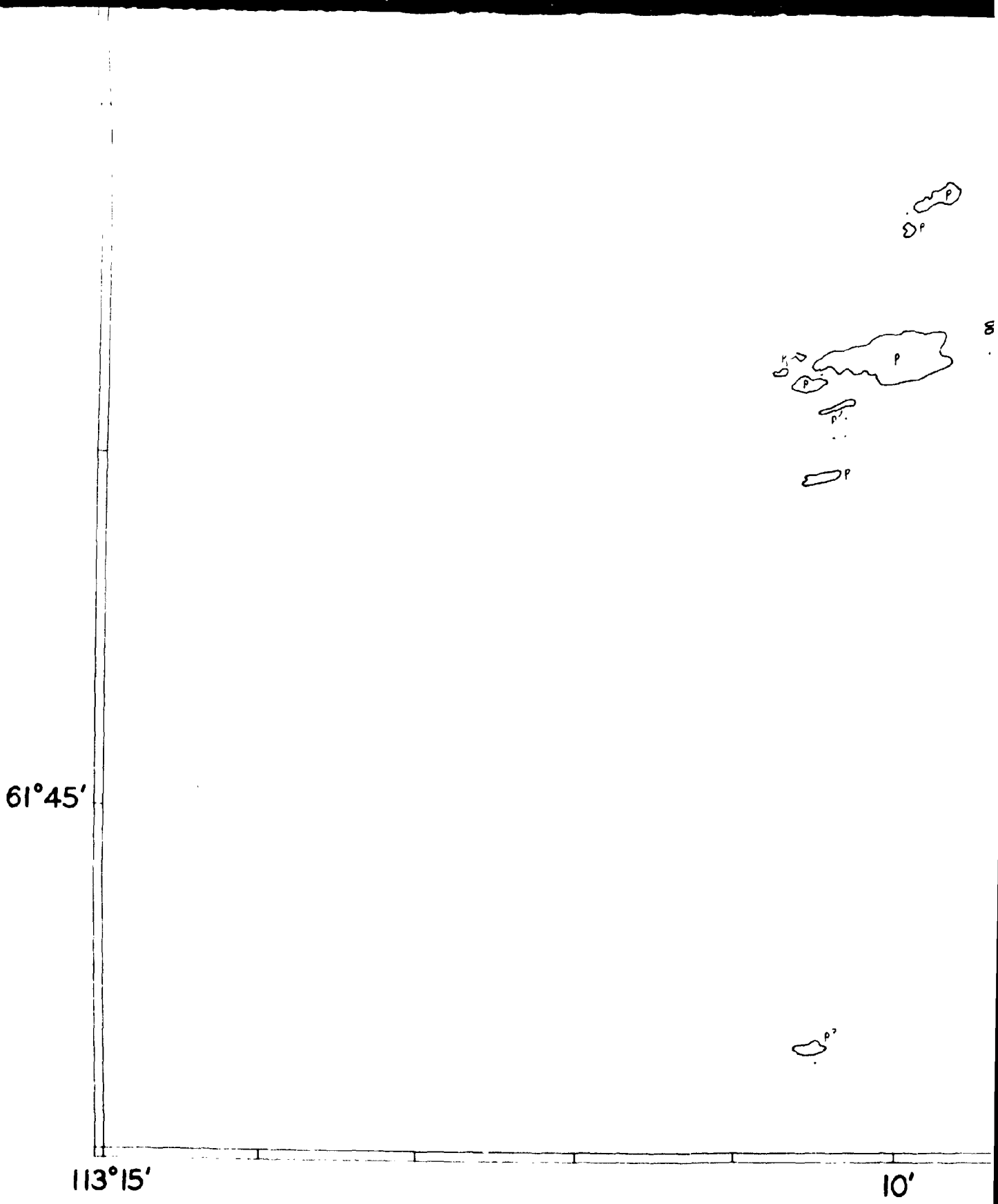
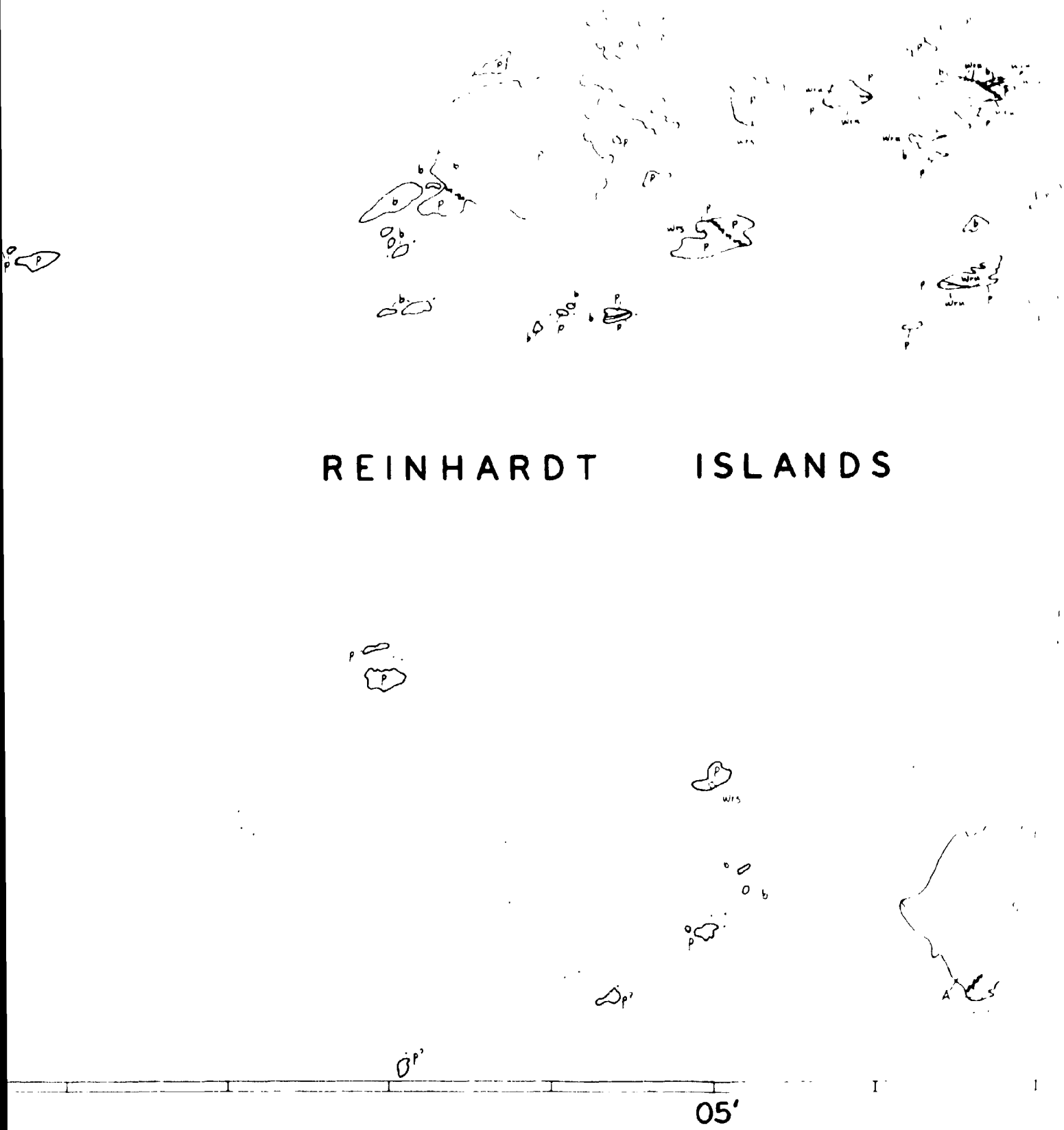
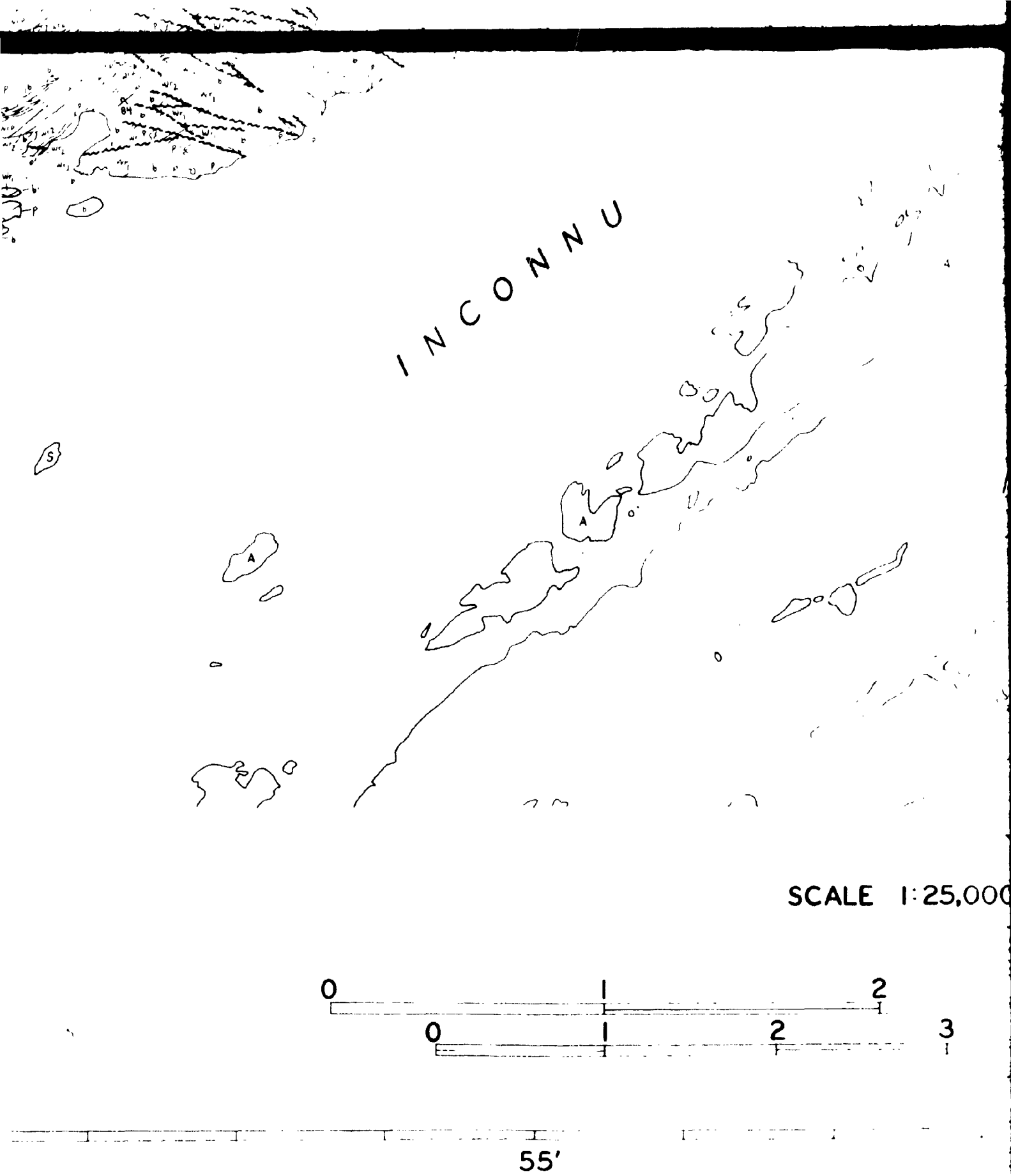

61°45′
113°15′
10′


REINHARDT ISLANDS
05'


ISLANDS
05'
113°00'


INCONNU
SCALE 1:25,000
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1
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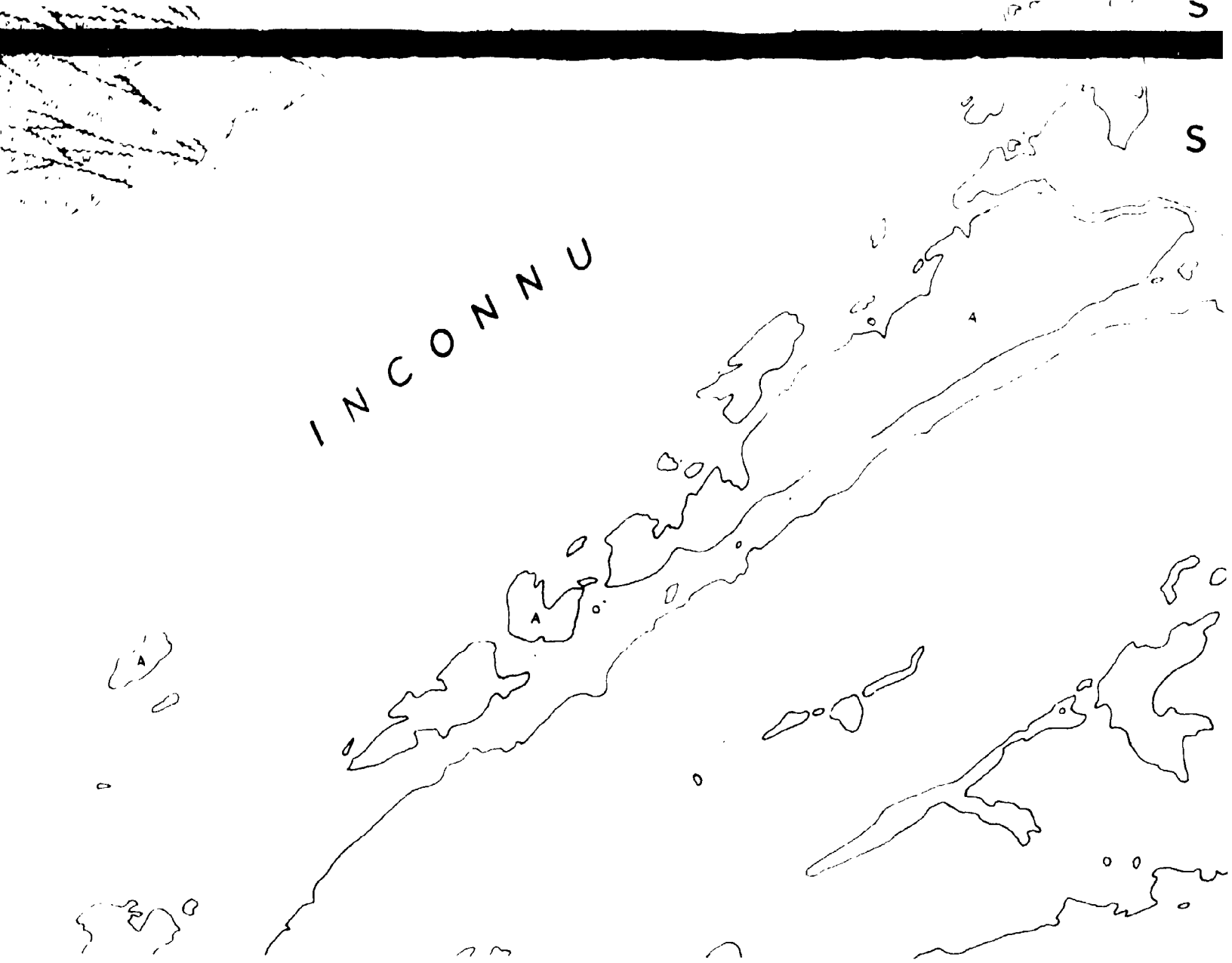
2
3
55'



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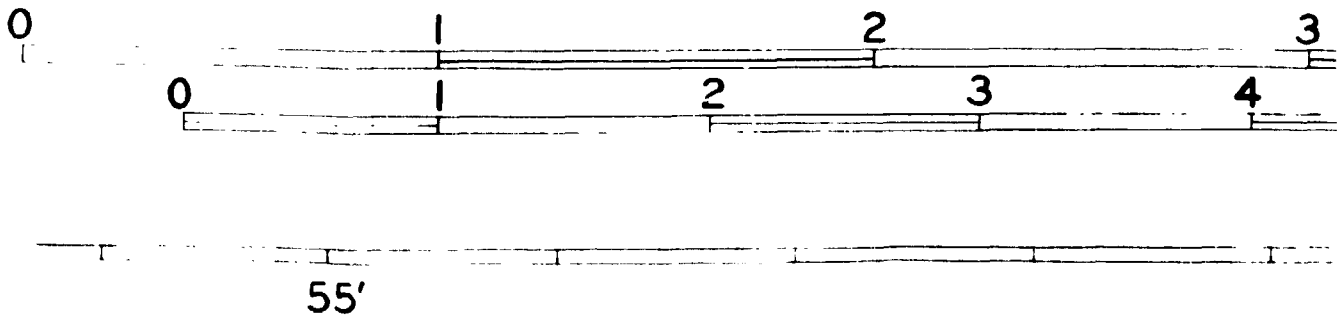


IMPSON

IMPSON

ISLANDS

61°45'

4 MILES

5 KILOMETRES

50'

112°45'

50′

EXPLANATION

Bedding (tops known, tops unknown, ove

Cleavage (in pelite, in quartzite, in ign

Bedding - cleavage intersection

Stretching lineation

Axial plane of crenulations

Crenulation axis

Minor fold axis

Axial surface trace of overturned antic

Axial surface trace of overturned sync

Form surface defined by geological c

Fault (defined, approximate, assumed)

05′

45′ 40′

rturned)

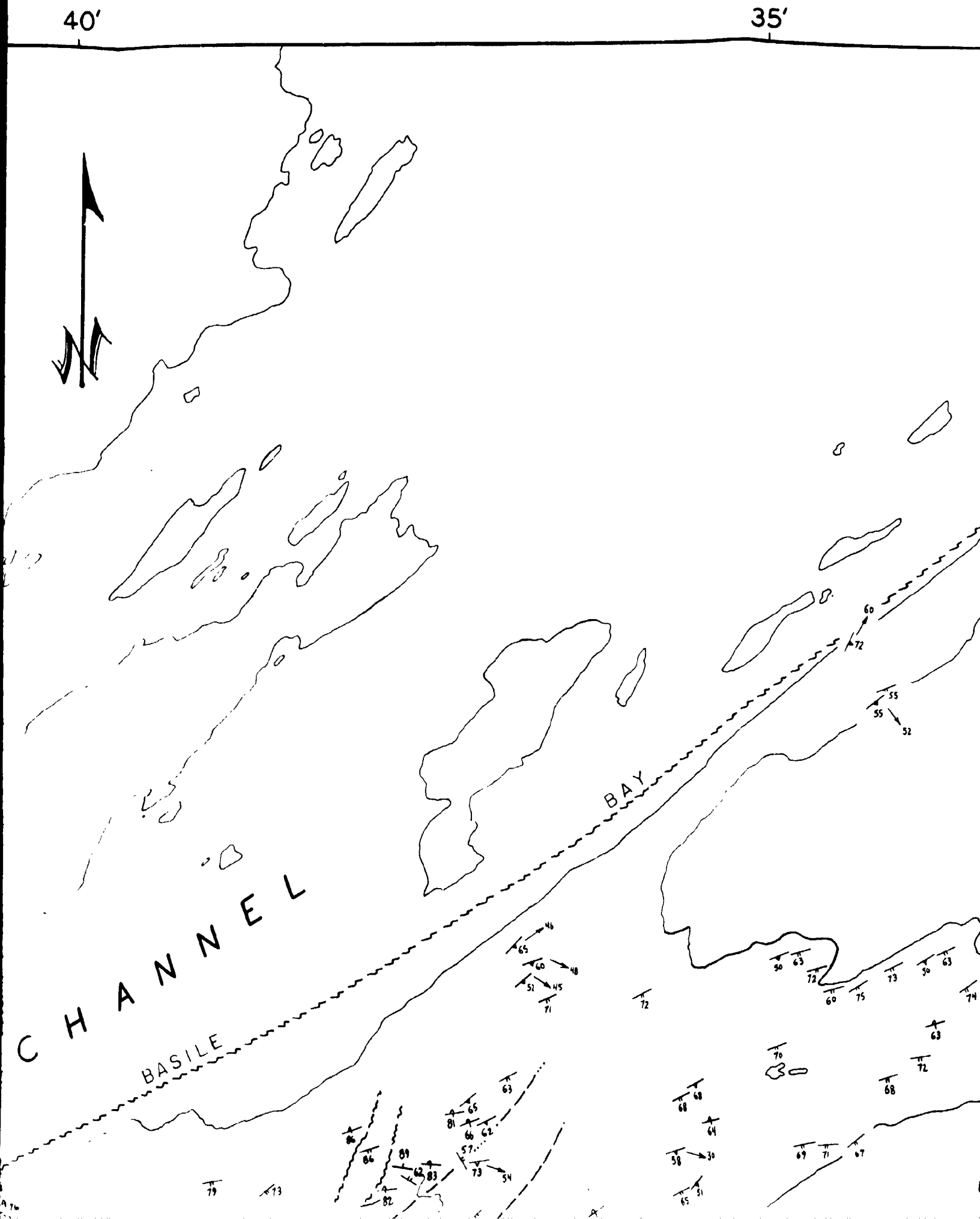
eous rocks)

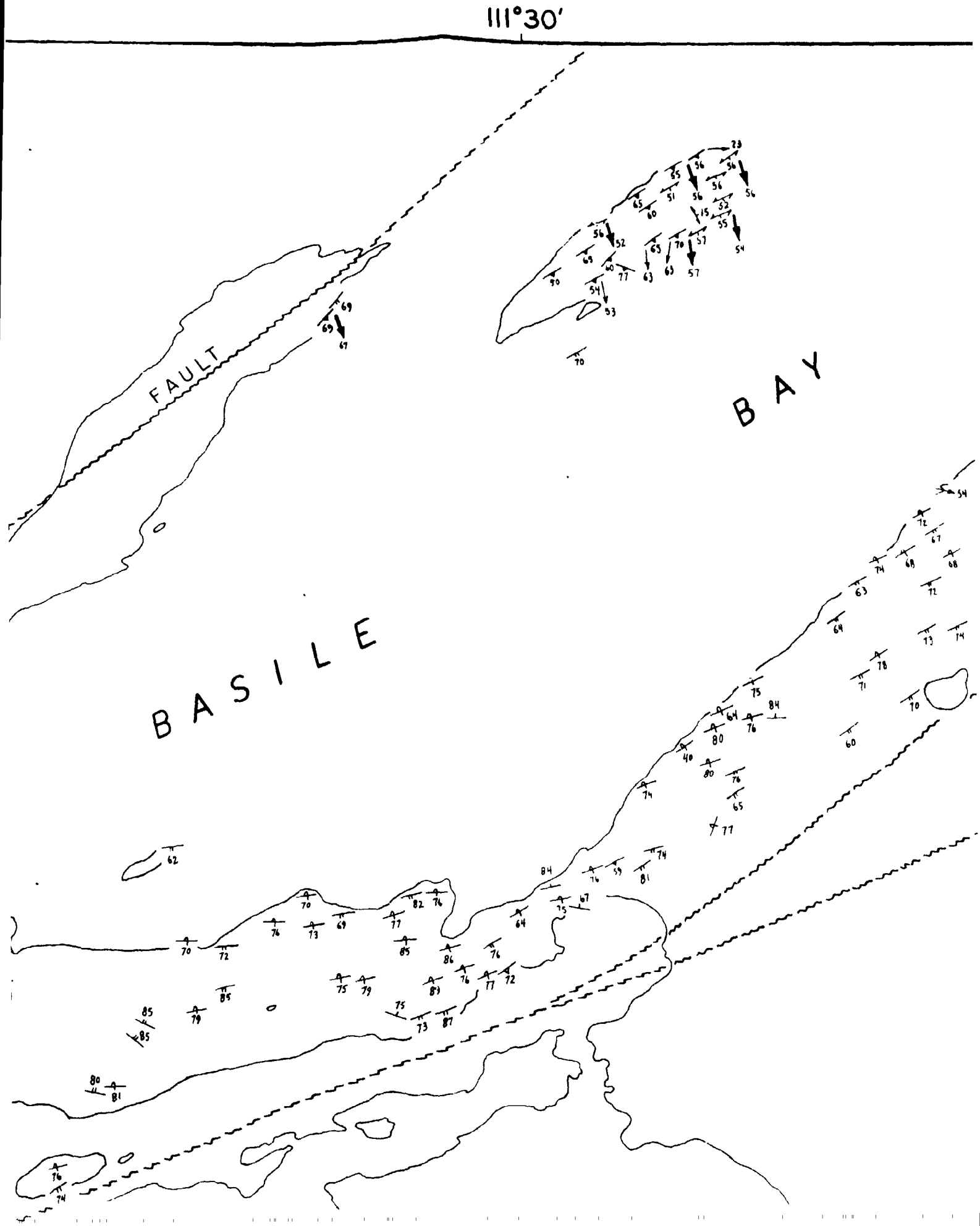
line

ine

ontact

ND


40'
35'
CHANNEL
BASILE
BAY


111°30'
FAULT
BAY
BASILE

EXPLANATION

Bedding (tops known, tops

Cleavage (in pelite, in qu

Bedding - cleavage interse

Stretching lineation

Crenulation cleavage

Minor fold axis (upward

Axial surface trace of p

Axial surface trace of p

Form surface

Fault (defined, approxim

unknown, overturned)

artzite and coarse-grained clastic rocks, in igneous rocks)

ction

facing, downward facing, facing unknown)

unging anticline

unging syncline

ite, assumed), arrow shows dip of fault plane

rocks)

MAP 1B

STRUCTURAL GEOLOGICAL MAP OF THE WILSON ISLAND AREA GREAT SLAVE LAKE, N

By B.J. JOHNSON, 1987

Geology mapped by B.J. Johnson (1984, 1985)

MAP 1B

TURAL GEOLOGICAL MAP OF THE WILSON

WILSON ISLAND AREA,

GREAT SLAVE LAKE, N. W. T.

By B. J. JOHNSON, 1987

Geology mapped by B. J. Johnson (1984, 1985)

ISLAND GROUP,

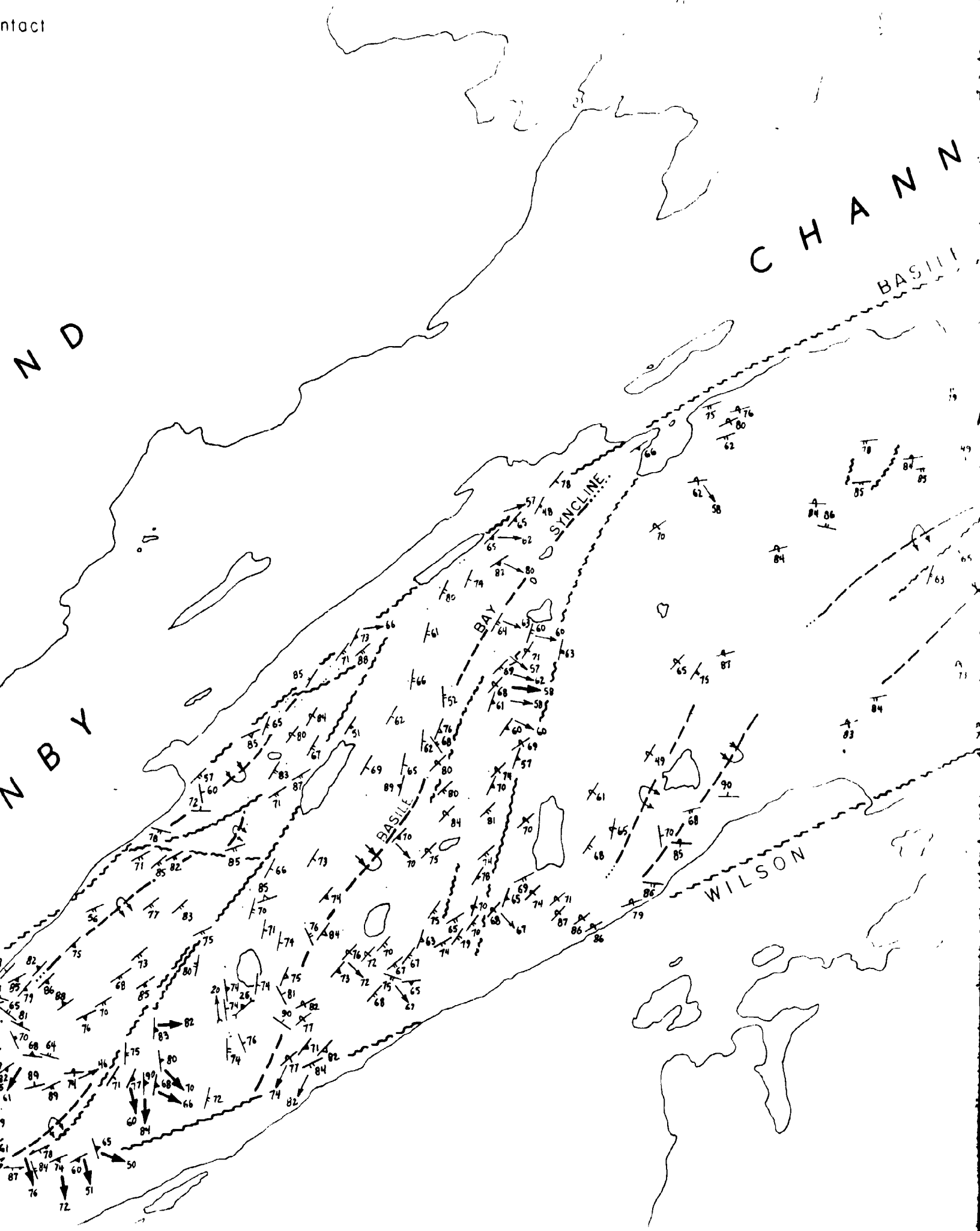
FAULT

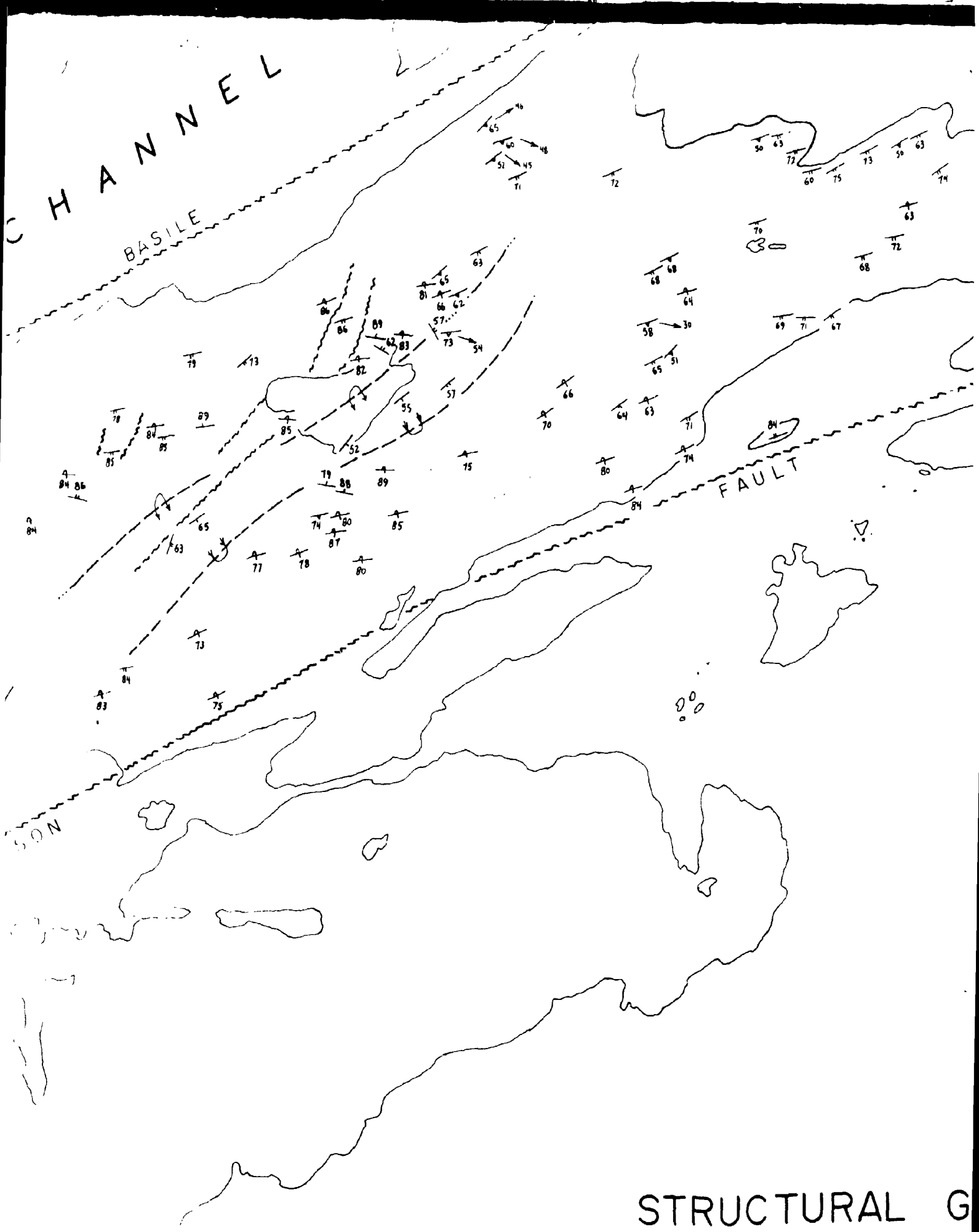
Form surface defined by geological co

Fault (defined, approximate, assumed)

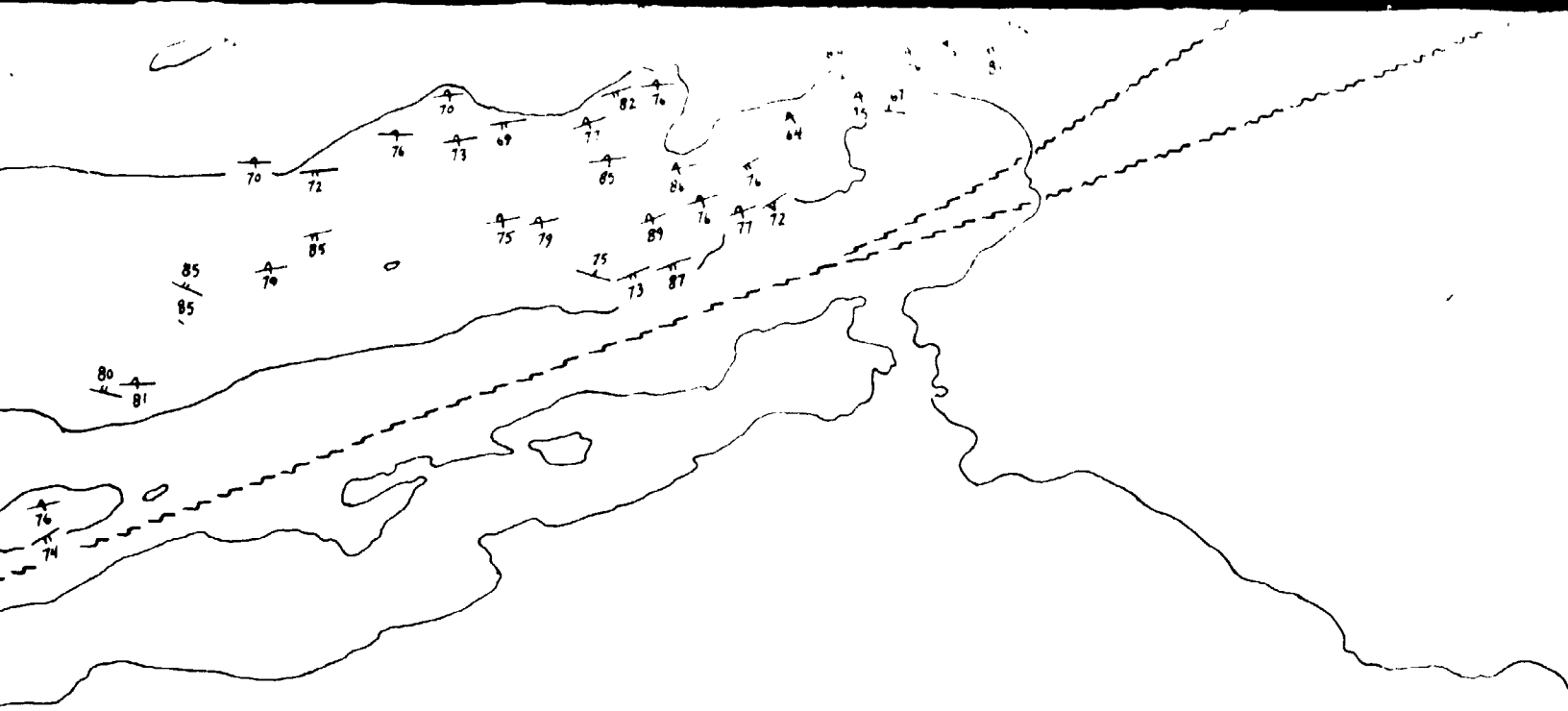
05'

ntact



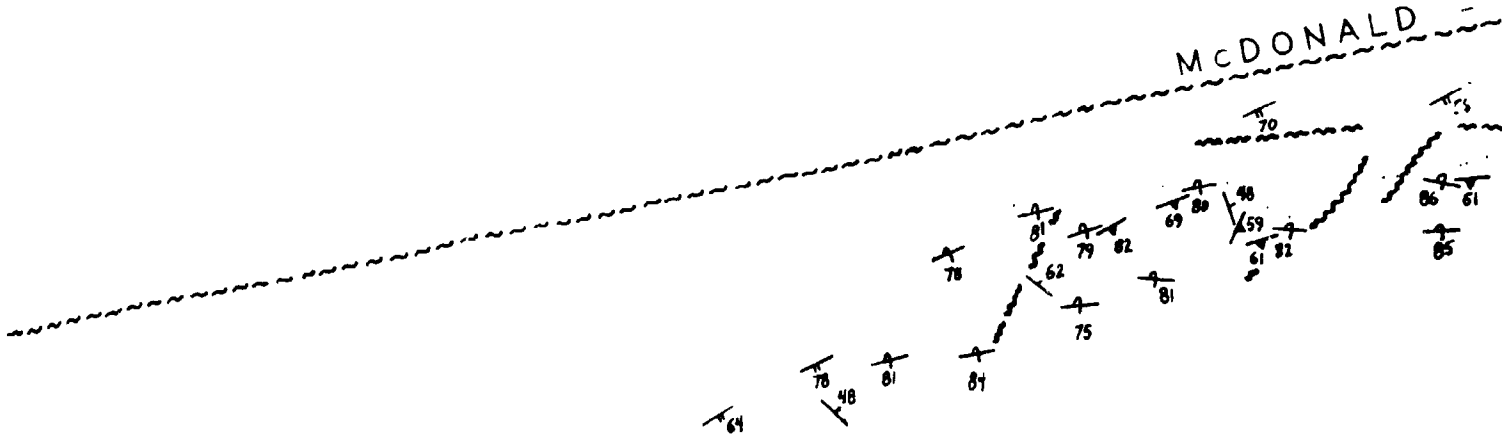


STRUCTURAL G



MAP 2B

GEOLOGICAL MAP OF THE


McDONALD
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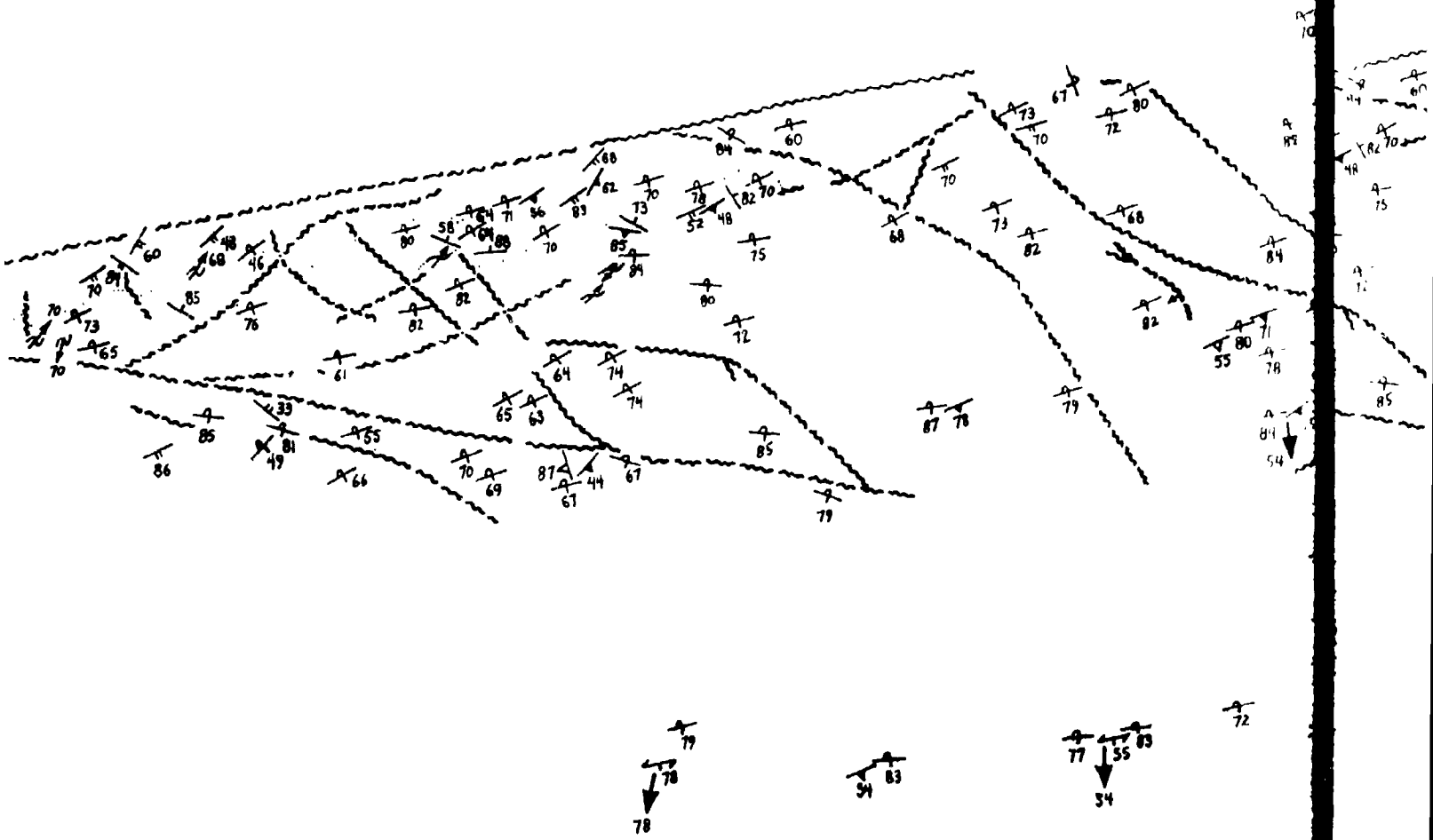
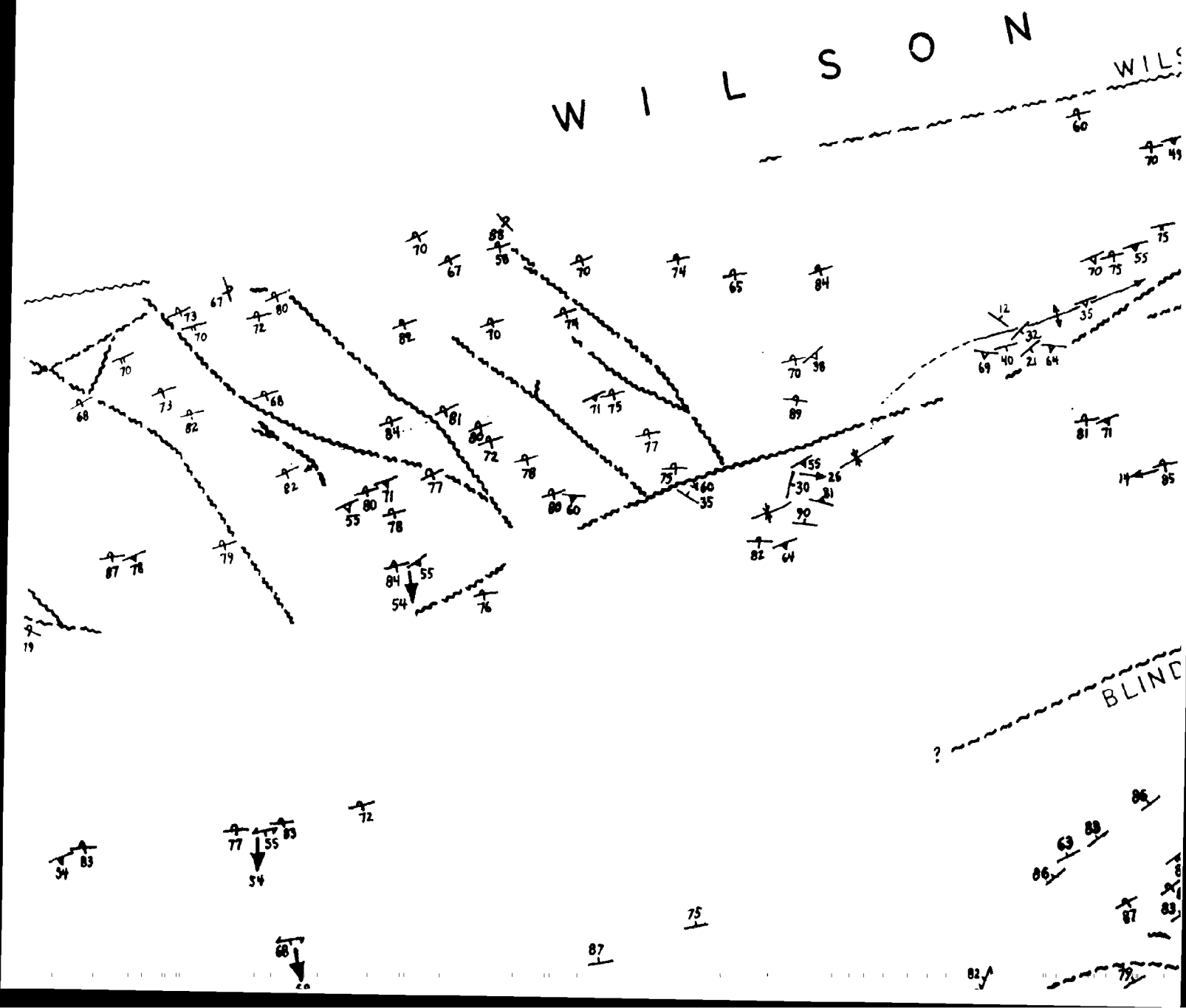
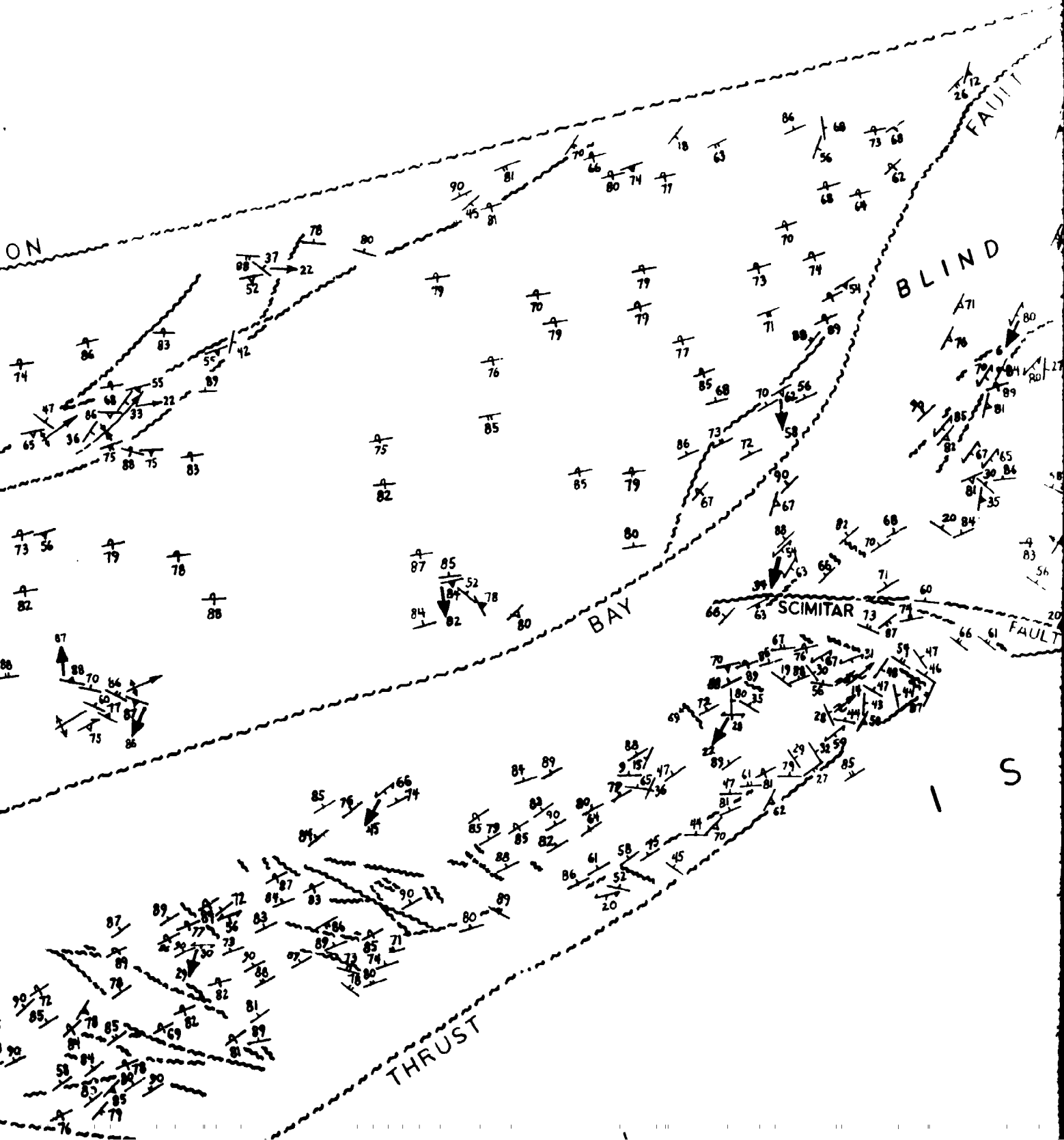
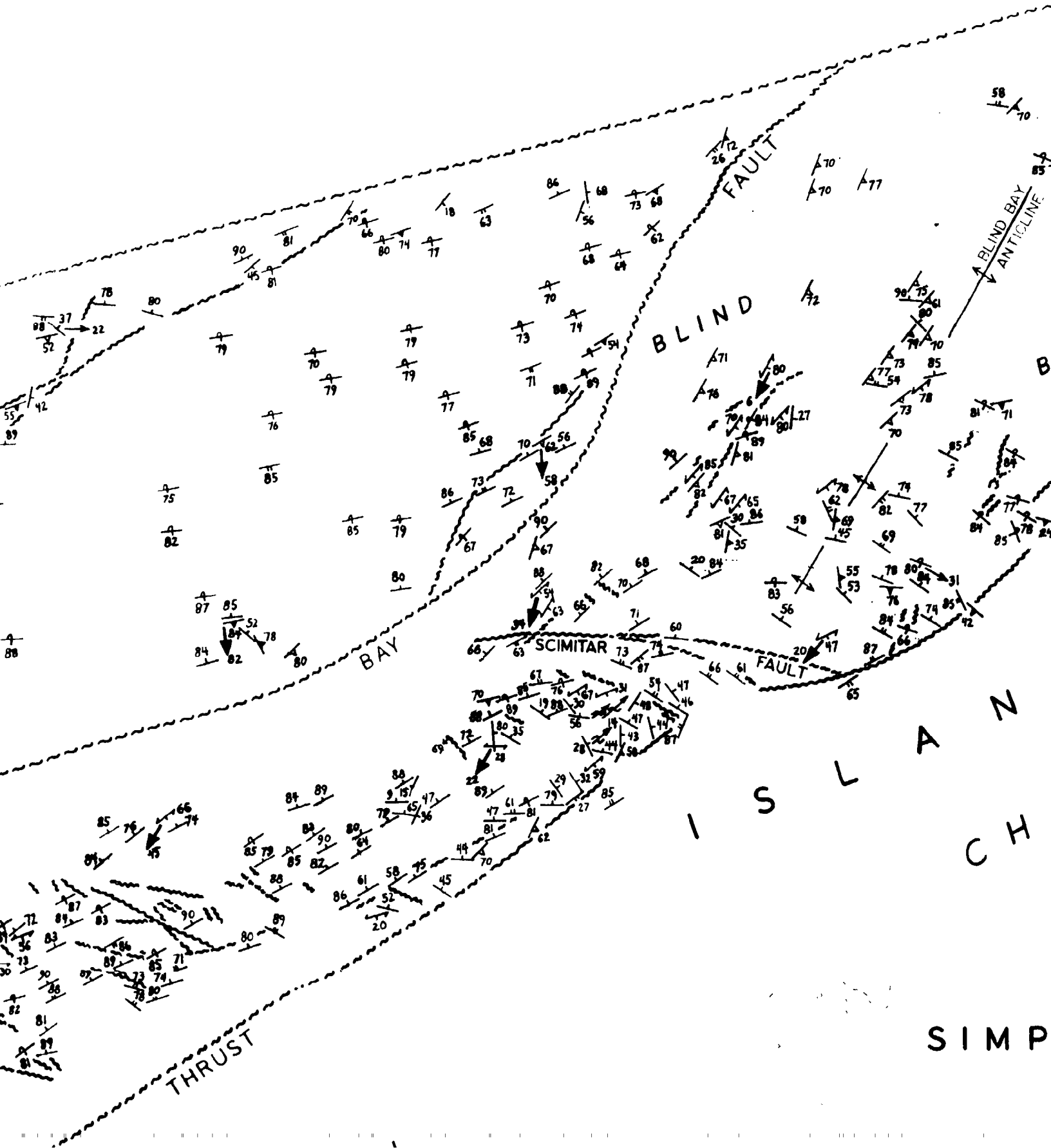



WILSON
WILS
BLIND


ON
BLIND
FAULT
BAY
SCIMITAR
FAULT
THRUST
S


BLIND
FAULT
BLIND BAY
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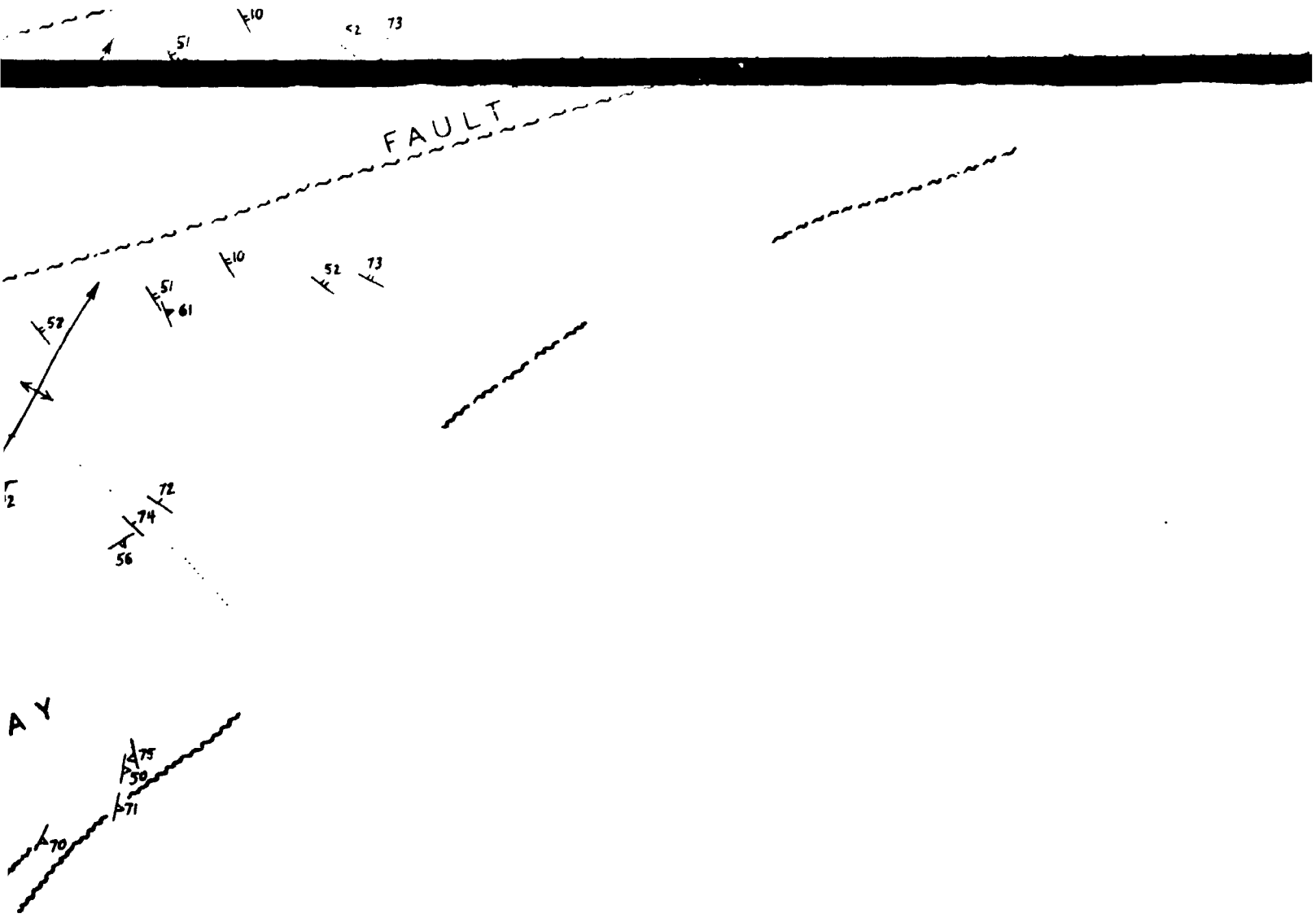
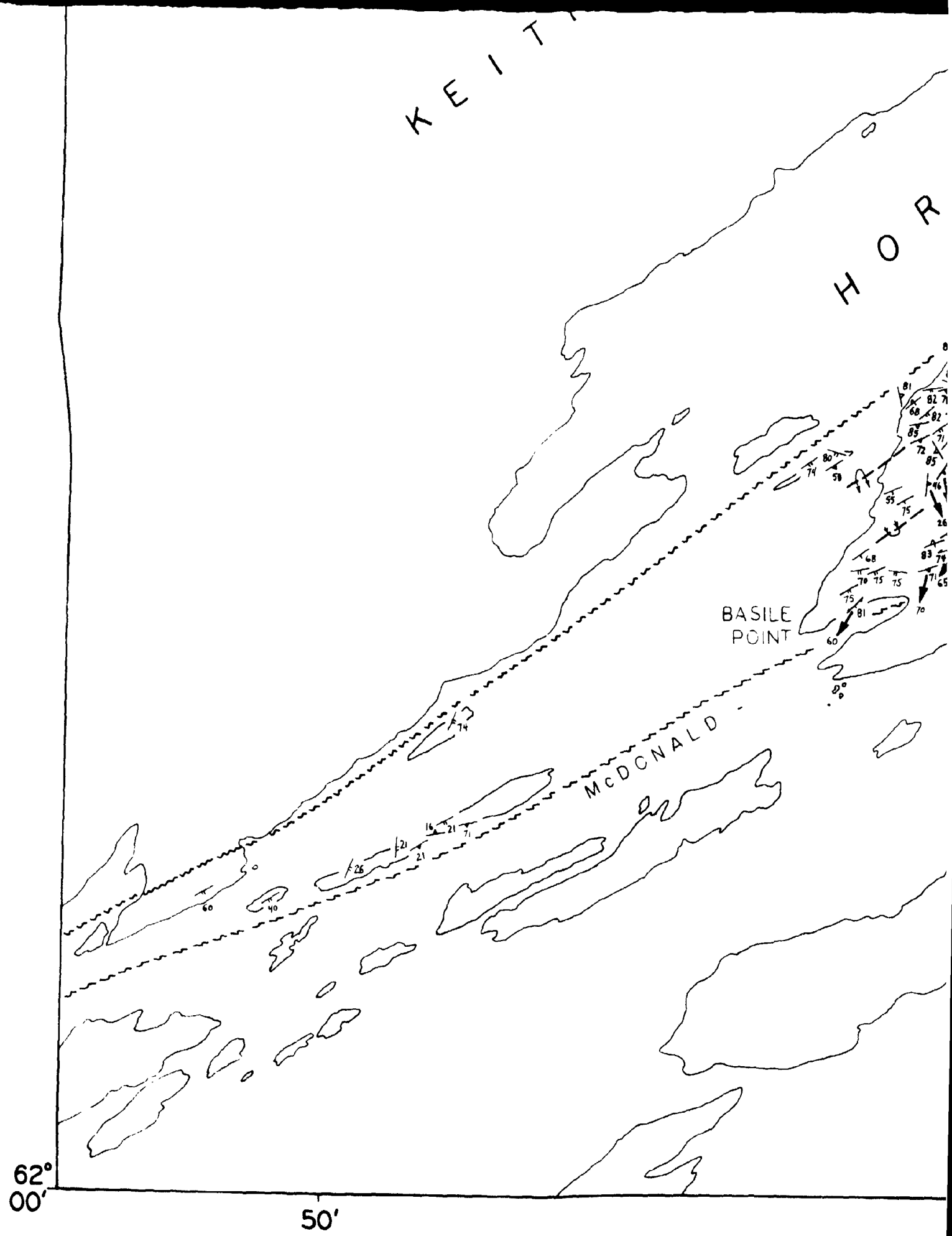


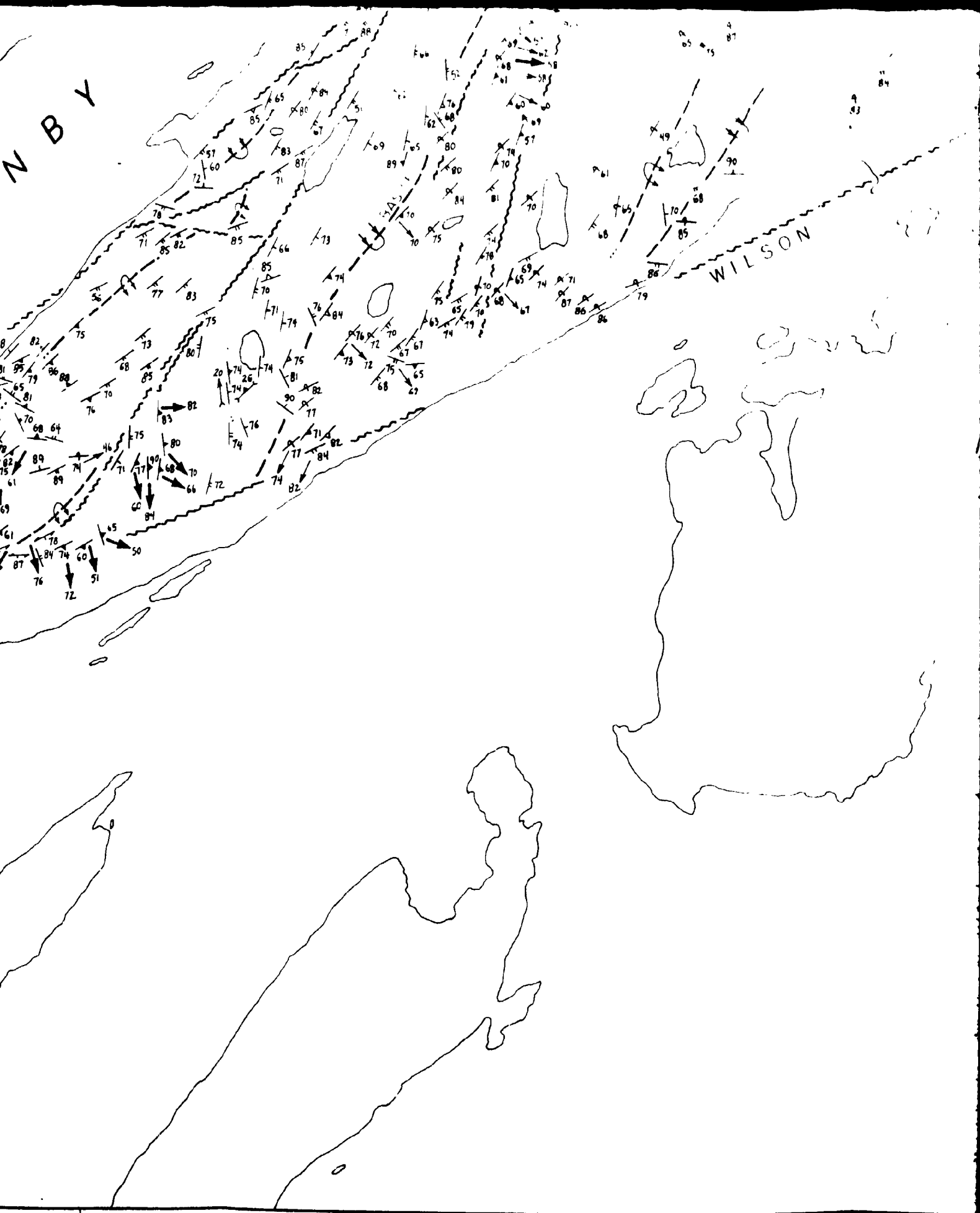
AY

D

ANNEL

SON


KEITH
HOR
BASILE
POINT
McDONALD
62°
00'
50'


NBY
WILSON
45'
40'

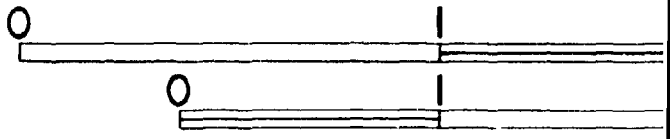
STRUCTURAL G

WILSON ISLAND

BASILE BAY, GE

By E

Geology mapped by B. J. Johnson (19



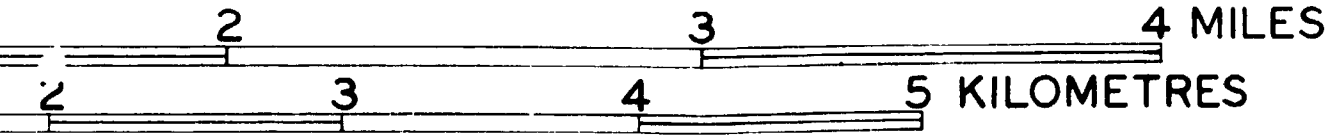
40′ 35′

MAP 2B

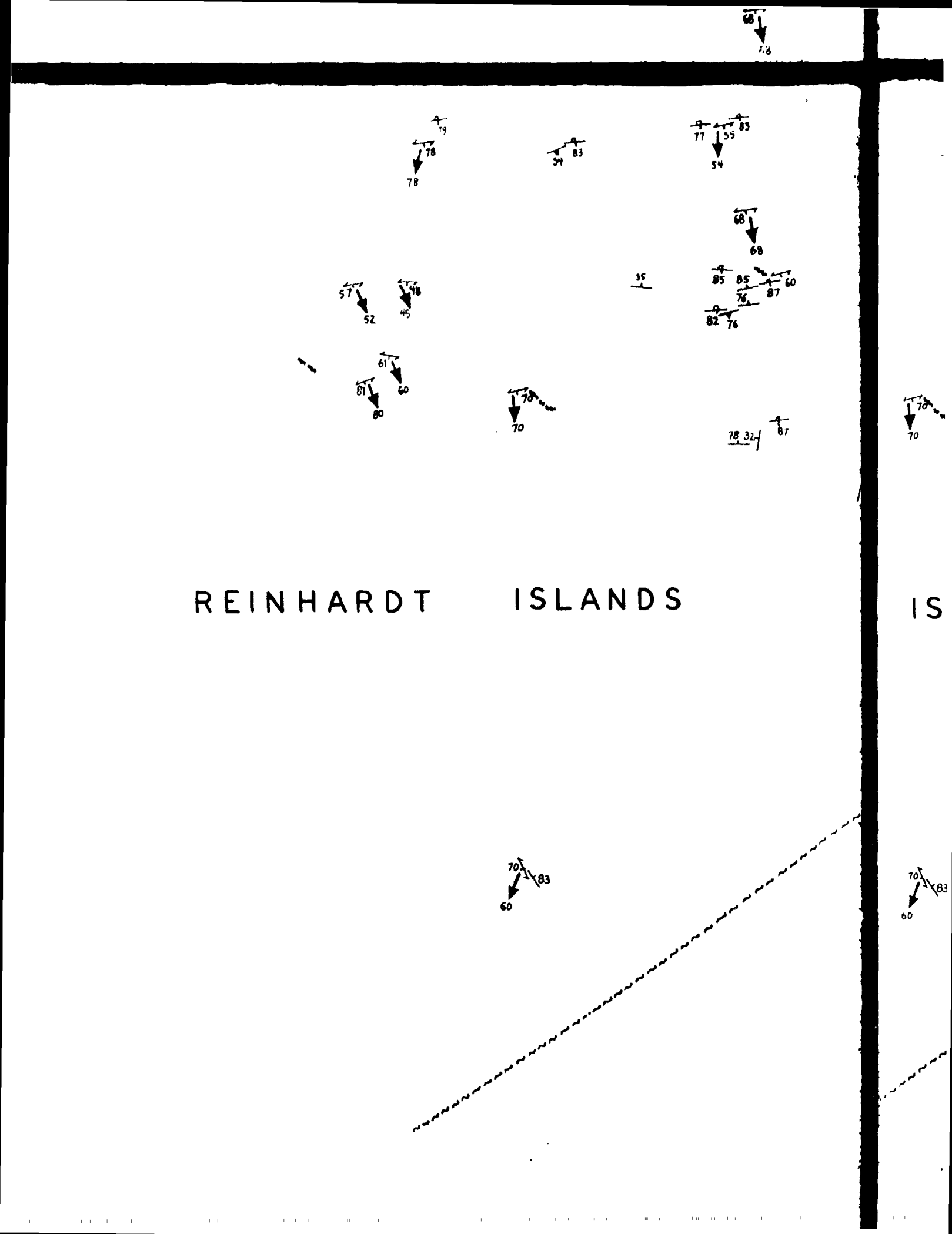
EOLOGICAL MAP OF THE GROUP SOUTHWEST OF REAT SLAVE LAKE, N.W.T.

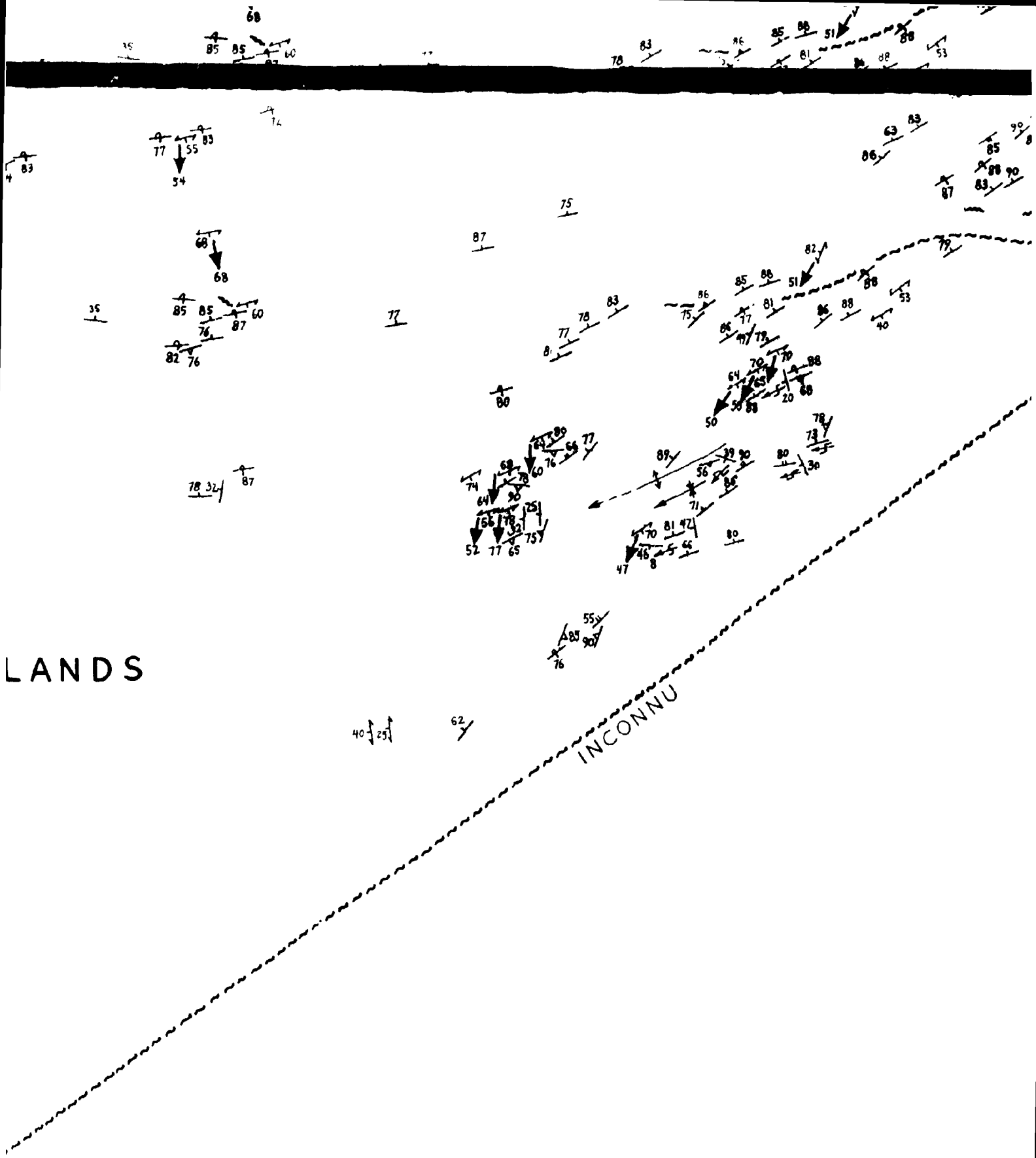
B. J. JOHNSON, 1987

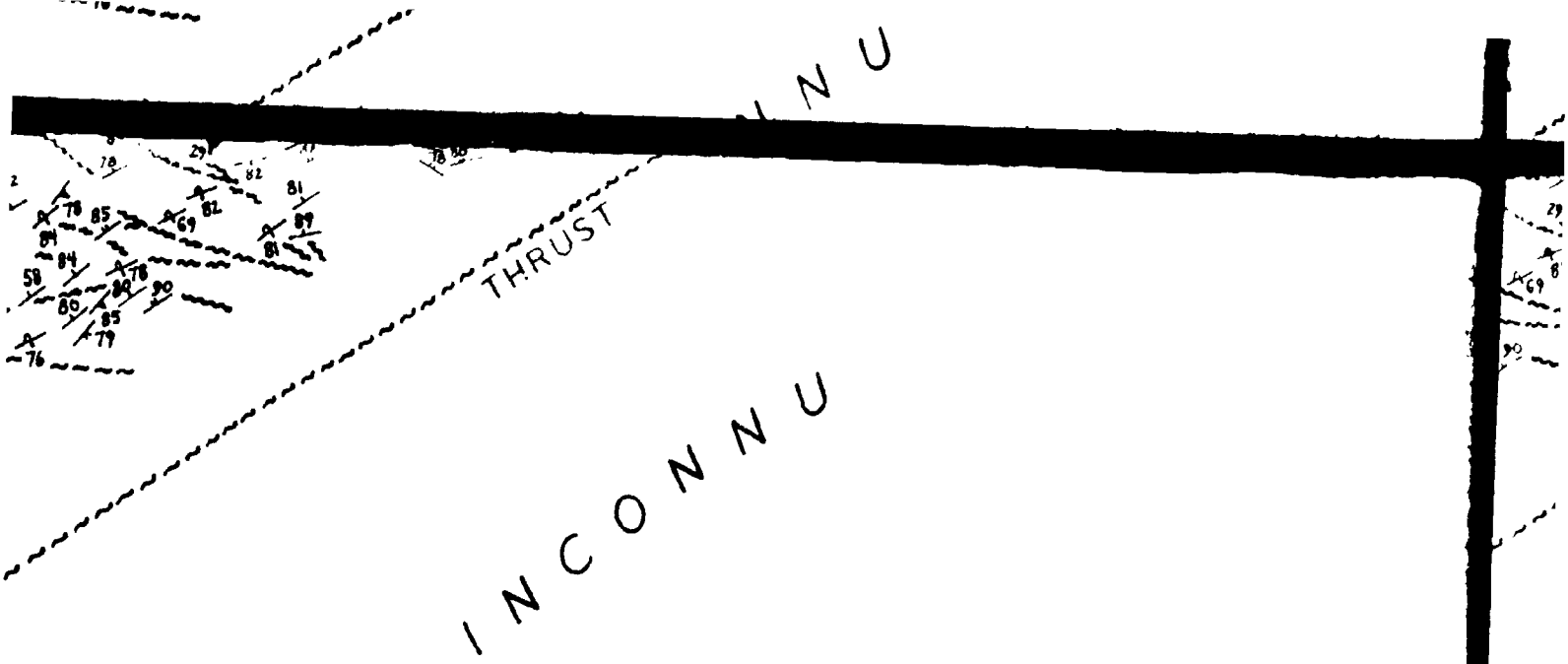
984, 1985), A. Enge (1984), C. D. Gault (1985), and J. M. Wahlroth (1985)



111°30′


REINHARDT ISLANDS


LANDS
INCONNU



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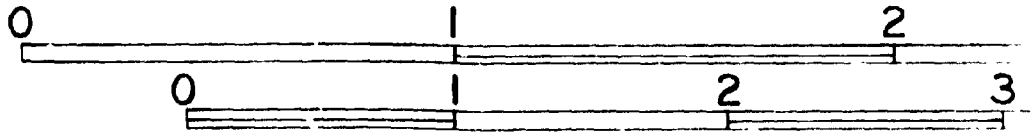

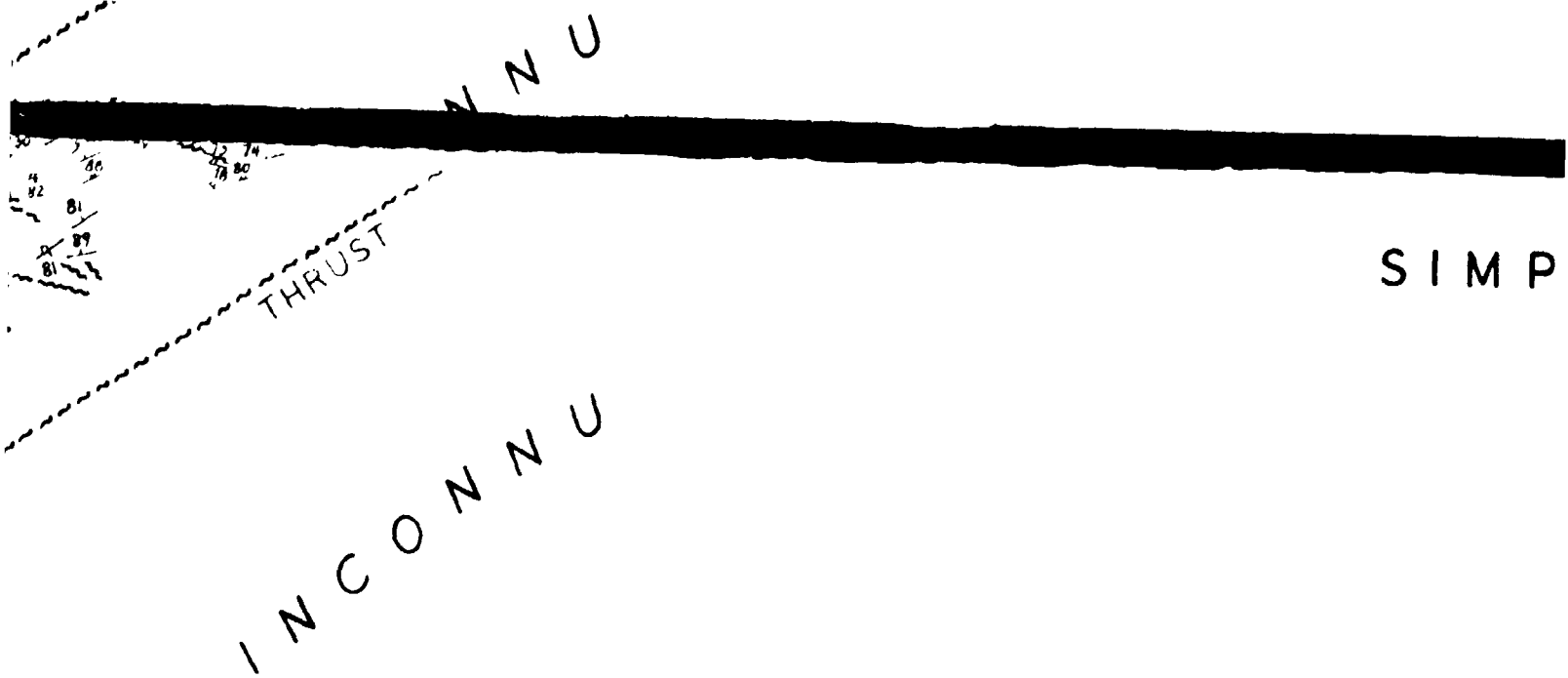

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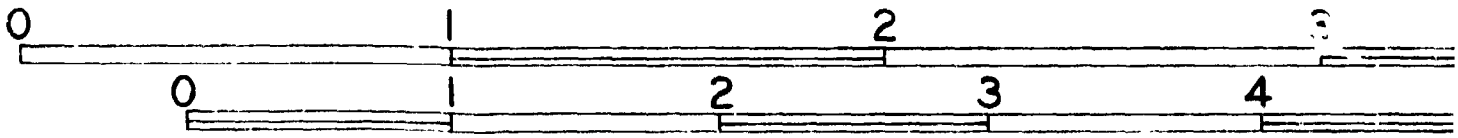


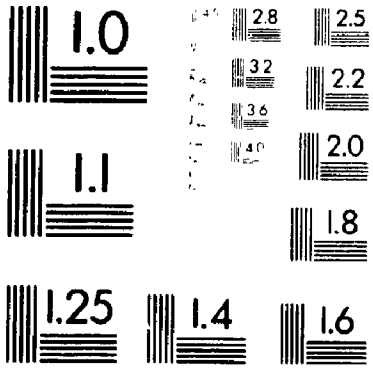
SON

ISLANDS

4 MILES

5 KILOMETRES

6 of/de 6



62° 15′ 50′

PROTEROZOIC	HELIKIAN	Md	Mackenzie diabas
		FG	Fortress Gabbro
		ET-THEN GROUP	
		E	undifferentiated
		GREAT SLAVE SUPERGROUP	
		GS	undifferentiated
		UNION ISLAND GROUP	
		U	undifferentiated
		WILSON ISLAND GROUP	
		BASILE FORMATION	
	APHEBIAN	Wbv	basalt or
		Wbm	mudstone
		Wbb	basalt
		Wbh	Graded member: s
		Wbf	Ironstone member
		Wbuq	Upper Quartzite
		Wbp	Lower Pelitic me
		Wblq	Lower Quartzite
		Wbds	dolomitic sandst

45′ 40′

EXPLANATION

dikes

(includes Murky and Preble Formations) conglomerate, sandstone

--------- unconformity ----------

(includes Sosan, Kahochella, Pethei, and Christie Bay Groups)

--------- unconformity ----------

basalt, diabase, dolostone, mudstone

---- relationship uncertain ----

qr sodic granite

mp mafic porphyry

di diorite

(plagioclase-phyric)

andstone, argillaceous siltstone

sandstone, hematite-magnetite ironstone, calcareous pisolitic ironstone

ember: feldspathic sandstone

mber: argillaceous siltstone, sandstone

ember: feldspathic sandstone

ne

40′ 35′

mestone

oupes)

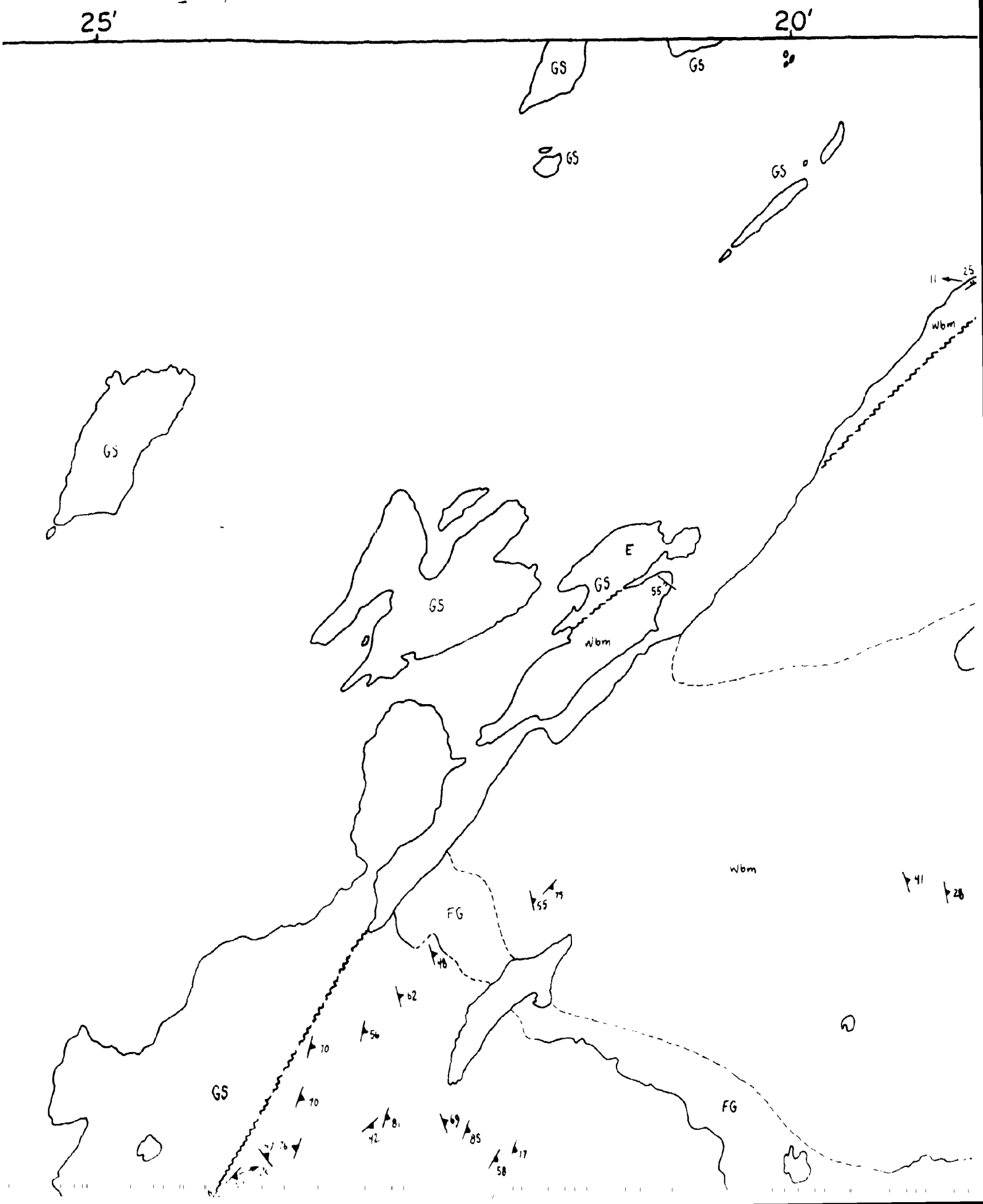
oolitic ironstone

G R

111°30'

E A T

S L A V E


25'
20'
GS
GS
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GS
GS
GS
E
GS
55
Wbm
Wbm
Wbm
FG
FG
GS
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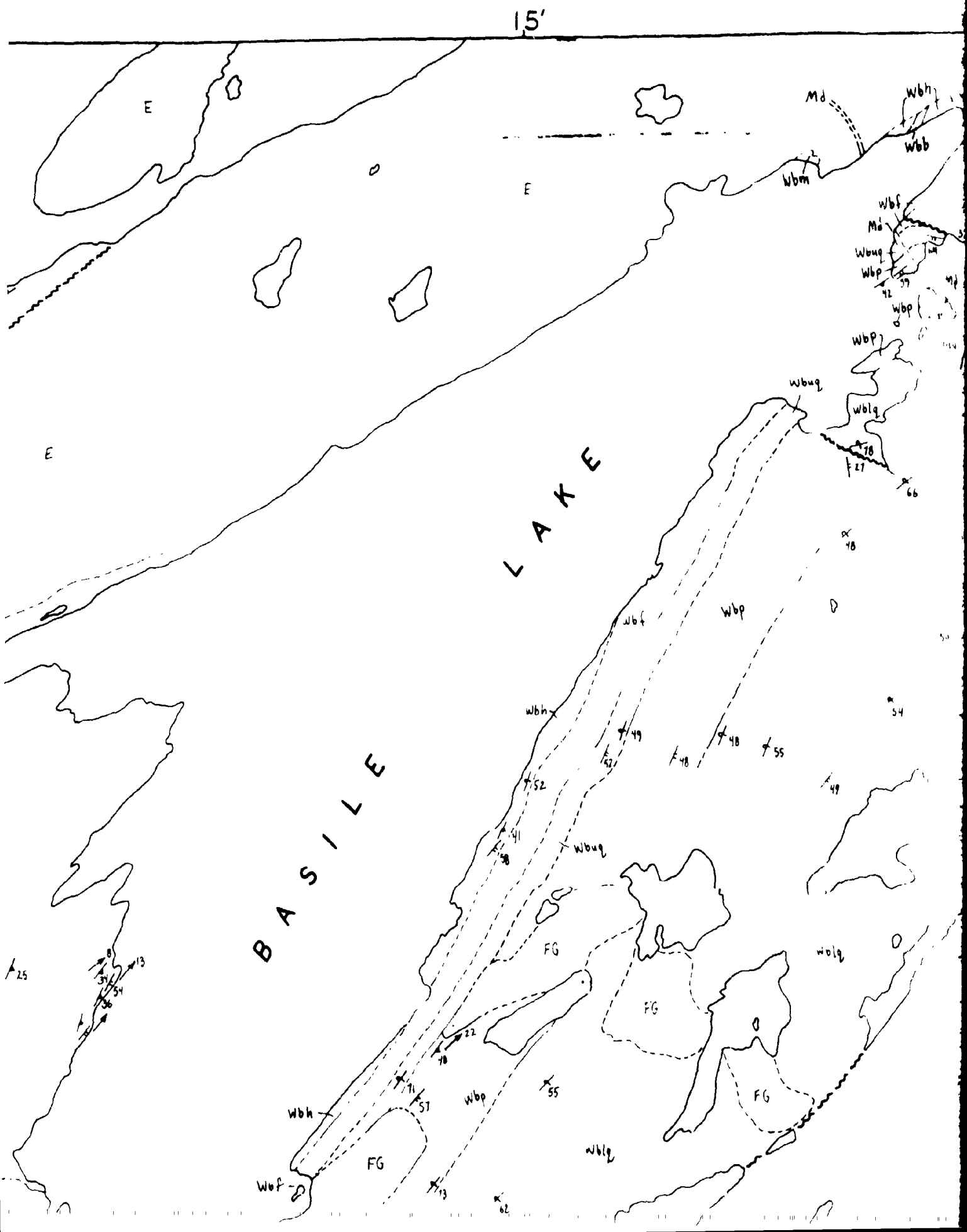
79
48
62
56
70
70
81
42
69
85
58
17
41
28
11
25


15'
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E
Md
Wbh
Wbb
Wbm
Wbf
Md
Wbuq
Wbp
Wbp
Wbp
Wbuq
Wblq
BASILE
LAKE
Wbf
Wbp
Wbh
Wbuq
FG
FG
FG
Wblq
Wbp
Wbh
FG
Wblq
Wbf


20'
15'
E
E
E
Wbm
BASILE
LAKE
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FG
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Wbh
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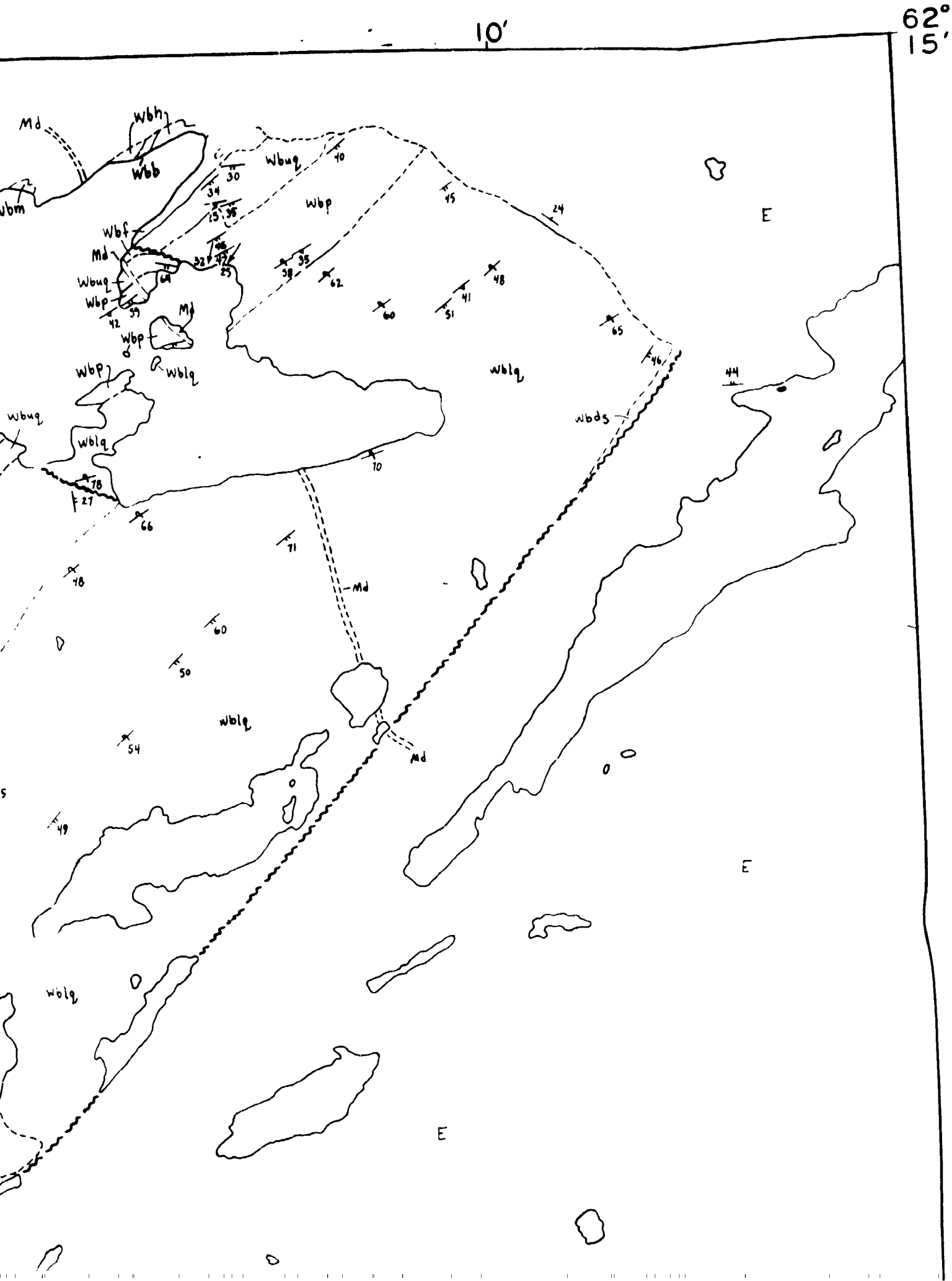
10'
62°
15'
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64
75

dstone, argillaceous siltstone

sandstone, hematite-magnetite ironstone, calcareous oolitic ironstone

mber: feldspathic sandstone

ber: argillaceous siltstone, sandstone

mber: feldspathic sandstone

e

---- inferred unconformity ----------

granitic, gneissic, and amphibolitic rocks

are described in terms of their unmetamorphosed equivalents

Geological contact (defined, approximate, assumed)

Bedding (tops known, tops unknown, overturned)

Cleavage (vertical, 15 plunging)

Bedding-cleavage intersection

Strike-slip lineation

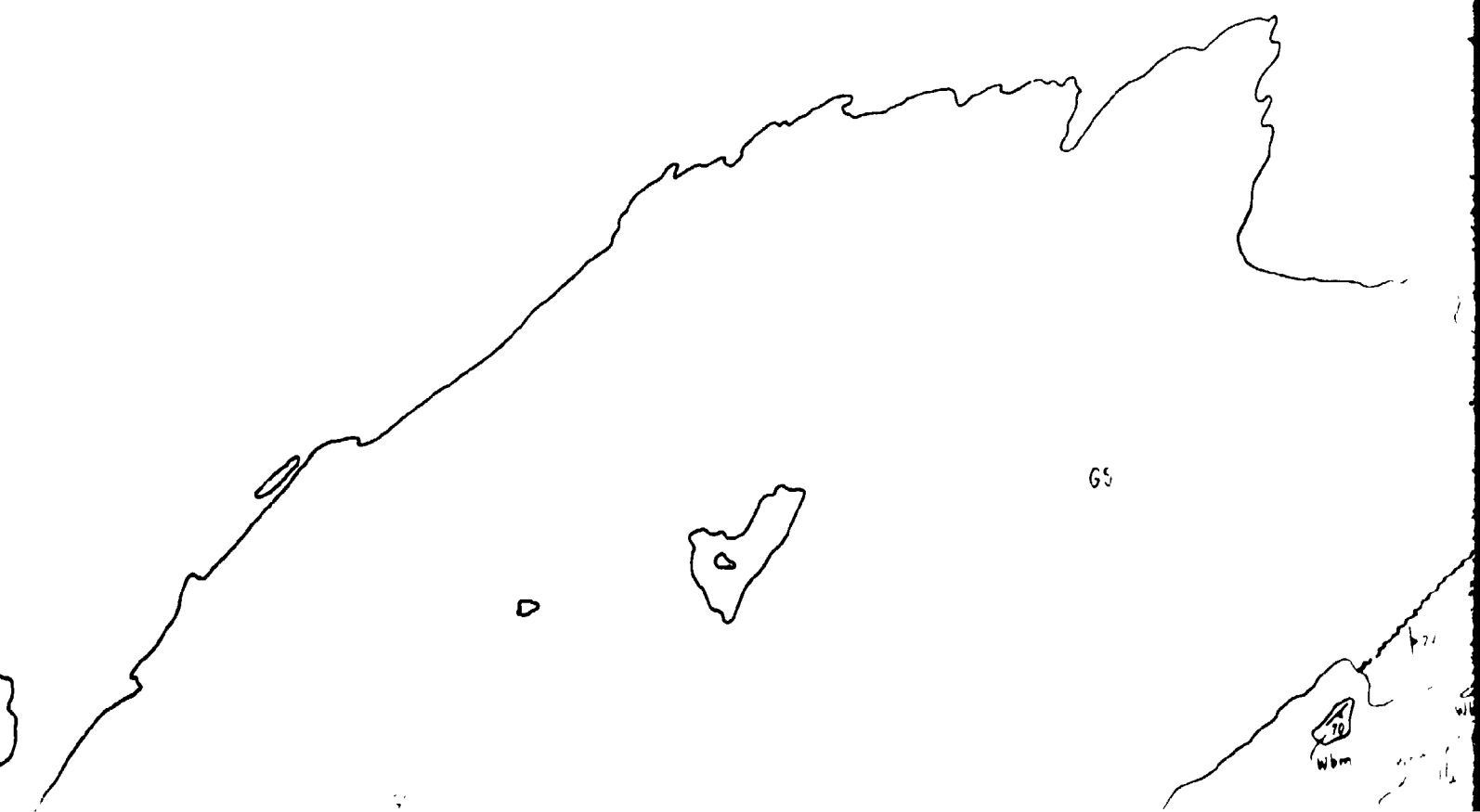
Fault (defined, approximate, assumed)

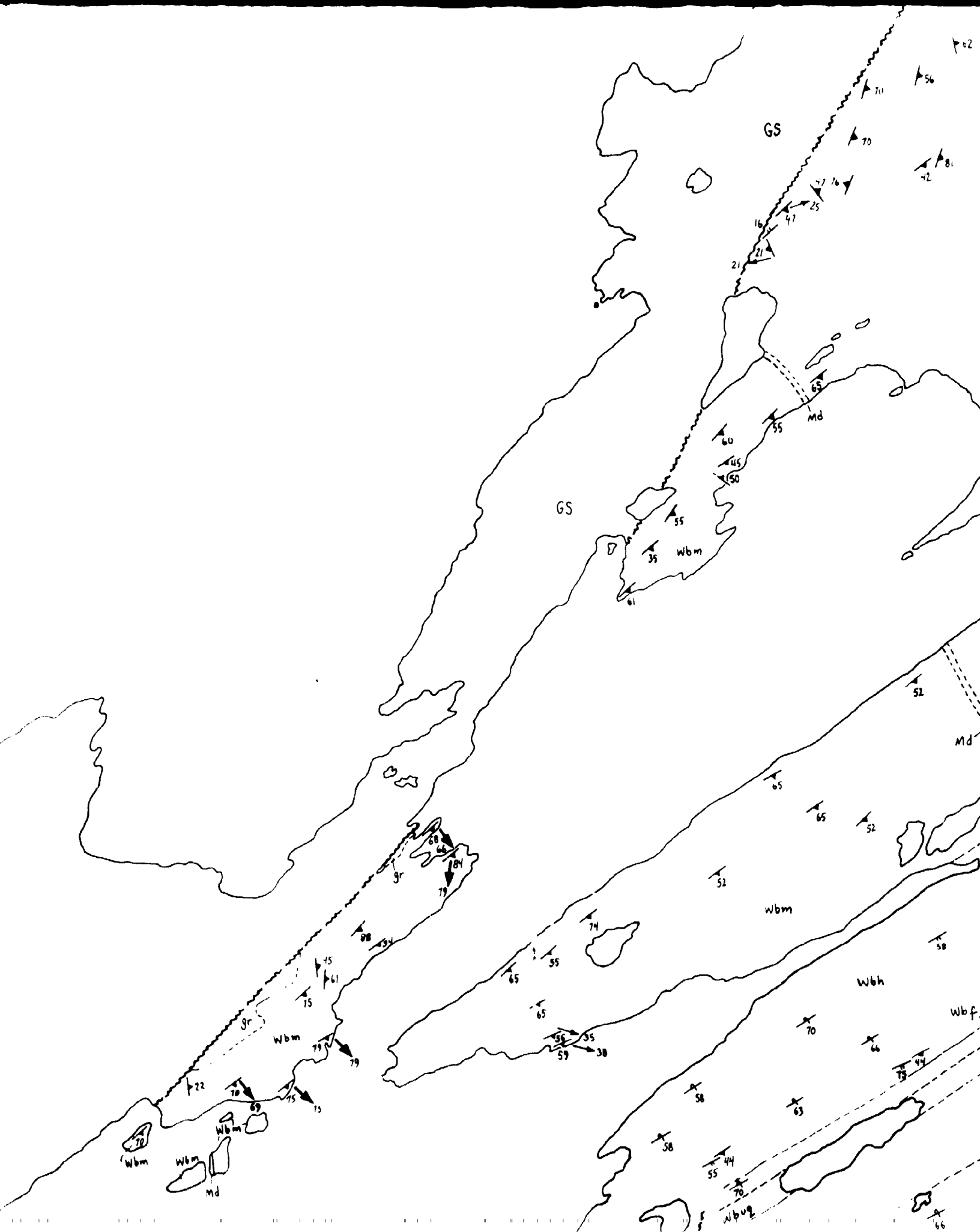
G R E

E A T

S L A V E

L A K E




GS
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Wbm
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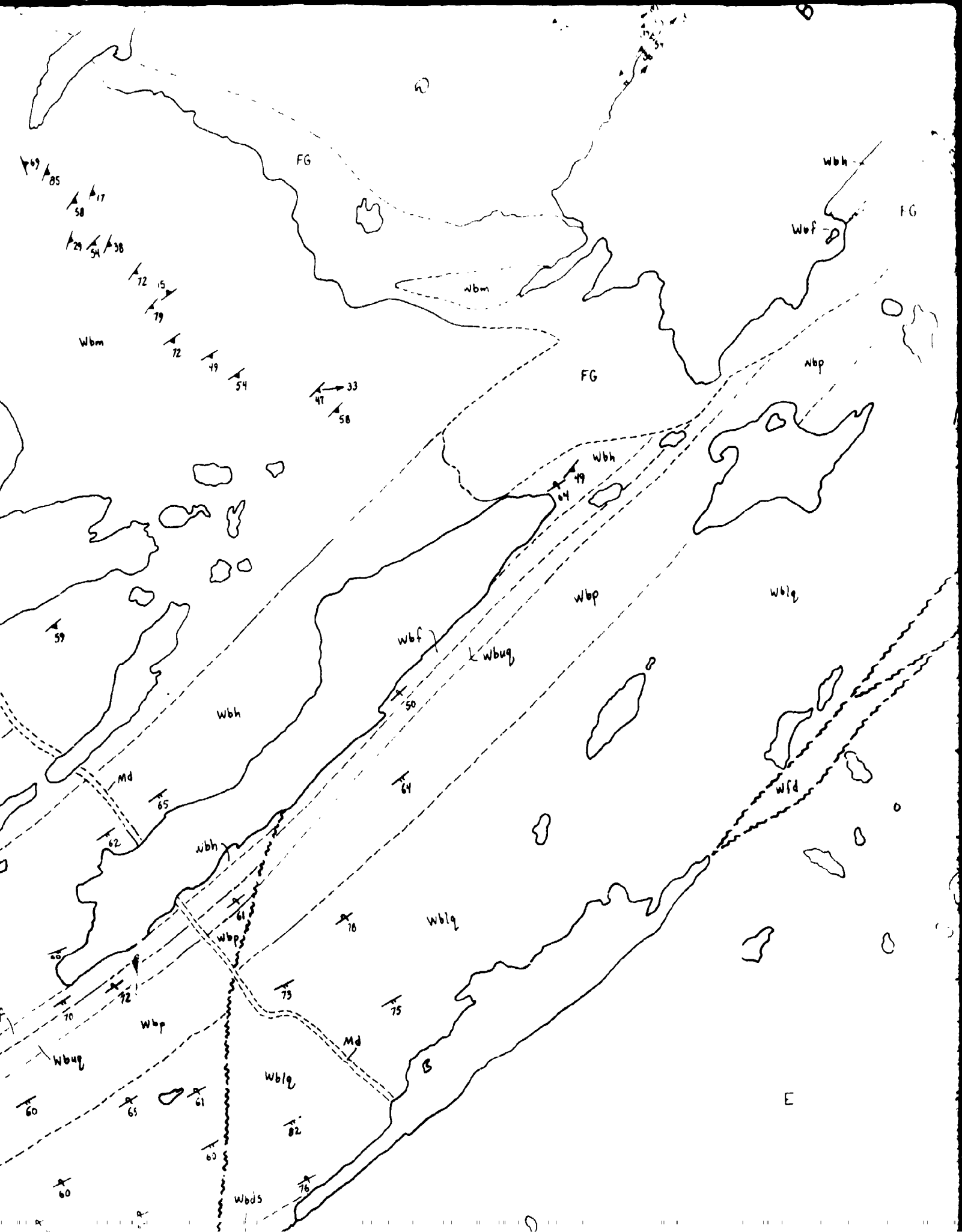
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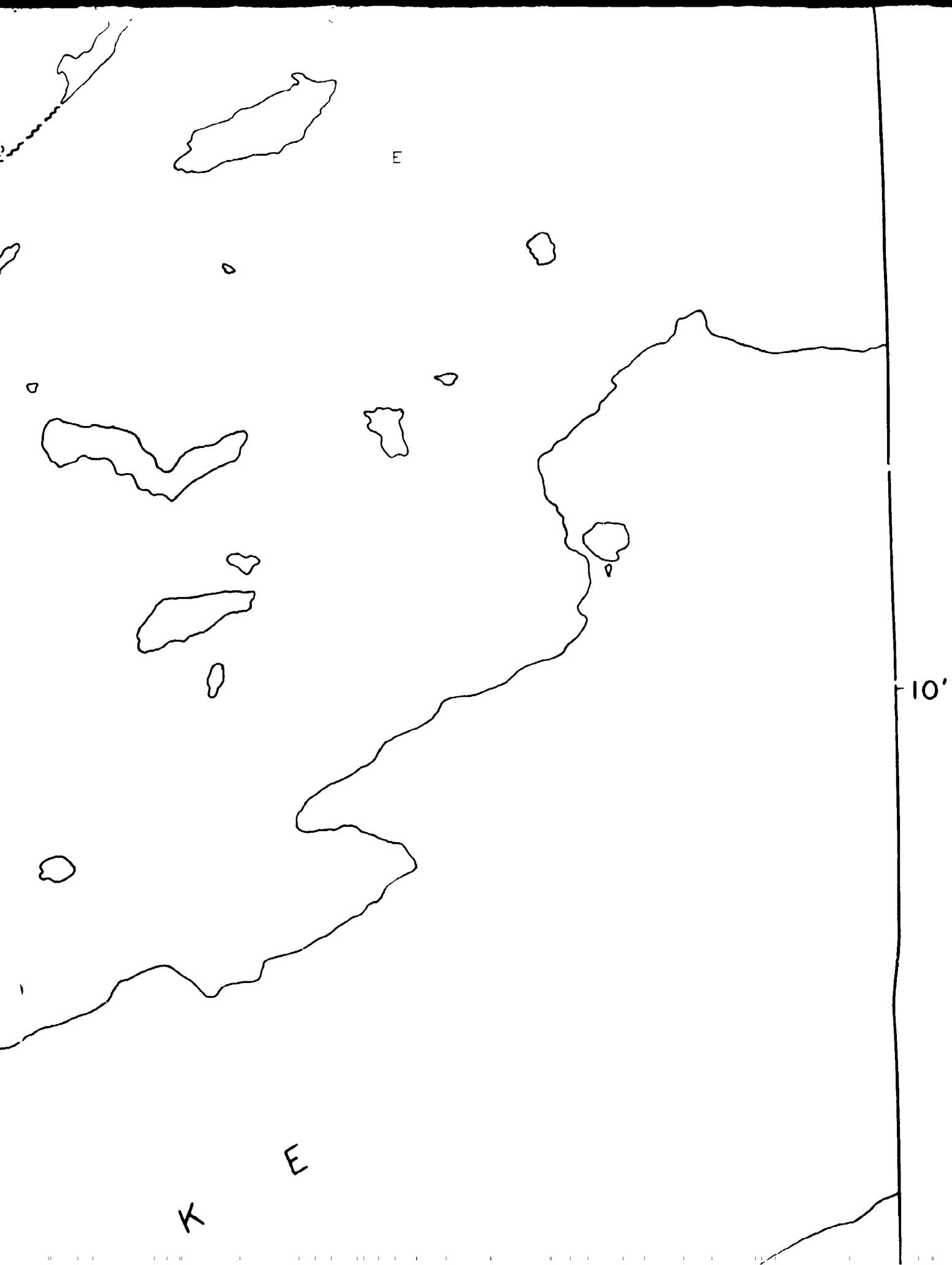
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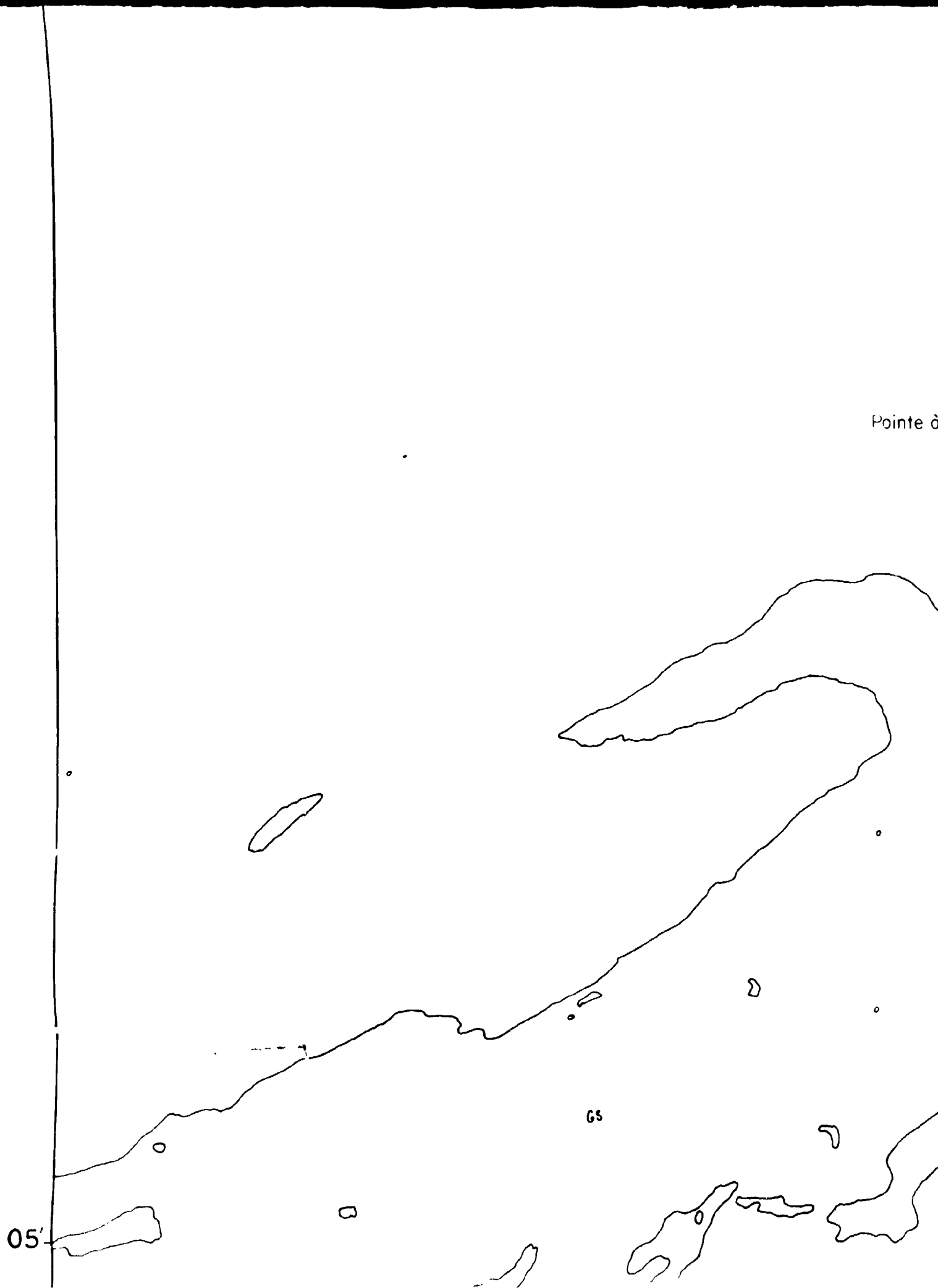
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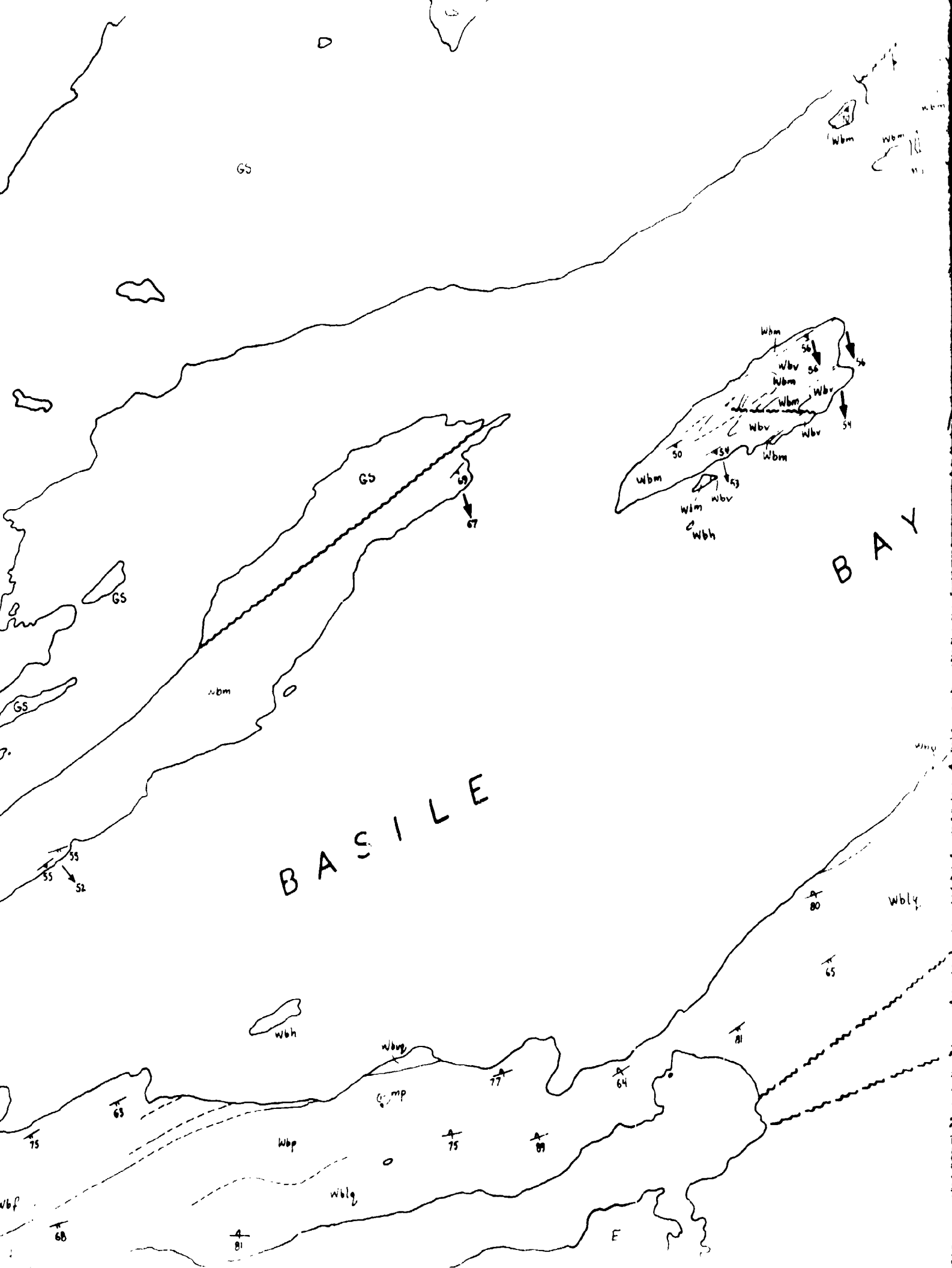
BASILE
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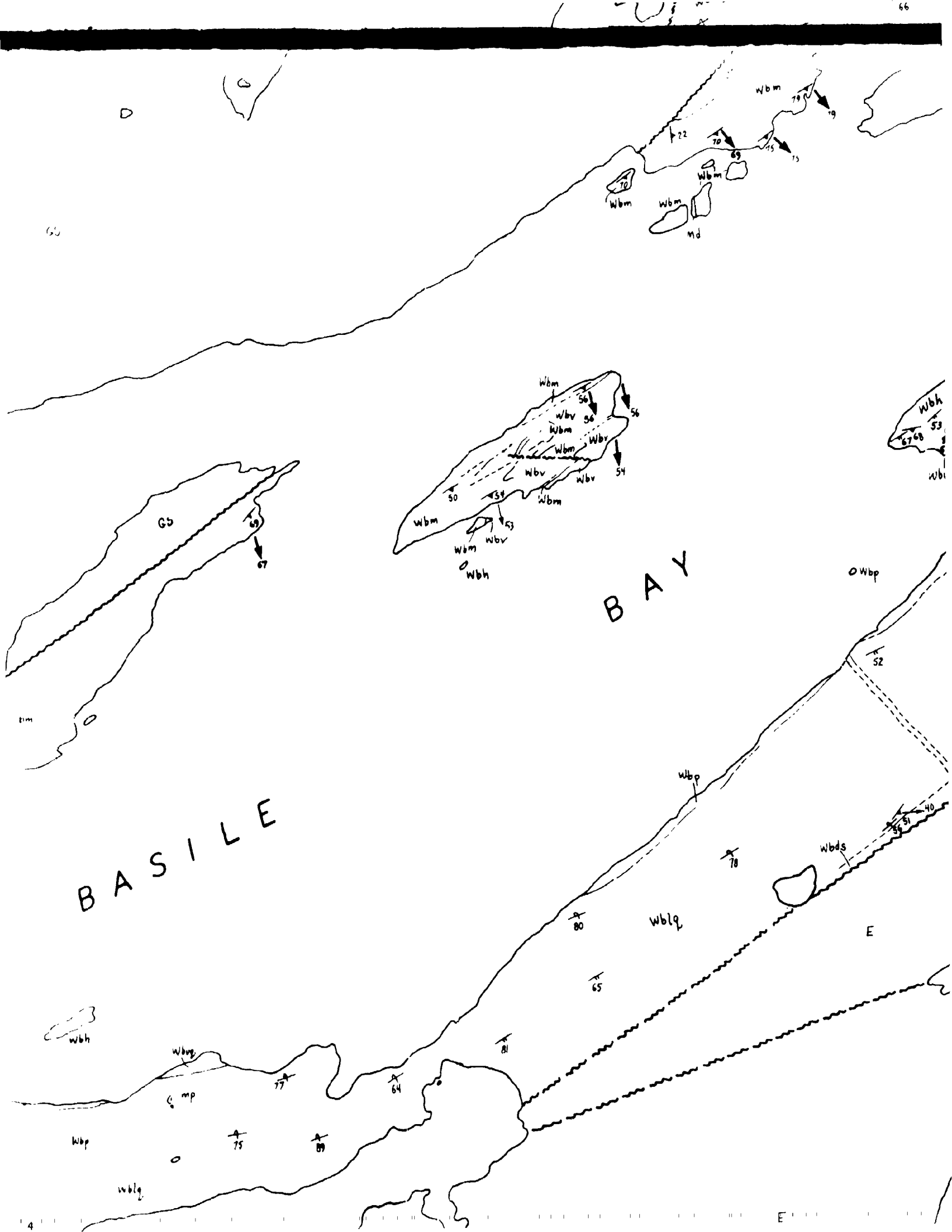
BASILE
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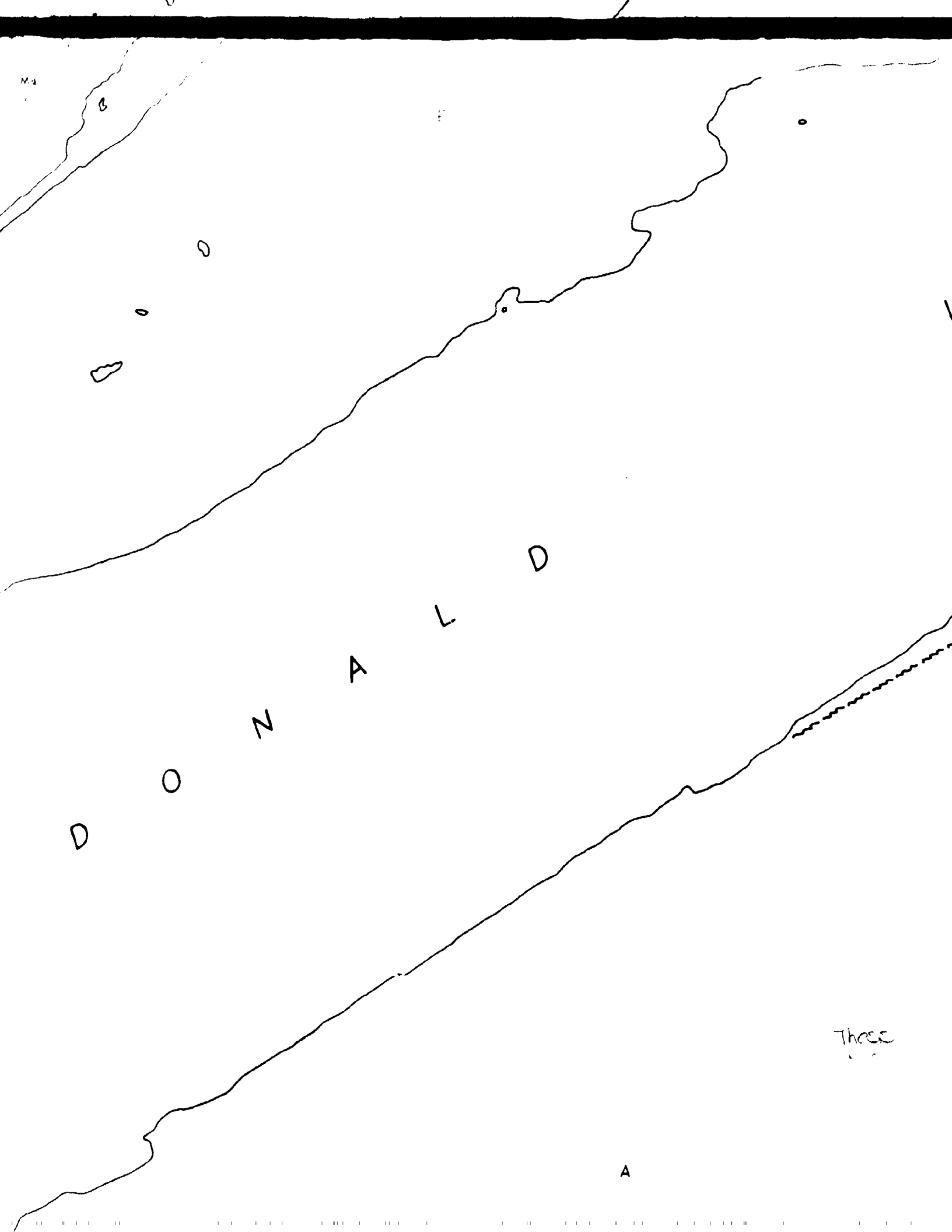
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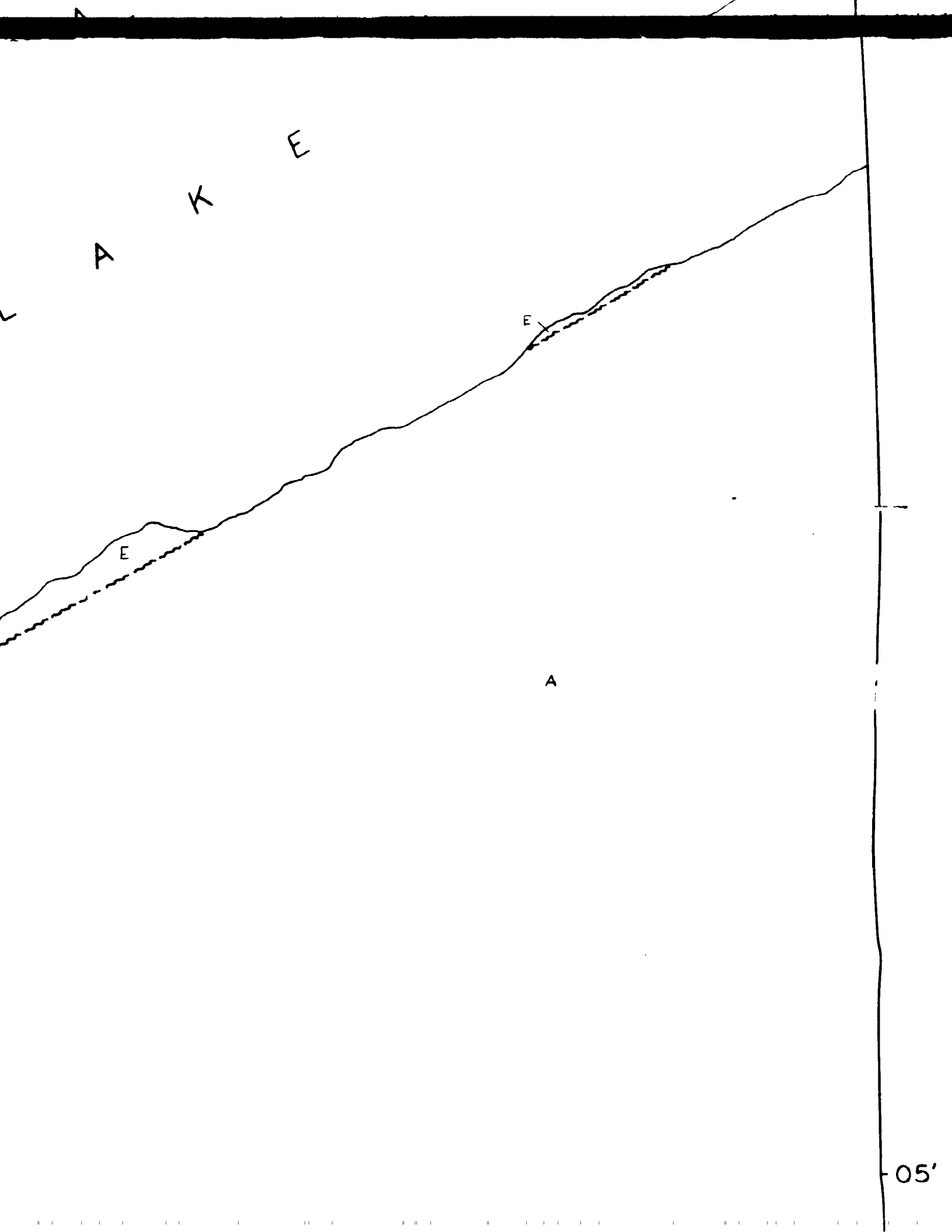
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05'


05'
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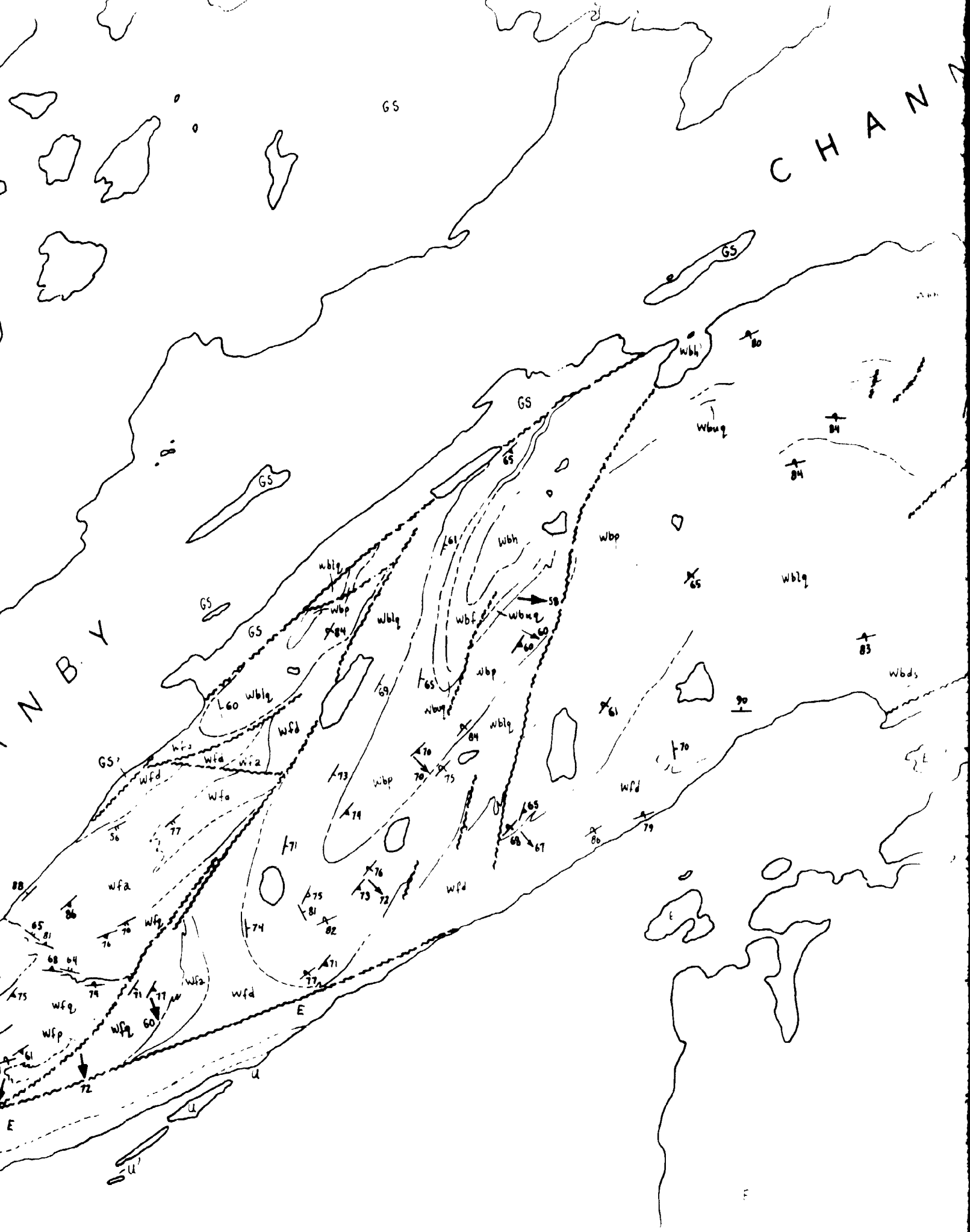
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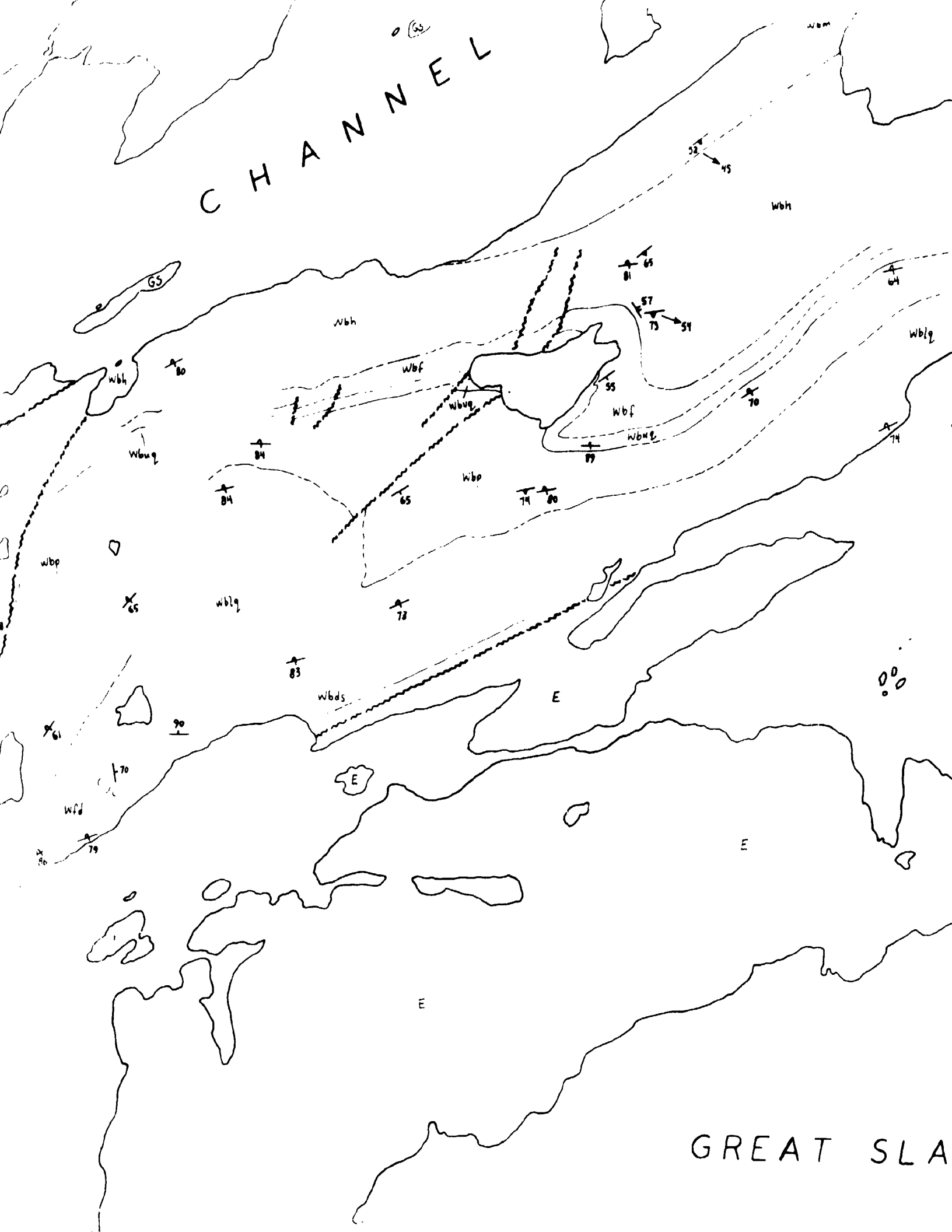
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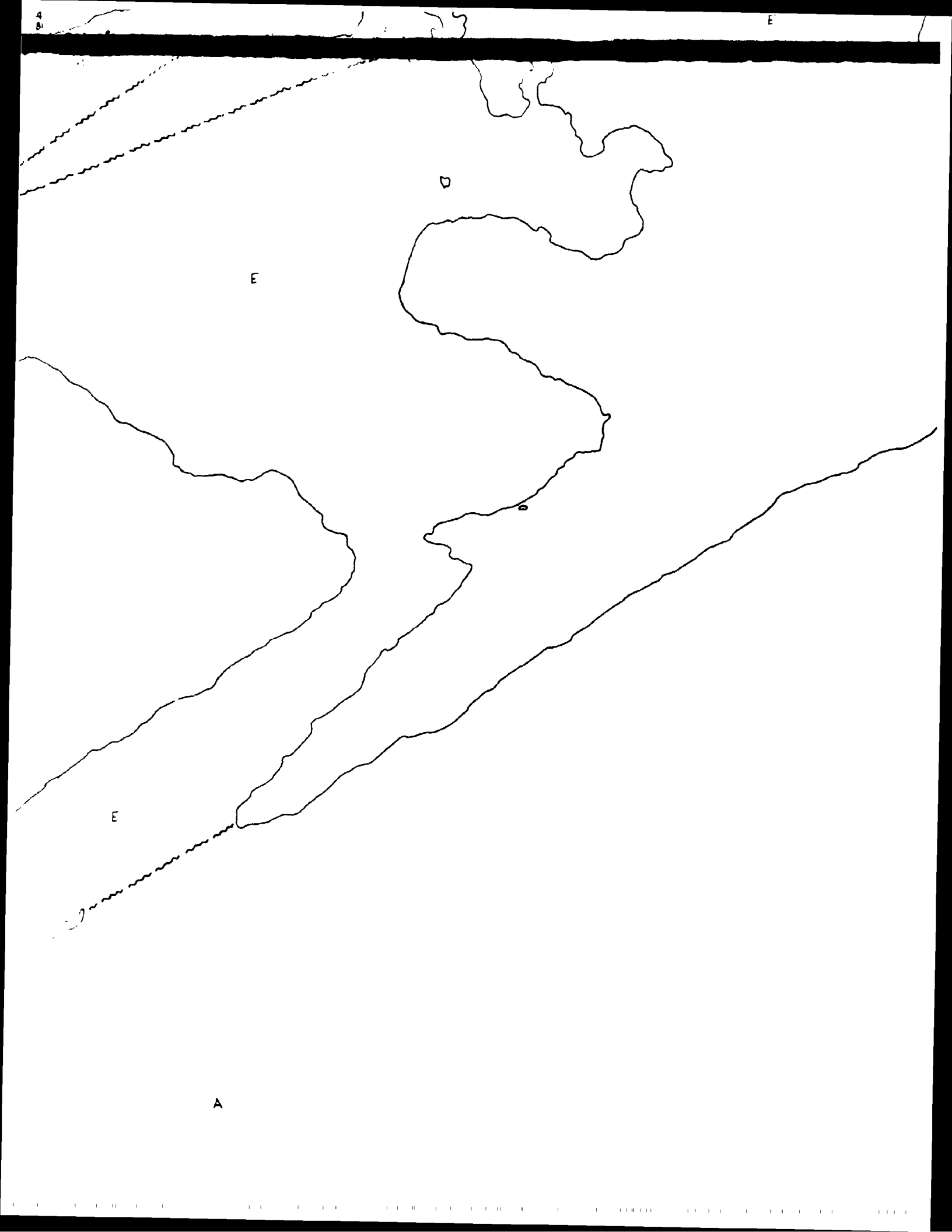
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MAP 2A

GEOLOGICAL MAP OF THE W
BASILE BAY
GREAT SLAVE LA

By B. J. JOHNSON

A

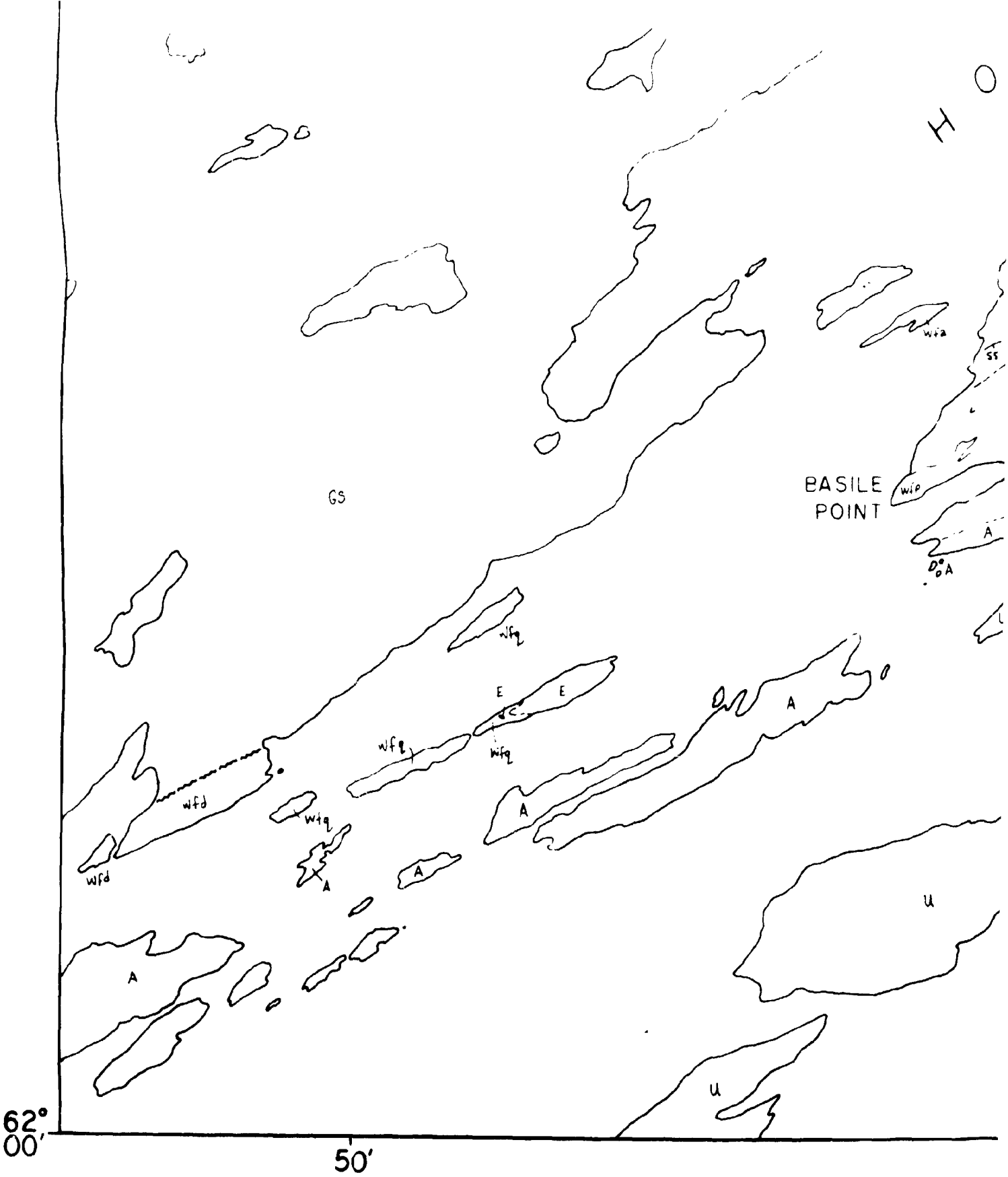
MAP 2A

GEOLOGICAL MAP OF THE WILSON ISLAND BASILE BAY AREA, GREAT SLAVE LAKE, N.W.T.

By B. J. JOHNSON, 1987

05'

D GROUP,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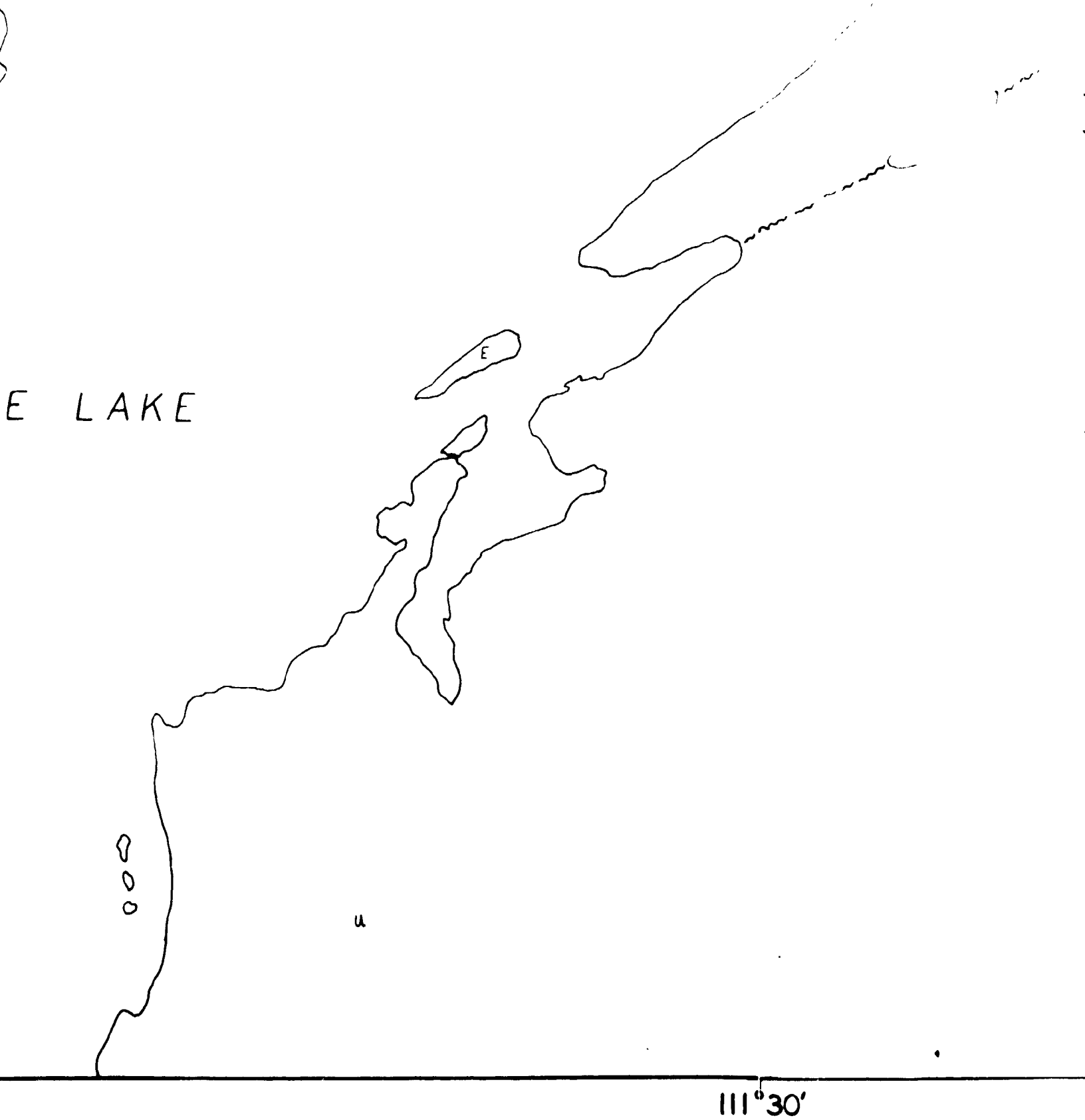
U
U
62°
00'
50'

45′ 40′


GREAT SLA
E
40'
35'


E LAKE
E
111°30'

A

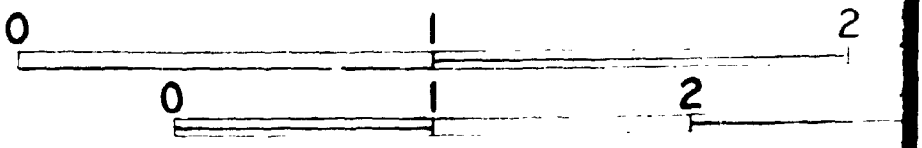
25'

GEOLOGICAL MAP OF THE
BASILE B
GREAT SLAVE

By B. J. JOHN

Geology mapped by B. J. Johnson (1984, 1985), A. Enge

Geology apart from Wilson Island



SCALE

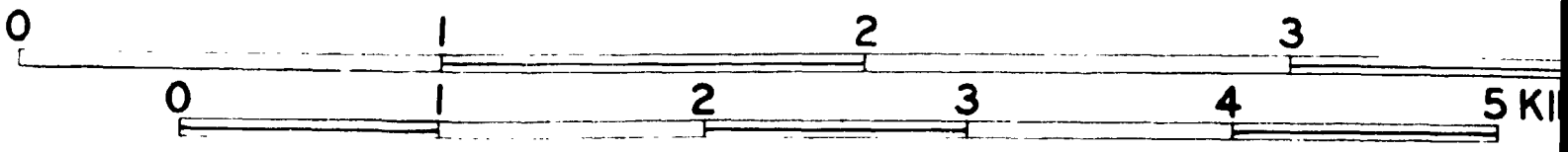
20'

GEOLOGICAL MAP OF THE WILSON ISLAND BASILE BAY AREA, GREAT SLAVE LAKE, N.W.T.

By B. J. JOHNSON, 1987

Geology mapped by B. J. Johnson (1984, 1985), A. Enge (1984), C. D. Goult (1985), and J. M. Wahlroth (1985)

Geology apart from Wilson Island Group in part simplified after Fettes (1977)



SCALE 1:25,000

20′

15′

GROUP,

4 MILES

LOMETRES

62°
00′

10′

END

18·03·90

FIN