THE UPLAND SURFACES OF WESTERN NEWFOUNDLAND

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ABSTRACT

In Western Newfoundland, erosion surfaces are recognised at the following elevations: 2350-2500', 2100-2300', 1900-2150', 1750-1900', 1550-1700', 1250-1500', 1100-1350', 1000-1150', 850-1000' and 750-850'. Surfaces below 750' were not investigated. All are considered to be of subaerial origin. By inductive reasoning, based on a recent hypothesis of the evolution of the continental margin of eastern North America, it is tentatively concluded that the group of surfaces above 2000' developed during a Cretaceous - early Cenozoic erosion cycle, and that lower surfaces are of later Cenozoic age. Uplift occurred at the beginning of the Cretaceous, in the mid-Cenozoic and in the late Cenozoic. Small vertical intervals between the surfaces are tentatively attributed to periodic positive isostatic responses of the earth's crust to the relief of load by denudation since at least the beginning of the Cretaceous.

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ERRATUM

Page 26, line 4: for "north" read "south".

Introduction

The level skylines of the Maritime Provinces and Newfoundland are one of the most striking features of the regional physiography of eastern Canada. The early geologists were much impressed by this characteristic but, due to the infancy of physiographic science, were unable to explain it. With the birth of the concepts propounded by W.M. Davis (1889,1899) at the turn of the century, the science became firmer established. From the Appalachian Mountains of Pennsylvania and neighbouring states, the study of landscape from a Davisian viewpoint spread farther afield, particularly into New England and into the Canadian Atlantic Provinces. In this latter area, work by Daly (1901), Goldthwait (1913,1924), Alcock (1926, 1928, 1935) and Twenhofel (1912, 1940) contributed much to an understanding of landscape evolution. Basing their hypotheses upon the Davisian concept of the development of surfaces of subaerial denudation, these workers attempted to establish a sequence of evolutionary stages.

This early work constitutes the bulk of studies of the broader aspects of the physiography of the Atlantic Provinces. Later, more detailed work has been undertaken to study the effects of glaciation in the region, but there has been no reconsideration of the earlier work on upland surfaces and drainage evolution. This work was done without the aid of reliable topographic maps and often without the relevant geological information. Workers such as Alcock and Twenhofel actually put forward their hypotheses of landscape development from field evidence gained during geological surveys.

Concerning the broad physiography of the Atlantic Provinces, Goldthwait (1924) wrote:

> "Overspreading the southern half of Nova Scotia, and appearing in many other parts of the Maritime Provinces, is an upland surface of striking uniformity. Although it appears only in detached fragments, ... this surface is believed to have been continuous, originally, throughout the region; for the several upland districts show a wonderfully systematic relationship of altitude. By matching the several parts it is possible to trace the ascent of the upland as a single feature, which may be called the "Atlantic Upland", from the coast at Yarmouth, Halifax, and Cape Canso, where it rises eastwards to the tablelands of Cape Breton and Newfoundland and Labrador. In some parts of this large area it forms undulating upland plains; in others, the flat crest-lines of mountains; and in Cape Breton and Newfoundland high tablelands of immense extent. Because this surface or facet cuts straight across the complex rock structure of the region, a structure which shows that it was occupied by high mountains in earlier geological ages, it is believed that the upland originated as a "plain of denudation" formed by the wearing down of the mountains to a lowland". (p.4).

The position of the island of Newfoundland within the geologic, tectonic and physiographic frameworks of eastern Canada (Fig.1) raised the question of whether or not the results of work in the Maritime Provinces can be corroborated and correlated in the former area. Goldthwait's observations in Nova Scotia led him to correlate the upland plain of that province with the high tablelands of Newfoundland as parts of the "Atlantic Upland". The work of Twenhofel (1912, 1940) in Newfoundland showed the same general picture as that outlined by Goldthwait, but the former work, while notable for its coverage of areas accessible only with difficulty, was necessarily of a reconnaissance nature and did not have the correlation of erosion surfaces as its main objective.

Western Newfoundland was chosen for this study both for the reason that it is relatively easy of access and because, being the highest

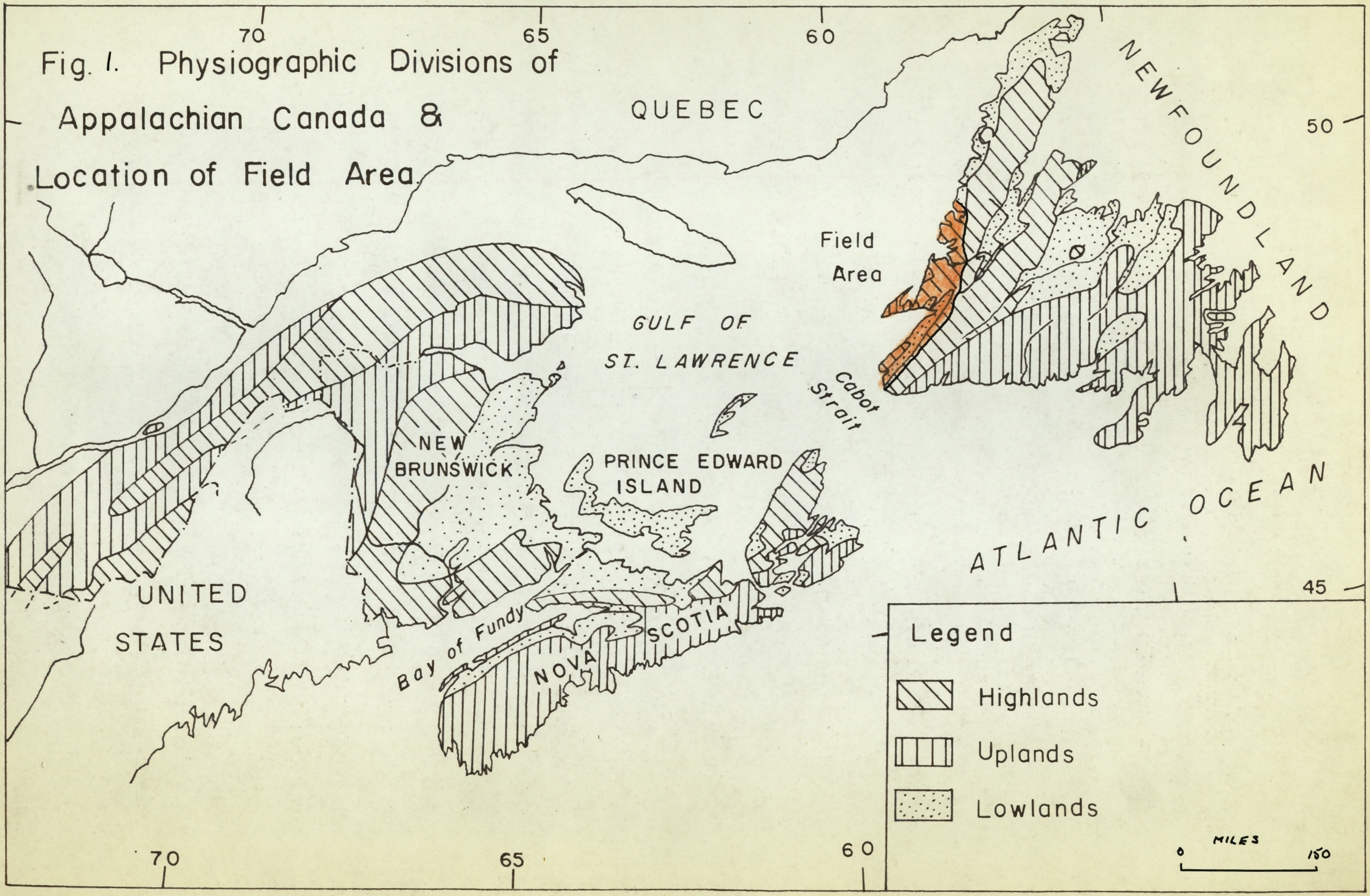

Fig. I. Physiographic Divisions of Appalachian Canada & Location of Field Area
QUEBEC
NEWFOUNDLAND
Field Area
GULF OF ST. LAWRENCE
Cabot Strait
NEW BRUNSWICK
PRINCE EDWARD ISLAND
ATLANTIC OCEAN
UNITED STATES
Bay of Fundy
NOVA SCOTIA
Legend
Highlands
Uplands
Lowlands
MILES
0
150
70
65
60
50
45

part of the island, it is there that a full sequence of erosion surfaces would be expected.

PART ONE

Chapter One: The Physiographic Regions of Western Newfoundland

The general impression of the surface of western Newfoundland is one of horizontality, with sweeping views of distant skylines. From their southern edge, overlooking Cabot Strait, the Long Range Mountains (sub-region Ia, end plate one) extend for 120 miles northeastwards to the vicinity of Grand Lake, and then continue into the Long Range Mountains of the Northern Peninsula as sub-region Ib. These mountains (Ia) present a steep front, ranging from 800-1300' in height, along their faulted western margin. The surface is extremely level when seen from the higher summits, a general level of 1600-1800' being maintained along their whole length. Residual smooth-topped hills rise 200-400' above this surface, which extends eastwards for several miles, and gradually declines in altitude to 1400-1200' in the valleys of large rivers, such as the Lloyds and Victoria.

In the southwest of sub-region Ia, the Long Range is separated from the Cape Anguille Mountains to the west by a steep-walled depression (sub-region II), averaging three miles in width, carved by the North and South Branches of the Grand Codroy River and Little Codroy River. The floor of this depression rises from sea-level, at the mouths of these rivers, to 500' northeast of the confluence of North and South Branches of the Codroy. From there, the land rises gently to the steep southern edge of an upstanding area with a level summit at 1600-1700', which closes the north-eastern end of the Codroy Depression. Two narrow passes, one on either side of this upland mass, lead to the northeast, towards St. George's Bay. The

Codroy Depression is bounded on the west by steep slopes rising up to the level surface of the Cape Anguille Mountains at a general elevation of 1500'.

The Cape Anguille Mountains (Region III) comprise an extensive dissected plateau, thirty miles long and up to eight miles wide, aligned southwest to northeast in the extreme southwest of Newfoundland. Steep cliffs drop to the Gulf of St. Lawrence along their western edge, which rises sharply inland to a level surface at about 1500'. Isolated summits rise above this level up to 1750', particularly along the eastern edge of the plateau. Broad upland valleys lie 100-200' below the general level, However, the lower and middle reaches of the rivers are incised up to 500' below the plateau top and the gentle upper valley-side slopes.

Northeast of these mountains, between them and the drift-covered lowland around the head of St. George's Bay, and between the western shore of this bay and the Long Range front, the land rises gradually from coastal cliffs, 100' high, to the foot of the Long Range scarp, in a distance of six to twelve miles. This sloping area is here referred to as the Long Range Foreland (sub-region IV). Dissecting this evenly-sloping foreland is a system of rivers which rise east of the Long Range high summits on the level plateau, and then flow west across the Long Range in broad, shallow valleys, which abruptly become incised six to eight miles east of the mountain front. They cross the front, marked along most of its length by a steep scarp, in steep-sided canyons, and continue their courses to St. George's Bay as a system of sub-parallel rivers which become increasingly shallow and wide towards the sea.

Between Humber Arm, Bay of Islands, and St. George's Bay, east of the mountain areas of the Lewis Hills and Blow me Down Mountains, and including the Port au Port Peninsula, the land surface consists of southwest to northeast-trending ridges and valleys, with an overall subdued relief, at a general elevation of 1100-1200'. The alignment of ridges and valleys is in general conformity with the folded and faulted structures in the Lower Palaeozoic sedimentary rocks. Interrupting the physiographic continuity of this ridge-and-valley region (sub-region VIa) is the Indian Head Range (Region VII), a line of hills, some 15 miles long, trending south to north and rising to 1800'. The physiographic character of the sub-region VIa is continued, with slight differences, in the area between Humber Arm and Bonne Bay (sub-region VIb).

West of these two sub-regions are the high, barren, steep-sided Bay of Islands Mountains (region VIII). These mountains, rising to 2000-2600', comprise the Lewis Hills (sub-region a), the Blow me Down Mountains (b), the St. Gregory Highlands (c), North Arm Mountain (d) and Table Mountain, Bonne Bay (e). They have an overall physiographic uniformity which reflects geologic homogeneity, but minor differences exist between them as a result of lithological and structural variation.

The Long Range Mountains of the Northern Peninsula (sub-region Ib) are distinct from those mountains farther east and southeast by virtue of their greater height, since they rise to 2600', and of the fact that the major drainage lines flow east from a divide located along their western edge whereas, in sub-region Ia, the drainage is to the west from a divide to the east of the higher summits. The coastal plain (Region IX) varies from two to eight miles in width, and rises gently from the sea to about 600' at the foot of the scarp along the western front of the Long Range Mountains (Ib). There is a longitudinally ribbed relief which is an expression of the differential erosion of steeply-dipping Lower Palaeozoic strata.

Chapter Two: Previous Investigations

Previous work on the erosion surfaces of Newfoundland is limited to the investigations of W.H. Twenhofel. The earlier of his two works, published in 1912, was based on very few observations made during geological surveys of the western part of the island. The later (1940) work was an elaboration and revision of the earlier.

Twenhofel, in his 1912 work, recognized that, over the whole of Newfoundland, "decrease in the elevations of the highlands southeastwards is fairly systematic, averaging a little less than ten feet per mile". He considered the tablelands as evidence "of the present dissection, but one-time perfection of a peneplain, a plain of erosion of remarkable perfection extending over the whole of Newfoundland". This peneplain declined from 2000' in the west to 700' in the east, a slope which Twenhofel ascribed to tilting during uplift. This uplift took the form of independent structural blocks, rather than a single warping. He tentatively correlated this upland peneplain with the Schooley Peneplain of the eastern United States, developed before the end of the Cretaceous period. Twenhofel also recognized another erosion cycle, initiated by uplift of about 800' at the close of the Cretaceous, during which the upland valleys developed to a stage of maturity.

The later work of Twenhofel (1940), undertaken over a longer period, was more directly concerned with physiographic problems. This study resulted in the recognition of three peneplains, in place of the original one. The highest mountains in the Long Range were seen as remnants of a surface at above 2000', which was named the Long Range Peneplain. This was represented east of the Long Range by isolated accordant summits at about the same elevation. A second peneplain, the High Valley Peneplain, was recognized as high-level valleys and dissected plateau on the Long Range at between 1300-1700'. This surface continued eastwards, first as the dominant erosion surface of the western parts of the Central Plateau, at 1200-1600', and farther east as accordant monadnocks and isolated plateau areas at between 1000-1200'. This surface sloped finally to the higher summits at 700-800' above the plateau of low relief on the Avalon Peninsula. The third, Lawrence, peneplain occurred in the west as a dissected upland surface at 1000-1100', with intermediate mature valleys at 500-1000'. In the Central Plateau this peneplain was found at the same heights in the wide valleys of the major rivers, at 200' around the head of the Bay of Exploits, and at 400' around Gander and on the Avalon Peninsula. The schematic profile across Newfoundland from Bonne Bay on the west to the Avalon Peninsula on the east, shown in figure 2.1., shows the heights at which these three peneplains were found. This information is tabulated for the major physiographic regions of Newfoundland in Table 1., p. 10.

TABLE 1

CORRELATION OF PENEPLAINS ACROSS NEWFOUNDLAND

Peneplain	Western Mountains	Central Plateau W.	Central Plateau E.	Avalon Peninsula
Long Range	2000'- plus	2000'	-	-
High Valley	1300-1700'	1200-1600'.	1000-1200'.	700-800'.
Lawrence	1000-1100'	500-1000'	400'	400'

Calculations made from the figures given by Twenhofel (1940) for the upper and lower limits of the peneplains, and the horizontal distances over which they are found, give the following gradients for the peneplains:

Peneplain	Gradient in feet per mile
Long Range	maximum - 5.5
	minimum - 3.5
High Valley	maximum - 3.0
	minimum - 1.0
Lawrence	maximum - 1.0
	minimum - 0.5

These gradients are well within the limits normally quoted for well-developed, undeformed peneplains. However, the fact that these gradients show successive reductions in amount, suggests that the diastrophism which initiated successively younger cycles of erosion was differential, and that it tilted the pre-existing erosion surfaces to the east. Thus, according to the above calculations, the Long Range Peneplain now has a slope which is the resultant of its original slope and three later tilting movements; the High Valley

Peneplain gradient reflects its original slope and two subsequent movements and that of the Lawrence Peneplain reflects its original slope and one final movement. Another possible explanation of the decrease in gradients is that each successively younger peneplain was better developed than the preceeding one and, therefore, possessed a gentler slope. This is not borne out by Twenhofel's conclusions, and the writer believes it is unlikely considering the known facts of landscape development in similar areas.

Twenhofel (1940) remarks that the present surface of Newfoundland seems to indicate that "the latest regional uplift was greater on the west coast than on the east, thus tilting the original surface to the southeast" (p. 1673). With no reference to a previous easterly tilt, this implies that only one period of diastrophism has been sufficient to effect the tilt of the peneplains. If this was the case, the gradients of the peneplains would be more nearly comparable, rather than showing a regular decrease.

Apart from the problem of the nature and the number of the earth movements affecting the surface of Newfoundland, there is the problem of the anomaly of the proposed peneplain slopes and the direction and slopes of the drainage lines. While the peneplains recognized by Twenhofel are said to have an easterly, or southeasterly slope, it is obvious even from a cursory examination of the drainage pattern, that this bears little relation to such a slope. Twenhofel himself states that streams must always have tended to flow parallel to structural trends which, in the main, run in most pronounced fashion southwest to northeast, at right angles to the proposed peneplain slopes.

In attempting to explain the easterly or southeasterly slope of the peneplains by earth movements of greater magnitude on the west than on the east, Twenhofel encountered the difficulty of accounting for the westerly courses of the rivers flowing into St. George's Bay. The peculiar course of the Humber River was ascribed to antecedence in both the early and later works. However, little attention was paid to the St. George's Bay river system. It remains to be seen if the results of this study can shed light on the evolution of that drainage.

It is clear that a number of problems remain to be solved concerning the evolution of the Newfoundland landscape. While the work of Twenhofel was of considerable value, it is now possible with the aid of topographic maps and improved conditions of accessibility, to furnish more evidence with which to reconsider these problems.

Chapter Three: Geological Outline

In a study of this kind, dealing with erosion surfaces, it is of prime importance to ascertain the extent to which the surface of the land reflects the lithology and structure of the underlying rocks. A major criterion for the recognition of erosional land surfaces, as against structural surfaces, is the former's lack of conformity with the geology and structure. General conformity of the form of a landscape to the gross lithology and structure of the underlying rocks can be expected as a result of the progressive adjustment of drainage to structure through two or more erosion cycles interrupted by periods of uplift. However, it is when, for example, the folds and faults within one formation, different in its gross lithology and structure from its neighbour, are truncated by the land surface that the existence of erosional landsurfaces can be suspected. Conversely, when morphological discontinuities unrelated to any geological discontinuity are recognized similar suspicions are aroused.

It is with a view towards examining the lithology and structure of the rocks of western Newfoundland that this geological outline is presented. When the land surface itself comes under consideration later, the relationship between it and the rocks will be understood more fully. The presentation is historical, an approach well suited to this kind of study since, when the periods covering the formation of the rocks and structures of the area are described, the scene is set for a discussion of events represented in the land surface itself. A simplified geological map of western Newfoundland is shown in end plate 2.

According to Eardley (1962), Newfoundland lay within the Grenville Province of the Canadian Shield during later Precambrian time. Clifford and Baird (1962) recognize the Great Northern Peninsula as a Grenville inlier and give a minimum age for emplacement of the granites after a major orogenic period as 945 million years. They also state that, by the end of the Precambrian, the main outlines of the peninsula were established and it was not subsequently affected by orogenic movements, although epeirogenesis has undoubtedly occurred.

At the opening of the Palaeozoic era, two geosynclines had developed across Newfoundland (Schuchert and Dunbar, 1934). One, the St. Lawrence miogeosyncline, was a continuation of that simultaneously occupying the New England - Gaspe region and the other, the Acadian eugeosyncline, continued across eastern Newfoundland that occupying most of Nova Scotia. These were separated from each other by the broad swell of the New Brunswick geanticline (Schuchert and Dunbar, 1934) of Precambrian rocks, whose extent in central Newfoundland has been limited by subsequent crustal shortening. In western Newfoundland Cambrian rocks are found in three main areas. (See map --end plate 2). They occur fringing the Precambrian massif of the Northern Peninsula to the southwest, fringing the Lower Palaeozoic granites and granite-gneisses of the Long Range west of Grand Lake and in the area north of St. George's Bay. In the first area, Cambrian strata comprise fossiliferous shale, slaty shale and quartzite, with bands of calcareous rocks in the upper horizons. Farther to the southeast, around Big Bonne Bay Pond, the rocks become slightly metamorphosed to mica schist and slaty shale, and the grade of metamorphism increases southeastwards until, near

Deer Lake, dark schists and gneisses are tentatively classified with the Cambrian (Baird, 1959). In the area around the head of St. George's Bay, Riley (1962) describes the Cambrian strata as follows:-

Lower Cambrian -	Kippens formation: grey and black shales and interbedded grey limestone bands.
Middle Cambrian-	March Point formation: only separate on Port au Port peninsula. Elsewhere, the same as -
Upper Cambrian -	Petit Jardin formation: dolomites, shaly and silty beds and basal quartzose sandstone. Shales more common east of Stephenville.

There is no evidence for Cambrian orogenic movements, but in the early and late Cambrian slight movements of the sea floor took place during sedimentation. Metamorphosed Cambrian rocks, similar to those described southwest of the Precambrian area of the Northern Peninsula, occur at the western end of Grand Lake in unfaulted contact with "metamorphosed Palaeozoics" (Riley, 1962) of the Long Range, the grade of metamorphism increasing eastwards.

The Cambrian sedimentary sequence was continued unbroken into the Ordovician. From Eardley's (1962) maps, it seems that the New Brunswick geanticline had subsided in this period, allowing marine transgression across the island. Riley describes a well-developed sequence of Ordovician sediments on the west coast. They are predominantly calcareous with a maximum thickness of 16,700 feet. The Lower Ordovician <u>Green Point Group</u> has only a limited outcrop on the Port au Port peninsula. The next youngest

St. George and Table Head Groups occur both separately as sequences of limestone, dolomite, shale and undifferentiated near the boundaries with the ultramafic rocks of the Bay of Islands Igneous Complex and near the Long Range Igneous and Metamorphic Complex. The Humber Arm Group is considered to be Middle Ordovician in age and generally appears to lie above the St. George and Table Head groups. It is composed predominantly of clastic sediments, together with some limestone conglomerates, massive limestone, sandstone and conglomerate, quartzites, phyllites and argillites. The volcanic rocks comprise andesitic and basaltic flows, tuffs and agglomerates.

The problem of the relationship of the two Ordovician sequences, the carbonate and the clastic, has recently been attacked by Rodgers and Neale (1963). They recognize the difficulty of accounting for the widely differing lithologic characteristics and structural relations of two apparently contemporaneous Ordovician sequences, each many thousands of feet thick, one a carbonate sequence, represented by the St. George and Table Head groups, and the other, a clastic terrane associated with the ultramafic rocks of the Bay of Islands Igneous Complex, represented by the Humber Arm Group. Both sequences appear to range in age from early Cambrian to Middle Ordovician according to these authors but, structurally, the clastic terrane is surrounded on three sides by, and generally seems to lie above, the carbonate sequence. The latter rests unconformably on Precambrian rocks and it is identified as autochthonous. Rodgers and Neale state that if the clastic terrane is also autochthonous, then either invisible unconformities or rapid facies changes must separate the two. An hypothesis is forwarded which not only appears to fit the known facts in Newfoundland (although it must be admitted that there are many more that are unknown) but also makes it possible to recognize similarities in the geological histories of that area and the Taconic province of eastern New York and southern New England. The hypothesis is that the clastic terrane is allochthonous and forms two large Klippe (see Fig. 3.1), one extending along the west coast from Port au Port to Daniel's Harbour and the other at the tip of the Great Northern Peninsula between Canada and Pistolet Bays. These allochthonous masses are said to have originated east of the carbonate sequence in an area where large bodies of lower Palaeozoic clastic and volcanic rocks now occur as remnants in a dominantly granitic terrane. This last, broadly speaking, is the shallow northeast-trending trough east of the Annieopsquotch Mountains. From this area, their place of origin, they are said to have been first squeezed up and then slid under gravity westwards into a basin where shale was being deposited upon the autochthonous carbonate sequence. If such a movement did occur then it was most probably during the Taconic orogeny whose effects are so marked in the Appalchian mountains of New England and the northern parts of the Maritime Provinces. The precise effects of this orogeny in western Newfoundland have remained in doubt, but if this hypothesis is acceptable then, together with the evidence from the Bay of Islands Igneous Complex, these may become better known.

The Bay of Islands Igneous Complex comprises four discrete masses of gabbroic and serpentinised rocks of plutonic character. The principal

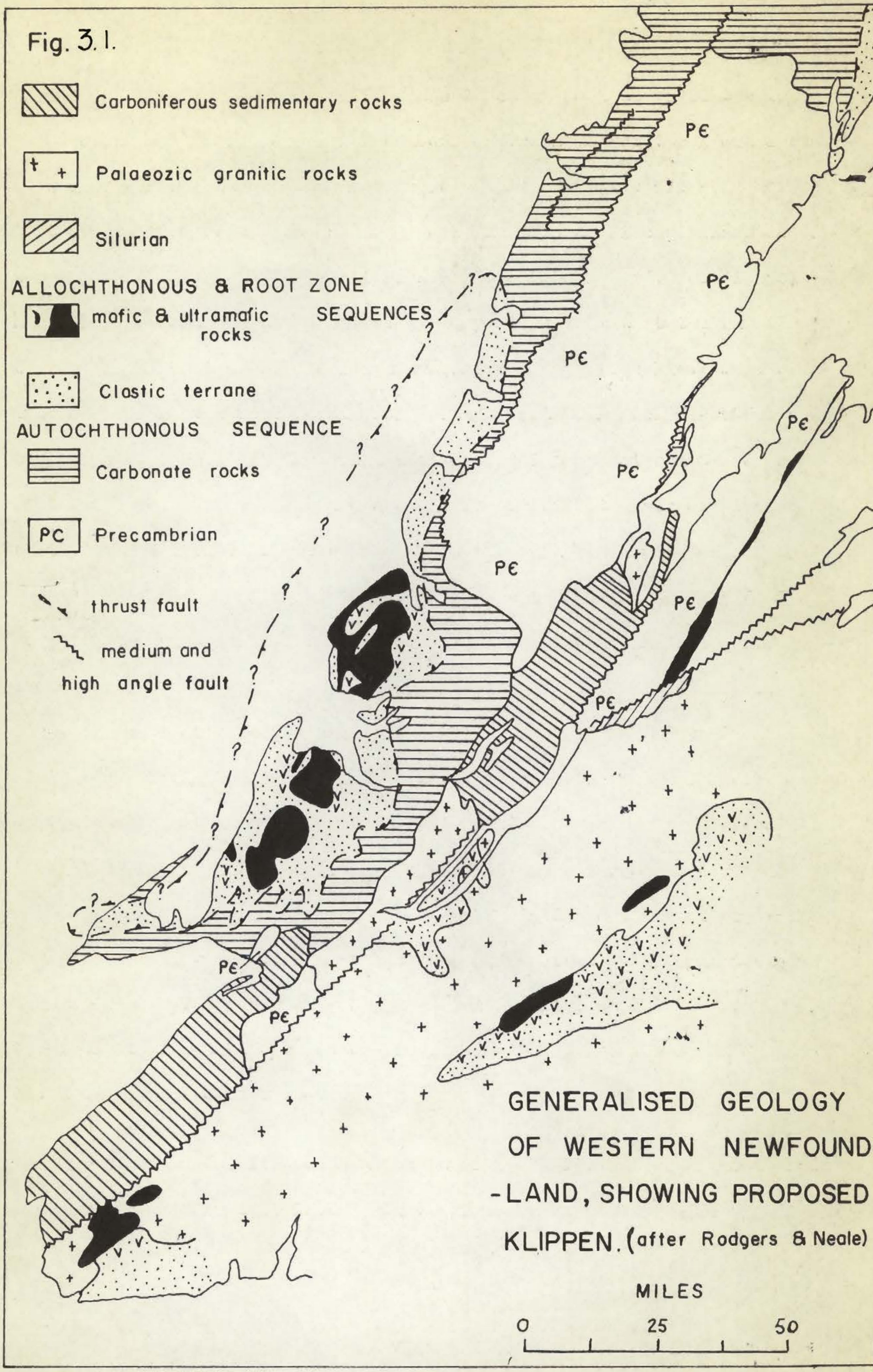

Fig. 3.1. GENERALISED GEOLOGY OF WESTERN NEWFOUND-LAND, SHOWING PROPOSED KLIPPEN. (after Rodgers & Neale)

masses are, from north to south, the Lewis Hills, the Blow me Down Mountains, Mount St. Gregory - North Arm Mountain mass and Table Mountain, Bonne Bay. The Complex, with its surrounding volcanic and sedimentary rocks, is only a part of the eugeosynclinal sequence exposed along the west coast. According to Smith (1958), the plutons show many of the characteristics of other classical gabbroic plutons like the Bushveld Complex, and yet are unique in that they combine features of typical gravity-stratified sheets with those of ultramafic plutons of orogenic belts, such as those of the northern Appalachians. Each pluton has a thick basal ultramafic zone, overlain by banded and massive gabbroic rocks. The major ultramafic plutons have westerly-dipping floors that are generally conformable with Ordovician sediments and volcanics. The latter are metamorphosed near their contacts with the plutons. Structural features within the plutons are well defined and generally conform to the northeasterly strike of major structures throughout western Newfoundland (fig. 81). Primary banding of ultramafic and gabbroic rocks and strata of the Humber Arm group have been folded along northeasterly axes, and are now offset by northwesterly-directed transverse faults, such as the Serpentine and Trout River faults. Probably, these masses were once part of a single body which has been disrupted and the separate parts moved into their present positions along high-angle reverse faults. They have acted as buttresses against which the sedimentary rocks to the east have later been folded (Smith, 1958; Walthier, 1949). The problem of accounting for the emplacement of the plutons is simplified in the light of the <u>Klippe</u> hypothesis of Rodgers and Neale (1963), that is, if these areas are also considered as allochthonous. Similar rocks to those of the Bay of Islands Igneous Complex are found in the Annieopsquotch Mountains, some fifty miles southeast, in association with Ordovician sedimentary and volcanic rocks, which are thought to represent remanent patches of clastic terrane left after the westward movement of the Klippe. Indeed, Smith (1958) states that "there is no indication that primary ultrabasic rocks were intruded into the Humber Arm rocks (the allochthonous sequence), although later deformation has brought them into contact in places", and that "there is evidence, ..., that the plutons were emplaced in an active tectonic environment in which their original form was modified considerably" (p. 51). Thus, it seems possible that the rocks of the Bay of Islands Igneous Complex could have taken part in the westward movement of the Klippe.

In western Newfoundland, Silurian rocks are known only in the White Bay area, which does not lie within the field area. Devonian sedimentary rocks are lacking, but radiogenic age-determinations of granites in the Long Range Mountains east and southeast of Grand Lake suggest a major orogenic episode during that period. This orogeny probably was the equivalent of the Acadian orogeny whose effects were so pronounced in Nova Scotia. The relative importance of Taconic and Acadian orogenies in Newfoundland is still disputed, but Neale and Nash (1963) favour deformation in both periods.

Knowledge of the Carboniferous rocks of western Newfoundland is largely due to the work of Hayes and Johnson (1938). The lithology and thickness of the various groups and formations can be seen in Table II. Structurally, the Mississippian Anguille Group occupies an anticline, plunging northeast and southwest, probably with subsidiary flexures superimposed upon it. Hayes and Johnson's work revealed the presence of faults on the margins of the anticline and it is likely that many more faults exist than have so far been mapped. A lesser anticlinal structure brings Anguille Group rocks to the surface between Barachois and Flat Bay brooks (see end plate 3). In its southeastern-most section this anticline is described by Bell (1948) as having a southwesterly plunge. The Anguille Group rocks are seen again faulted against the Long Range igneous and metamorphic rocks across Crabbes Brook, where the structure is domal.

The Upper Mississippian Codroy Group is conformable with older and younger Carboniferous rocks. Their relationships to Long Range rocks, the only others with which they come into contact, are twofold. Along the northeast-trending section of the Long Range fault they lie in faulted contact with them, as do other Carboniferous rocks. Fault-contact relations are also seen along the northward-trending fault bounding an anorthosite body to the west, across Flat Bay Brook. However, an unfaulted, unconformable contact is found along the north-northwest-trending anorthosite boundary and around the anorthosite inlier of Mount Howley, south of Fischell's Brook.

Carboniferous rocks also outcrop in a broad trough aligned southwest to northeast between the Precambrian massif of the Northern Peninsula

TABLE II

COMPOSITE SECTION OF THE CARBONIFEROUS ROCKS OF THE CODROY SHORE, ST. GEORGE'S BAY

PENNSYLVANIAN	BARACHOIS GROUP		5000 '	Conglomerate, sandst., arkose, shales & sm. lenses of coal; plant frags.
			Fault with v. small offset	
MISSISSIPPIAN	CODROY GROUP	Woody Pt. sandst.	700'	Cross-bedded & ripple - marked coarse to fine-grained sandst.
		Woody Cove shales & lst.	500 '	Fossiliferous marine shales, limy shales & sandsts.
		Black Pt. Lst.	80'	Limestone & red-grey shales
		Codroy shales and gypsum	3700'	Gypsum Red massive shale Gypsum Red & green shales Red, green & grey shales & sandy shales
			Fault contact	
	ANGUILLE GROUP	Snakes Bight shale	Fault	Thin to thick-bedded black to greyish shale with numerous veins of calcite & quartz
			2600'	Thin-bedded grey sandsts. with plant frags., massive sandsts. with dark grey to black shale partings; thin to thick bedded sandsts., ripple-marked & cross-bedded & intercalated with beds of conglomerate.

(after Hayes & Johnson, 1938.)

and the lower Palaeozoic granitic terrane east of Grand Lake. Structures there are characterized by simple plunging folds, sometimes complicated by longitudinal faults. The rocks are similar in lithology to those found farther southwest.

Evidence of post-Pennsylvanian orogenic movements is furnished by the folded and faulted structures within the Carboniferous rocks. The alignment of these structures follows the dominant northeasterly trend of older structures in western Newfoundland, but it is thought that the effects of this orogeny were less severe than previous ones, in that no regional metamorphism took place. Betz (1943) assigns a post-Carboniferous age to the initiation of the Long Range fault. He quotes the observation by Hayes and Johnson (1938) of an easterly dip of 65-75 degrees on the Long Range fault in the Bay St. George area, and a relative movement of Precambrian rocks (as the rocks of that part of the Long Range were then thought to be) over Carboniferous. Evidence from other sources points to the reverse nature of this fault, according to Betz. Post-Carboniferous faults in the White Bay region appear to have easterly dips, with the latest movement in the vertical plane occurring on the east, suggesting thrust faulting.

No rocks younger than Pennsylvanian are found in Newfoundland, except for the Pleistocene deposits associated with the glaciation and deglaciation of the island. Such absence could possibly be taken as evidence of a long erosional period affecting the surface of Newfoundland throughout post-Pennsylvanian time. However, it is possible that deposition took place one or more times during that period, over at least a part of the island, and that all traces have been subsequently removed by denudation. Since the implications of the former existence of a cover rock for the development of the landscape are profound, as they are in the Appalachian Mountains of the eastern United States, it is of considerable importance to ascertain whether such a cover could ever have existed in Newfoundland.

Douglas Johnson's now classic theory of regional superposition of Appalachian drainage (Johnson, 1931, chapter II) demands the former extension of the Cretaceous Coastal Plain rocks of the eastern United States 125 to 200 miles northwestwards of their present inner margin. From Johnson's (1931) figure 11. (p. 29) the approximate furthest inland extent of the Cretaceous cover can be judged by tracing a line between the former headwaters of rivers which acquired southeasterly courses upon this cover. Examples of these rivers include the North Branch Susquehanna, the upper Delaware, the Housatonic, the Merrimac and the Saco. This line trends southwest to northeast through the Finger Lakes region and the upper Hudson into central New Hampshire. Studies of the former extent of the Cretaceous cover northeast of these areas do not appear to have been made, nor was possibility of such a northern transgression entertained. Projecting northeastwards the line roughly traced through the New England states, it seems probable, owing to the relatively low-lying nature of the southern parts of Maine, New Brunswick and Nova Scotia, that the Cretaceous sea also transgressed those areas. Such reasoning does not lead to any conclusions; rather it prompts other lines of approach to the problem. One such approach lies in the interpretation of seismic profiles of the continental margin of eastern North America.

Seismological studies of the continental margin province off New England,[1] Nova Scotia,[2] and Newfoundland [3] have revealed the existence of several rock layers of distinct seismic characteristics. Oceanic crustal rocks, with high velocities of propagation of longitudinal waves, underlie the outer margin of the continental shelf and the whole of the ocean floor. Above this, "basement" rocks with lower seismic velocities occur, solely on the continental shelf. These "basement" rocks have been identified with lower Palaeozoic and possibly Precambrian granitic rocks. Above this layer are found what are termed "consolidated sediments", having a lower seismic velocity than underlying layers, identified with post-Palaeozoic and pre-upper Cretaceous sedimentary rocks. Occasionally, a layer of "semiconsolidated sediments" has been encountered and is identified with lower Cretaceous rocks. Above the "consolidated sediments", and having the lowest seismic velocities, are "unconsolidated sediments", identified as upper Cretaceous, Tertiary and Quaternary in age.

Engelen (1963) has recently challenged the overall interpretation by Heezen et al. (1959) of the structural relationships of the various units of continental marginal rocks as due to large-scale warpings and subsidence during deposition of a continental terrace wedge. The discontinuities which appear on seismic sections Engelen has interpreted as major faults or steeply-dipping downfolds (fig. 3.2). By reconstructing the interfaces of the rock units along these discontinuities, he has been able to trace the tectonic and depositional development of the continental margin in three

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(1) Drake, C.L., et al., (1954)

(2) Officer, C.B., and Ewing, M., (1954)

(3) Press, F. and Beckmann, W.C., (1954)

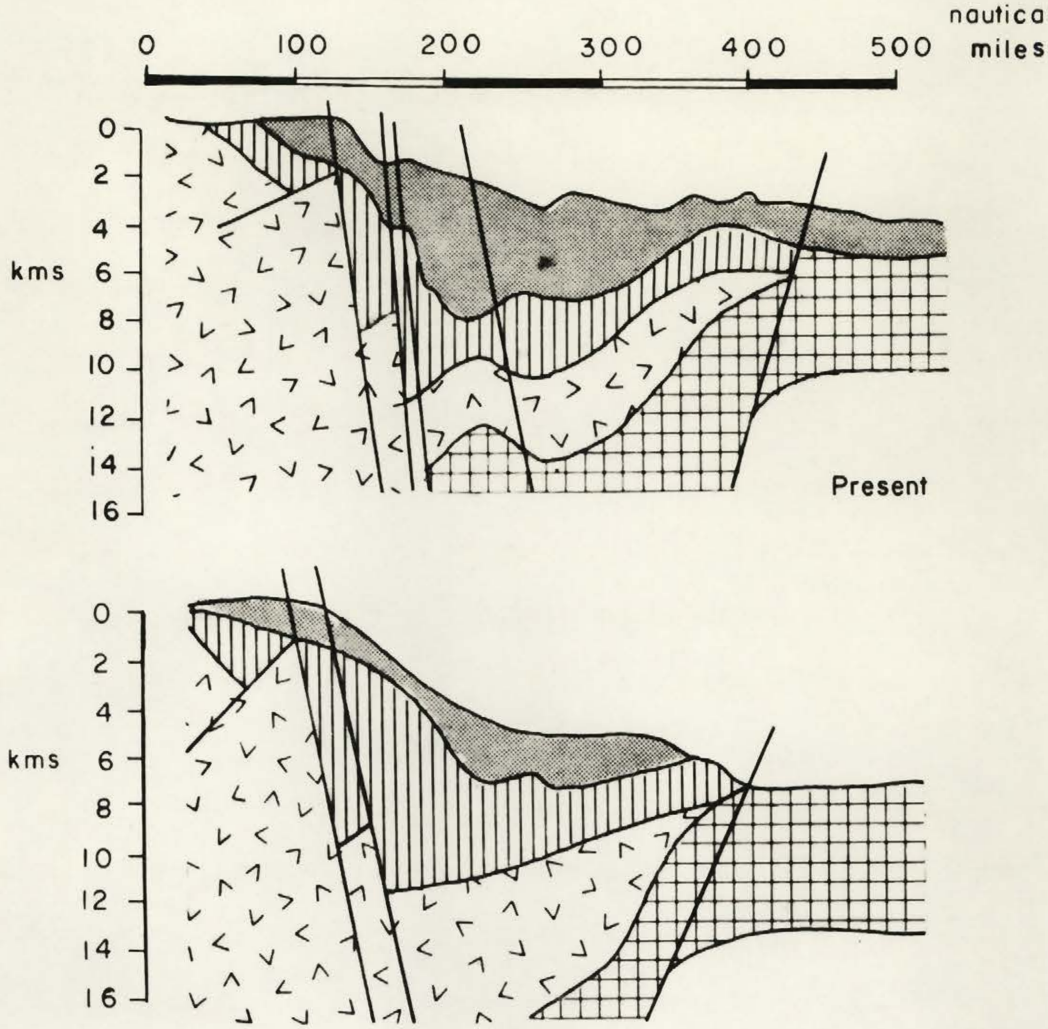
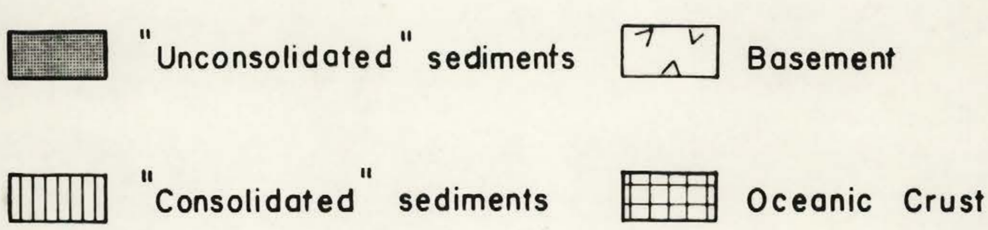


Fig. 3.2. The Development of the Continental Margin off Southern Newfoundland (after Engelen, 1963)



stages, as follows: -

(i) Originally, continental shelf, slope and rise formed one structural unit, covered by an almost homogeneous stratigraphic sequence in its various parts.

(ii) Block-faulting, leading to graben formation (Fig. 3.2, lower diagram) disrupted this crustal unit, causing vertical displacements of its different parts. These movements began in the early Cretaceous in the northeastern section of the continental margin and moved southwards, so that the margin off the Bahamas was not affected until after the Miocene. Thus, the thickness of the undisturbed sedimentary layer increases southwards.

(iii) Later, syntectonic and post-tectonic sedimentation caused variations in the thickness of the sedimentary columns covering the tectonic blocks (Fig. 3.2, upper diagram). Therefore, greater variations in thickness of the rocks deposited after graben formation occur in the northern continental margin, for example, off Newfoundland.

From these conclusions it follows, if downward movement of the continental margin, (Engelen does not discuss causes, but the most likely cause would seem to be isostatic compensation for sedimentary loading) occurred in the early Cretaceous off Newfoundland, that a corresponding upward movement of the continental area occurred approximately simultaneously. There might have been a time lag but no estimate can be given of its length. The relative movement between undisturbed parts of the ocean floor and the continent in the Newfoundland area is estimated by Engelen to be ±2000 fathoms (or ±3.8 kls.), a movement which was likely to have effectively removed continental land areas from the influence of the succeeding upper Cretaceous marine transgression. The downward movement of the continental margin off Newfoundland and the Maritime Provinces took place earlier than in areas further north and it might be accounted for as the result of greater sedimentary loading of detrital material brought by rivers flowing southeastwards from a divide which lay far to the northwest of the coastline of the time, possibly in the Canadian Shield. Farther south, the Atlantic watershed probably was of lesser areal extent because of the proximity of the Appalachian divide to the coastline.

The implications of these conditions for landsurface development and drainage evolution in Newfoundland will be discussed in chapter eleven following presentation of the field evidence of upland surfaces.

Chapter Four: Upland Surfaces - General Considerations and Methods of Analysis

(i) General Considerations.

The definition of an upland surface is an arbitrary one, depending upon the position of a surface within the landscape more than upon its elevation. An upland surface may therefore be considered as a planation or erosion surface, not genetically defined, found in the "uplands" of a region. Hence, in Newfoundland such surfaces are found usually above about 1000 feet and, when viewed in profile, occupy much of the skyline. In an area of lower elevation and relief, such as Nova Scotia, an upland surface may be seen at lower elevations. The surface occurring at 700 to 800 feet atop North and South Mountains in Nova Scotia was termed an "upland plain" by Goldthwait (1924) and qualifies for the name "upland surface" in the sense used in this study.

In order to present as complete a picture as possible of the erosional stages which have affected the Newfoundland landscape lower erosion surfaces, not designated "upland surfaces", are here taken into account. These are found mainly in the wide lowland areas, in the major valleys and fringing the mountains and tablelands. However, on the whole, attention will be directed to the upland areas in this study.

The approach to the study of erosion surfaces is a hazardous one. For some time, actually since the inception of the Davisian geomorphological ideas which laid the foundations for their interpretation, the very existence of such features has been questioned. When the existence of erosion surfaces has been accepted, doubt has often been expressed as to the possibility of their origin by processes of submarine and/or subaerial denudation, periodically interrupted by negative movements of base level (Rich, 1938; Hack, 1960; Kennedy, 1962).

Objectivity of approach is an obvious necessity, but very difficult to practice. Burton (1963) has recently remarked that "the moment that a geographer begins to describe an area, ..., he becomes selective (for it is not possible to describe everything), and in the very act of selection demonstrates a conscious or unconscious theory or hypothesis concerning what is significant" (p. 156).

It is quite commonly the case that two workers, possessed of opposing opinions concerning erosion surfaces, will find evidence from the same landscape to support their respective views. Often, one piece of evidence can itself be used to support either view, depending upon its interpretation. In general, morphologic evidence can be examined with some degree of objectivity, but its interpretation depends more upon the view of the researcher. An example may be found in the earlier geomorphological works of Lester C. King (1950, 1953), whose grand scheme, involving the world-wide age equivalence of erosion surfaces over continents which have drifted apart from a parent super-continent, depended upon his interpretation of hillslope forms and the processes moulding them. King's views were opposed to those of European geomorphologists whose ideas, born of Davisian tradition, did not allow them to accept such universality, and prompted them to restrict the processes of landscape development envisaged by King to areas of certain climatic and structural conditions, allowing their own theories to hold sway "nearer home".

It was with the hypothesis of multiple erosion surfaces in mind that this study was undertaken. Such an hypothesis has been shown to account adequately for the geomorphic evolution of nearby areas, such as the Appalachian Mountains of the eastern United States (Johnson, 1931). Opinions have differed, however, as to the processes responsible for the development of the erosion surfaces and as to the time of their formation. These are two major problems which are investigated in this study.

Studies similar in nature and purpose to the present one have used two lines of evidence: that from contoured maps and that from the field. The former consists of cartographic analysis, employing techniques of varying degrees of refinement, such as generalization of contours, various types of profiles and statistical methods. The latter employs surveying techniques and field observation in conjunction with the use of contoured maps.

(ii) Cartographic analysis.

The short time available to the writer for laboratory work, and the large size of the field area, prohibited the use of all but the simplest cartographic techniques. Those used were: (1) generalized contours on a scale of 1:250,000, over an area far wider than that covered in the field, and (2) superimposition of composite profiles, with a horizontal scale of 1:50,000 for most of the area covered in the field and extending some distance out of that area when interesting relationships came to light.

(a). Generalization of contours.

From the 1:250,000 generalized contour map (end plate 3) it can be seen that, along the western front of the Long Range mountains, wide summit areas appear at between 1800 and 2000 feet. In the extreme southwest these overlook the marked structural depression of the Codroy rivers; but further northeast they are set some way back from this front to the east of a platform at 1600 feet. This last is also well developed between the higher areas and to the east of them, where it is seen as a wide, level plain extending some thirty or forty miles east of the 2000 foot summits of the Long Range, with a maximum relief of 200 feet, and with isolated hills rising 100 to 200 feet above it.

Further northeast along the Long Range Mountains other areas at 1800 to 2200 feet are found, more extensive than those to the southwest, overlooking the broad lowland of Harrys River above both the eastern and western shores of Grand Lake. These apparent levels are developed upon the granites, granite-gneisses and small outcops of mafic rock of the Long Range Igneous and Metamorphic Complex and show no relationship to rock type. Less extensive but significant areas at above 2000 feet are found upon the upstanding gabbroic mass of the Annieopsquotch Mountains, about twenty miles east of the Long Range front.

The 1600-foot level appears west of the Long Range in the summit plain of the Cape Anguillie Mountains, well developed across the anticlinal structure. A link between this level here and in the Long Range is found upon the small upstanding mass of Codroy Group rocks at the northeastern end of the Codroy depression. West of the Long Range front and northeast of the Cape Anguille Mountains the landsurface slopes gently to the sea from the foot of the Long Range scarp at about 1000 feet over a distance of up to fifteen miles. The surface is very even and bears no relationship to the folds and faults in the Carboniferous strata beneath.

Thus, it seems that a high level at 1800 to 2200 feet occurs extensively throughout the western Long Range, especially near to the scarp which marks a major zone of dislocation, extending from Cape Ray in the far southwest to White Bay in the northeast. Below this level, and partly or wholly surrounding it, there is a level at about 1600 feet developed as an extensive plateau east of the Long Range, as smaller platforms along its eastern front and as dissected pl¢ateau, for example in the Cape Anguille Mountains. From the trend and spacing of the generalized contours the lower surface would seem to slope to the southeast, east of the Long Range front and towards the west on the western side. No slope is discernible upon the higher level from the map.

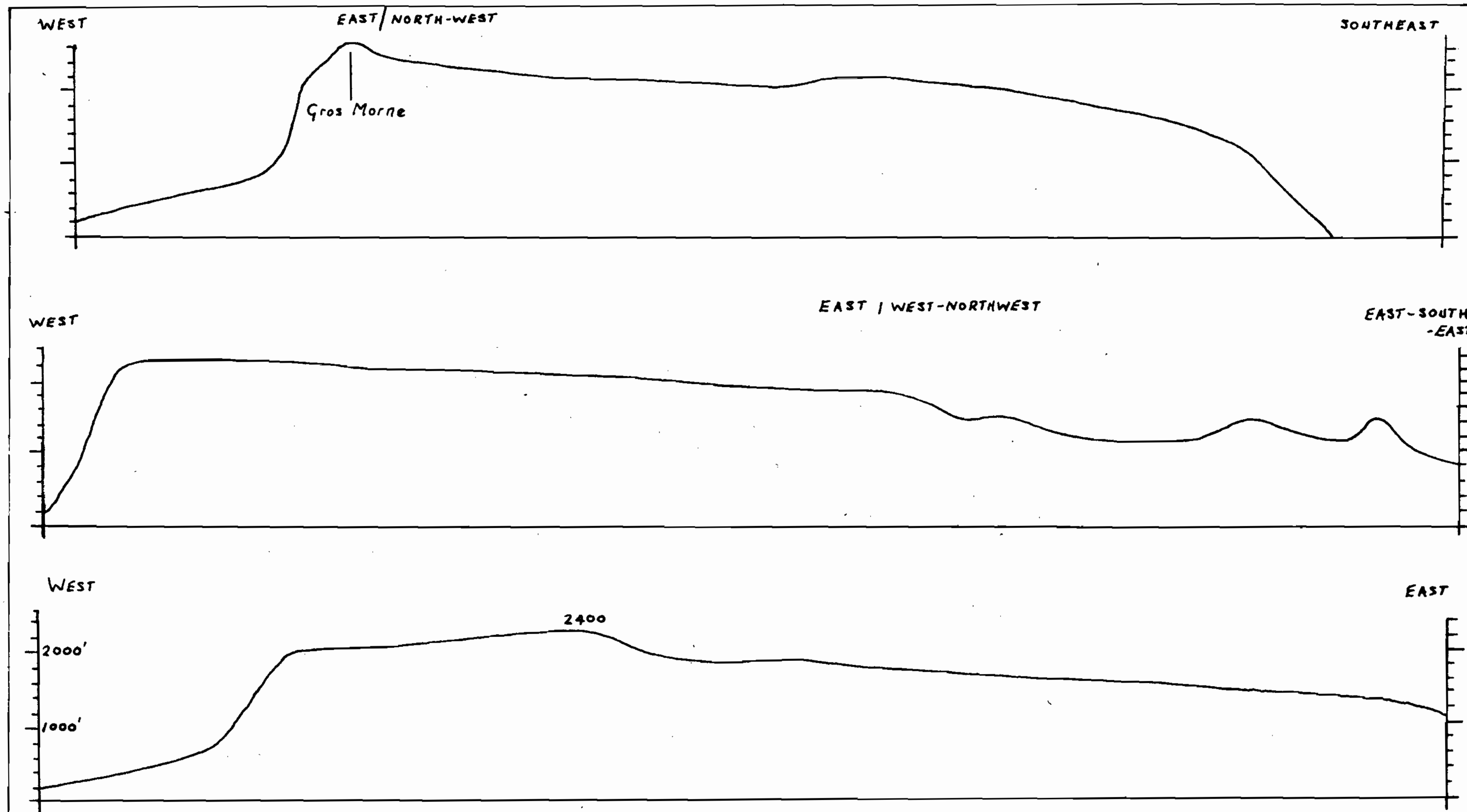
The southern limit of the 1600-foot level is marked by a pronounced break of slope, unrelated to any geological boundary, along a line roughly parallel to the south coast of Newfoundland. This slope to sea level seems fairly regular, but slight flattenings appear in places at 1000 feet and 800 feet. The nature of this slope and the break above it are better seen on the profiles to be discussed below.

West of the Long Range and north of St. George's Bay, in the physiographic region VI, (end plate 1) the landsurface shows nothing of the complexity of rock structure in the region. Across the clastic and carbonate terrane alike is developed a level at approximately 1200 feet. In sub-region VIa the surface slopes down to the axis of a trough carved by Harrys River and its tributaries and streams flowing north to Humber Arm. In sub-region VIb, between Humber Arm and Bonne Bay, the surface shows a similar disregard for geology and structure and a level at between 1200 and 1400 feet is developed across it. This, however, slopes up to a central axis which culminates at about 1600 feet.

West of sub-regions VIa and VIb lies region VIII, comprising the four bodies of the Bay of Islands Igneous Complex (subregions a,b,c,d,& e). These appear very strikingly on the generalized contour map by virtue of the steep scarps separating them from the surrounding sedimentary lowland.

The Lewis Hills VIII (a) possess a summit level at 2200 to 2400 feet developed across the two rock types of the complex, the gabbro and the serpentinised dunite, A small remnant appears above this level at 2600 feet. Below this level a fairly wide area at 1600 to 1800 feet occurs on the west side of the mass of the Lewis Hills, and at 1800 to 2000 feet on the eastern side. Lower levels do not appear on the 1:250,000 scale with a 200-foot contour interval. The Blow me Down Mountains(VIII b) present a well-marked level area at 1800 feet, which is surmounted by three higher areas displaying an accordance of elevations at 2200 to 2400 feet. Again these levels bear no relationship in altitude and form to the underlying rocks and structures. The North Arm Mountain VIII (c) - St. Gregory Highlands VIII (d) mass shows a markedly smooth surface, but with a greater height range than that on the other masses. It slopes from the summit of North Arm Mountain at above 2200 feet northwestwards to 1600' where the land dips more steeply to the Trout River ponds area. There is also a level at 1600-2000 feet sloping northeastwards to the same area. Table Mountain, Bonne Bay VIII (e) is surmounted by a level area which slopes from above 2200 feet in the central area west to 1800 feet and east to 2000 feet.

To the northeast of Bonne Bay the Long Range Mountains of the Northern Peninsula, composed of Precambrian granites, granite-gneisses and schists of the Grenville Series, extend for some 200 miles northeastwards to the Strait of Belle Isle. Comprising the backbone of the Northern Peninsula, these mountains are delimited to the east and west by faults (see regional description, Chapter 12). The mountains rise from the back of



SCALE: 1,250,000 ⟶

VERTICAL EXAGGERATION: X 10.

Figure 4.1. Three Profiles across the Southern Part of the Northern Peninsula - from generalised contours

the coastal plain at about 600 feet up to 2600 feet atop the scarp. Occasionally, the higher summits of the mountains are set back, east of the scarp above a lower level at 2200-2400' which forms the predominant level over much of the western edge of the Northern Peninsula. Generally, the surface of these mountains appears as a plane sloping gently east and southeast, but the profiles constructed from the generalized contour map reveal noticeable flattenings on this slope (fig. 4.1).

(b) Superimposed composite profiles

In contrast to simple line profiles, composite profiles are drawn of a strip of country between two parallel lines, in this case one mile apart. Whereas, in the case of the simple line profile, the intersection of contour and profile line is projected on to the grid for a composite profile the highest point between two successive contours between the parallel boundary lines is projected on to the grid. Figure 4.2 shows the different methods of construction and the results of the use of each for an ideal contoured strip of country. The resulting composite profile represents a trace of maximum elevation between the boundary lines, the emphasis being given to the hilltops and plateaux. Representation of the valleys, with regard to their true depth is false but, since this work is concerned with the uplands and not the valleys, the importance of this falsity is minimised. In a more detailed study truer representation of valleys is necessary. The advantages of this type of profile are (i) it enables the form of the land to be generalized to a degree which does not render it invalid as a method of physiographic portrayal, (ii) its emphasis

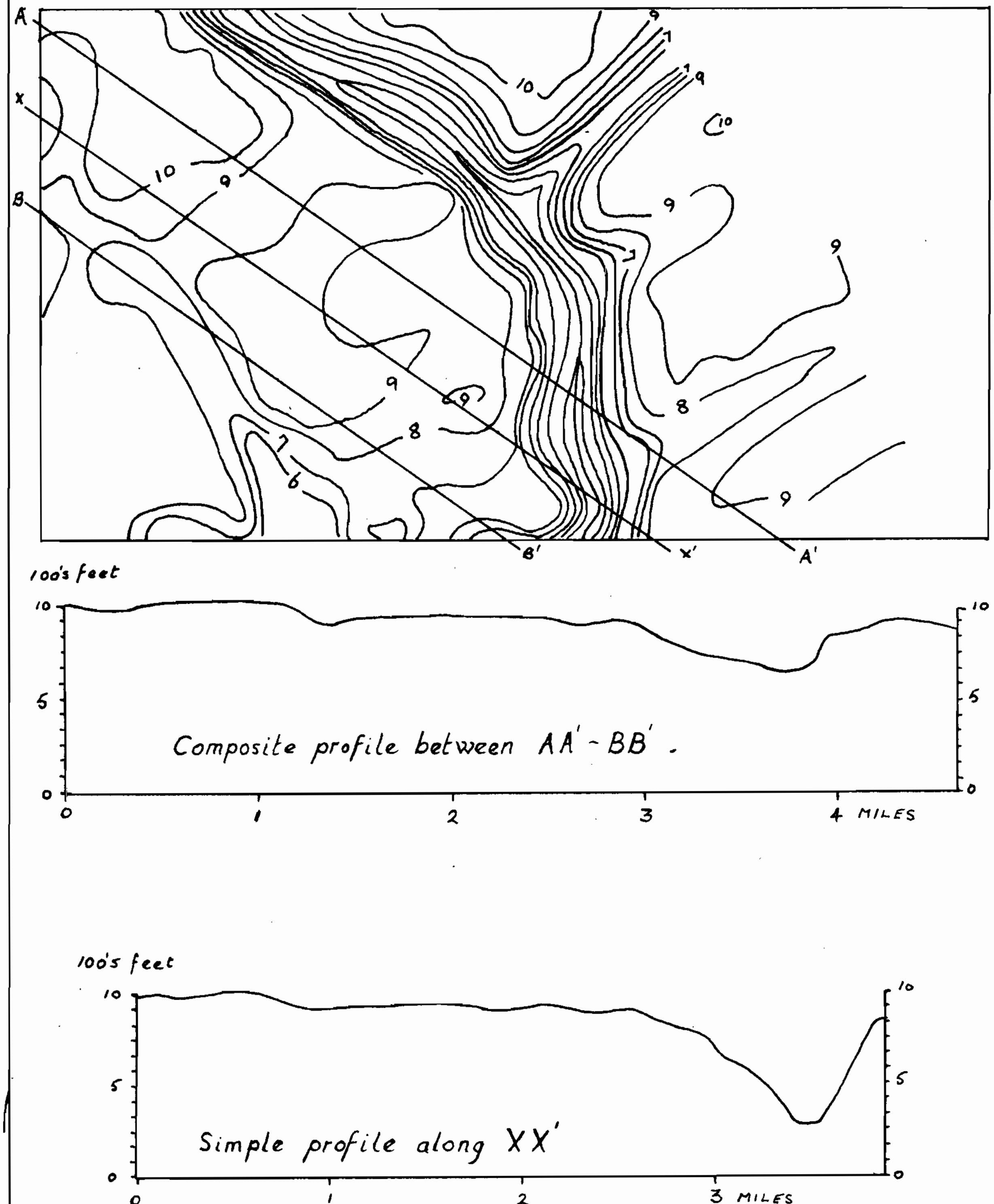


Figure 4.2. Comparison of composite & simple profiles.

is on that part of the landscape which is under examination in this study, and (iii) it enables a whole area to be portrayed in this general way, rather than merely that infinitely narrow strip of country along the line of a simple profile which must, unless chosen subjectively, show many false slopes where the profile runs at an angle to the line of greatest slope.

The profiles were drawn in a northwest to southeast direction because this runs at right angles to the dominant structural trend of western Newfoundland and it parallels the main drainage lines. Complications arise when the structural trend changes, as it does in the southwest, where the surface slopes southwards. Hence, in diagram (1) of end plate 5, the southeastern end of the profiles is really steeper than shown since the profile line runs at 45 degrees to the surface slope. Three selections of five adjacent profiles and one of four were made in order to represent typical tracts of country. Two sets of profiles were chosen which traverse the Cape Anguille Mountains, the Codroy trough and the Long Range Mountains (end plate 5) and two were taken covering the Long Range Foreland and those Mountains and extending some twenty miles eastwards to include the level plateau areas termed the "High Central Plateau" by Twenhofel (1940) (end plate 6).

The first set of profiles, traversing the southwesternmost parts of the Cape Anguille Mountains, the Codroy trough and the Long Range (end plate 5, diagram 1) shows that a well-developed level exists atop the Cape Anguille Mountains at 1500' to 1600'. In does not appear to be warped in any way, any slope that it shows being to the northwest and southeast from a central higher area. There are also signs of a lower level found around the margins of the mountains at 1300-1400', sloping more markedly in the same directions as the upper level. Signs of minor benching are visible on the steep seaward slope of the mountains at about 500 to 600 feet, which could be similar to benches at 400-500' on the southeastern side of the mountains, bordering the Codroy trough. Apart from these benches, the flat tops of the mountains descend 1200' to the sea on one side and the Codroy trough on the other. The steep western side of the Codroy trough coincides with the boundary between the Anguille Group and the Codroy Group along the eastern limb of the anticline. It is a structural slope but not a scape, since the bedding planes of the rocks comprising it are more or less parallel to the slope. The Codroy trough appears on the profiles as a flat-floored, steep-walled depression about three miles wide, with signs of benching on either side. The rocks underlying it belong to the upper Mississippian Codroy and the Pennsylvanian Barachois groups, rocks less resistant to erosion than those on either side of the trough. The origin of this feature presents an interesting problem. Was it once filled with sediments to the level of the surrounding plateaux and subsequently hollowed out by rejuvenated denudation? Or was it let down between faults at one or more times in the relatively recent geological past? The answers to these questions will have to be postponed until after discussion of the field analyais of upland surfaces bordering the trough. The eastern side of the trough presents a steep wall to the west, a scarp developed along the Long Range reverse fault line. Signs of a threshold bench appear at 400-500 feet at the base of the scarp on the profiles, matching a similar feature seen on the western side of the depression. Above this, the scarp rises steeply to a flatter area at between 1300 and 1400 feet, signs of which are also seen on both sides of the Cape Anguille Mountains. Another narrow high-level platform occurs above this at about 1600 feet, a level which is continued in the floors of wide upland valleys, and which reappears in eastward-sloping plateaus east of the highest summits atop the Long Range Mountains. This level also forms the summit plain of the Cape Anguille Mountains. These highest summits appear to conform to a general level of about 2000 feet. The 1600-foot level gently declines to the southeast for a distance of twelve miles, where it has fallen to 1000 feet. Above it, isolated summits and small plateaus rise to 1500-1600 feet. At its south-eastern extremity this level increases markedly in slope, from over forty feet per mile to over 140 feet per mile. The trend of the profiles in the section where the gradient increases is at forty-five degrees to the slope of the land surface and, therefore, the true slope of the surface is calculated to be about 200 feet per mile. Such an abrupt change of gradient suggests that the latter surface is not a part of the former and that it is possibly an older surface, steeply tilted and bevelled by the more widely developed inland surface at 1600-1000 feet. The higher summits standing above this last-mentioned surface seem to represent accordant remnants of the surface seen in the highest summits found atop the Long Range Mountains.

Similar relationships are seen on the set of profiles across the Cape Anguille and Long Range Mountains further northeast (end plate 5, diagram 2). The Cape Anguille summit surface at about 1600 feet continues across the Codroy depression to the edge of the Long Range Mountains and the plateau areas further east. This surface does not seem to slope so markedly southeastwards as it does in the previous set of profiles, maintaining a level of 1400-1600 feet for a distance of about sixteen miles. It then increases in gradient from less then 10 feet per mile to about 176 feet per mile. Such relationships suggest a similar origin for the steep facet as that given above for the feature seen in diagram 1, end plate 5. The Long Range Summit surface is represented by 2000-foot summits along the western edge of those mountains and by accordant monadnocks at 1800-1900' to the southeast.

The first set of profiles across the Long Range Foreland and those mountains (end plate 6, diagram 1) runs from the coast of St. George's Bay forty miles southeast to the western shores of the fjord-like feature of La Poile Bay on the south coast of Newfoundland. From the former coast the land surface slopes up to the foot of the Long Range scarp at about 1000 feet in a distance of thirteen miles: a gradient of 77 feet per mile. The slope is regular and is an erosional feature bevelling the folds and faults in the Carboniferous rocks of the St. George's Bay lowland. The scarp behind this surface is up to 500 feet high but the profiles show no definite surface to the east. This is probably due to a difference in the trends of profiles and landsurface slope, since further east there are definite signs of a level at around 1600'. A high summit level appears in the Long Range at 1800-2000' and extends eastwards almost to the shores of La Poile Bay. The question arises as to the origin of the scarp bounding the Long Range Mountains to the west. If this is an erosional feature, due solely to the differing existances to erosion of the Carboniferous sediments to the west and the granites and granite-gneisses to the east of the fault line, the recent diastrophism which raised the Long Range Mountains was of equal effect on either side of the fault, and the sloping surface of the Foreland is younger in age than the youngest of the surfaces bevelling the scarp. Or, if diastrophism was unequal in effect on either side of the fault, raising the Long Range Mountains above the foreland along the fault-line, this would imply that the Foreland surface is of the same age as the latest surface developed on the Long Range before the diastrophism took place. Since re-activation of old faults is a well-known tectonic occurrence, this latter possibility may be more worth entertaining than the former, but no more definite statement can be made before analysis of the field results.

The set of profiles drawn across the Foreland and the Long Range mountains further northeast (end plate 6, diagram 2) is more complicated. In this area the boundary between the uplands of the Long Range and the Carboniferous and drift lowland of St. George's Bay strikes northwards, along the edge of an anorthosite mass lying to the west of the Long Range fault. Thus, the position of this boundary on each profile changes with reference to the base line marked on the profiles. A foreland is still discernible, having the same slope as before, but at a varying distance from the coast this is replaced across a scarp-like feature by an upland area at 1100-1300' on the anorthosite and the Long Range igneous and metamorphic complex. Above this level and further eastwards a very well developed level is seen at 1600'. It is found west of the Long Range summit surface as narrow platforms and upland valleys and to the east of this as an extraordinarily level plateau extending twenty miles southeastwards without a perceptible change in elevation. At its easterly extremity on this set of profiles, northwest of the upper Lloyds River , isolated summits rise above this plateau to 1800 feet and probably represent outlying remnants of the Long Range summit surface which is found better developed near the western edge of these mountains at 2000-2200 feet.

The generalized contour map and the composite profiles have enabled a broad picture of the physiography to be given. It remains now to describe the landscape, enumerating the levels, ascertaining their elevations and their relationships to each other and to drainage lines, in order to see whether this is a true enough picture and to solve the problems raised by this preliminary analysis of the generalized contours and profiles, and broader problems of landscape evolution throughout western Newfoundland.

(iii) Field Methods.

Field conditions in western Newfoundland make a detailed study of erosion surfaces extremely difficult. In order to compensate for the impossibility of studying all areas in detail the writer carried out field studies in a number of areas, the choice of which was dictated more by their accessibility than by their inherent value and significance for the solution of the problems under investigation.

The use of surveying instruments was almost entirely prohibited by the terrain and vegetation conditions, except for the limited use of an Abney Level. Aneroid traverses of sufficient length to be of any value could not be made due firstly to the slowness of progress on foot across difficult terrain, through thick forest and impassable bog land and, secondly, to the inconvenient situations and distribution of bench marks and horizontal control points. For example, even in the area which, from the topographic maps, would appear to offer good opportunities for aneroid traversing, that is the southern end of the Long Range Mountains bordering the Codroy depression, terrain and vegetation conditions ruled out any such work. The bench marks in that area are situated on the railroad track which in places closely approaches the foot of the Long Range scarp, and horizontal control points are found atop these mountains up to a distance of one mile from their western rim. However, the traverse from the railroad tracks to the mountain rim cannot be made in the three hours maximum time allowable for uncontrolled aneroid traverses. Where a fixed point is not available for closure of the traverse atop the mountains and closure involves a return to a bench mark on the railroad, the problem is insuperable. Weather conditions also were not often amenable to the execution of uncontrolled aneroid traverses, since stable, anticyclonic pressure conditions were rare in occurrence during the field season of 1963.

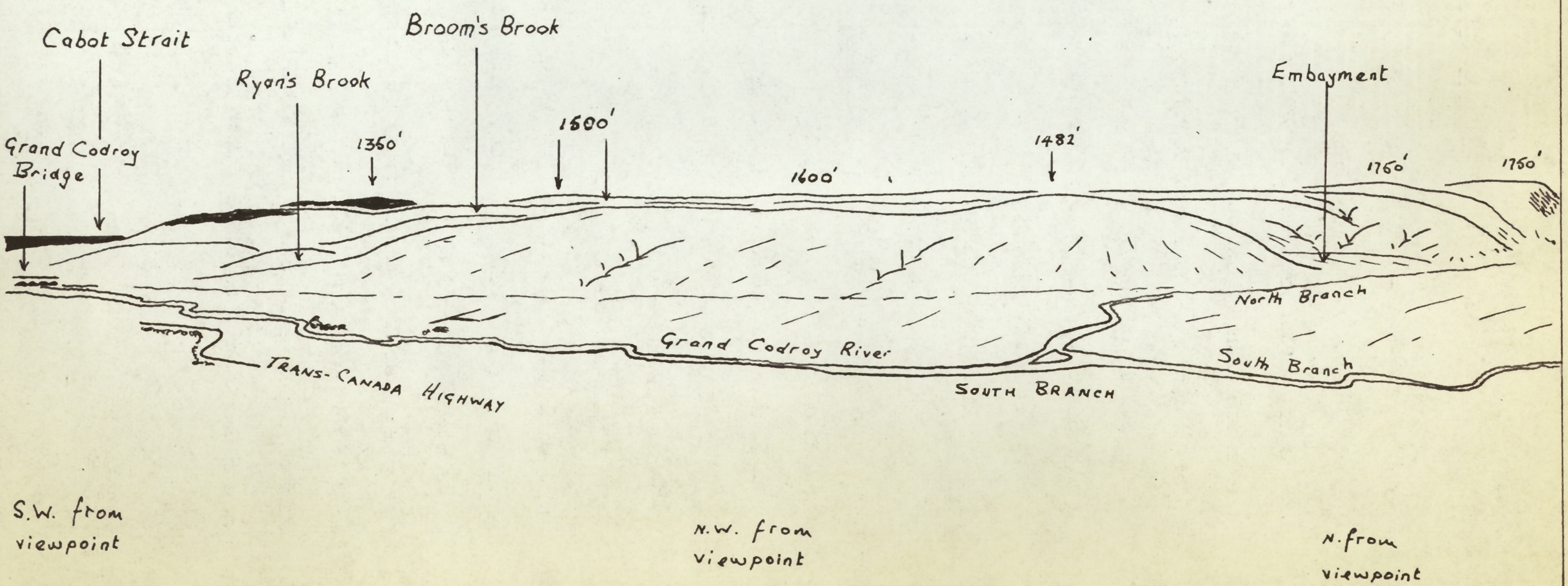
Field work, thus, was largely visual, and entailed transferring to the 1:50,000 contoured maps the remnants of the erosion surfaces whose extent was determined by eye. The contour pattern of the maps was a good guide to the areal delimitation of the various levels, but the 50-foot contour interval was often a hindrance, necessitating some uncertain judgements as to their exact extent.

By following the surfaces by eye and on foot in each of the areas visited, continuity of surfaces was fairly well established, both on the condition of elevation and that of the position of a surface relative to others. The picture built up in this way is inexact to a degree, but is considered valid for the purposes of this study.

Chapter Five: The Cape Anguille Mountains

The Cape Anguille Mountains comprise a thirty-mile-long upland plateau whose summits lie at a general elevation of 1500-1600'. They extend from Cape Anguille in the southwest, northeastwards to overlook the Highlands River. The bedrock is predominantly Mississippian sandstone, and is arranged in a broadly anticlinal structure which pitches both northeast and southwest (end plate 4). Steep slopes bound the mountains on all sides and on the northwest and southeast margins these slopes reflect the steep dips of the rocks off the axis of the anticline and marginal faulting. The surface presents a remarkably even skyline, especially when viewed from the summits of the Long Range, across the Codroy depression (fig. 5.1). Streams drain the Cape Anguille Mountains in all directions, mainly to northwest and southeast, from a central divide which trends roughly along the axis of the structure (end plate 7). The lower and middle reaches of the valleys are deep and steep as a result of recent rejuvenation. In the upper parts of these valleys rejuvenation has only yet succeeded in producing a small incision in the broad shallow upland valley floors, usually to a depth of 100'. Structural weaknesses would seem to be of little importance in guiding the courses of most of the transverse streams, but many smaller streams tributary to them have coincidences of alignment which follow the strike of the rocks and which, therefore, may be due to lithological weaknesses or control by the bedding planes. Faulting may also guide some stream courses, such as those of Broom's and Ryan's brooks in the southwest of the mountains, which have a pronounced northeast to southwest alignment.

Figure 5·1. Panoramic View of the Cape Anguille Mountains from the Long Range Mountains above the village of South Branch.



The line of the former brook is continued to the northeast in the steep head wall of a wide embayment in the eastern flank of the mountains and, further northeast, in the line of the headwaters of small brooks flowing from the mountains to the Codroy depression. This major alignment of features is due to faulting in the southwest according to the map in Hayes and Johnson's (1938) report, but further northeast may be due to subsidiary flexuring of the eastern limb of the anticline. The Cape Anguille Mountains were studied in three areas which are reasonably accessible; one in the extreme southwest and two in the northeast.

(1). The Southwestern Area.

The southern part of the mountains is reached by a sheep track which climbs the steep slope above the village of Cape Anguille to the plateau above the headland of that name. From Hayes and Johnson's (1938) map it appears that the upland is comprised entirely of rocks of the Anguille Group. Where it is exposed on the steep southern edge of the mountains, the bedrock is seen to be a flaggy, fine to medium-grained sandstone, predominantly quartzose but with some arkose. It is thinly bedded and splits easily along bedding planes, especially under the action of frost-shattering which in places has sorted the individual weathered plates concentrically around a central finer core with the axes of the planes vertical. The dip appears to be at thirty degrees to the east, which suggests that either folding or faulting has complicated the general westerly dips along the western flank of the Cape Anguille anticline.

Cape John

Long Range Mountains

northeast east southeast

Figure 5.2. Panoramic view north to east across southern Cape Anguille Mountains. (From viewpoint A. fig. 5.3).

Once the surface of the plateau is gained, at about 1150', a broad panorama of level plateau country stretches away to the northeast, east and southeast (fig. 5.2). To the southeast (fig. 5.3), the view is of a spur possessing a marked flattening at between 1100' and 1300', about $1\frac{1}{2}$ miles long between its distal and proximal ends. Towards the back of this level marshy vegetation is dominant and is an excellent indicator of its extent since, at the break of slope separating it from the isolated hilltops above it, the vegetation cover changes to a darker coloured dwarf spruce-juniper-scrub type, a change which is discernible at a distance of one mile (fig. 5.2 right.) From this break of slope at 1300' this spur level, and its counterpart across an incised valley to the northeast of it, appears almost level, declining at an angle of less than $\frac{1}{2}$ degree. Above this break the land rises more steeply to a summit area at above 1350', crowned by two isolated flat-topped hills. These hills represent parts of a higher erosion level. More extensive remnants of this higher level are seen on interfluves between streams draining the western slope of the mountains to the northeast of this spur just described. (See map of surfaces, fig. 5.3).

The view northeast shows two major levels. Just below the skyline spurs sloping with the drainage to the west show marked flattenings in the range of elevation 1100' - 1250'which prompts their correlation with the spur levels further southwest (fig. 5.3). The skyline itself is remarkably even, the greater part comprising a flattish, whalebacked ridge sloping in all directions from the indeterminate crest of the mountains at above 1400' to about 1300'. This level is equivalent to that in the southwest, small

Fig. 5.3. The Erosion Levels of the S.W. Cape Anguille Mountains

1350'-1400' levels
1150'-1300' "
1000'-1100' "
850'-950' level
750'-850' "
500'-600' "
Geological boundary

remnants of which are seen atop the two isolated hills at above 1350' (see fig. 5.3).

A lower level on some spurs, less extensive but none-the-less significant occurs at 1000-1100' in this southwestern area map (see fig. 5.3). The nature of the spur and ridge-top levels in this area, with regard to their elevations and their relations to each other and to drainage, preclude all but a subaerial origin. They slope with the drainage lines, the break of slope separating them from higher levels is not at a constant height, the pattern of the levels conforms in general to that of the valleys and interfluve levels often "finger" up into the valleys; all features which can only be attributable to fluvial erosion during periods of relative still-stand of base level, interrupted by rapid negative changes of base level.

Before continuing the examination of the landscape it is wise to consider questions raised by the small vertical intervals between the levels so far found (and to be found in the following descriptions) and the height ranges of the levels identified, which is large relative to the vertical intervals. It may seem absurd to speak of a level at 1100-1300' below another at 1300-1400', since the height ranges of the levels are greater than the vertical intervals separating them. However, when dealing with subaerially formed erosion level remnants, it must be recognized that they have a greater original slope and relief than wave-cut terraces and, therefore, have an appreciable height range, depending upon the degree and the extent of their development. While this accounts for the greater height range, the explanation of the small vertical interval must lie in an understanding of the diastophic force at work in the earth's crust and subcrust. The most useful concept, with regard to erosion surfaces, is that of periodic isostatic uplift in response to the relief of load resulting from denudation. Relatively abrupt uplifts of the order of 100' do not seem impossible by this mechanism (Schumm, 1963), although it may be difficult to understand why such a small uplift should occur in one area when, in another, greater vertical intervals between surfaces suggest greater periodic uplifts. Possibly, this may be explained by (a) faulty observations of the vertical intervals between erosion surfaces, or (b) local differences in crustal strength which may require that different amounts of load be removed before there is a crustal response, or (c) differences in denudation rates which cause variations in the time elapsing between periodic responses.[1].

The recognition of many erosion surfaces, separated by relatively small vertical intervals, seems justified in this study by virtue of the recurrence of a pattern in the landscape. Once a pattern of erosion surface remnants is discerned it is a relatively simple task to reconstruct a former landscape and to separate it from a landscape whose remnants occur above and below it. If difficulties arise in comprehending a mechanism responsible for the patterns observed, research should be directed towards the framing of more refined geophysical hypotheses, such as has been attempted by Geyl (1960).

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1. It must be added that these factors are presented as possible explanations of the variations in the isostatic uplifts of the land surface. Other epeirogenic movements and eustatic sea-level changes are important in a discussion of other causes of the variations in the vertical intervals between erosion surfaces.

Figure 5.4. Looking north along Codroy Pond from eastern slopes of Cape Anguille Mountains. Note 1550' - 1700' surface at right, bluff below 1350-1500'surface north of Morris Brook at left.

(ii) The Codroy Pond Area.

From the steep eastern slopes of the Cape Anguille Mountains above Codroy Pond (see Fig. 5.4) a view is obtained of an upland mass lying between these mountains and the Long Range. This mass is composed of Mississippian rocks of the Codroy Group, younger than those of the Cape Anguille Mountains. Although no dip observations are seen on Riley's (1962) geological map, a synclinal axis, just to the northeast probably passes through the western part of the mass, while easterly dips of twenty degrees to the southeast in Pennsylvanian strata suggest that rocks in the eastern part of the mass have similar dips. Therefore, it seems possible that an upfold passes through the eastern half and an downfold through the western half of the upland mass (see end plate 4). These structural complications are not manifest in the form of the land surface, which exhibits smooth summits, declining gently southeast from a high bluff bounding the mass to the north. Several spur and summit flattenings are visible (see fig. 5.5). Three gently rounded summits at above 1700 feet can be interpreted as remnants of an erosion level at that height. They are too small in extent for any surface slope to be discernible upon them. From these summits spurs slope south and north and possess noticeable flattenings at 1500-1650'. The cols between the higher summits, occurring at 1550-1600' are part of this level. That these cols represent former through drainage lines seems unlikely, if only for the reason that there is no likely source area for the drainage flowing through them. Further, the spur levels associated with them slope both north and south from the summits above them, implying that drainage at that stage flowed in both directions from a divide crossing the mass from west to east.

Fig. 5.5. The Erosion Levels of the N.E. Cape Anguille Mtns. & East of Codroy Pond

	1750+' level		1150'-1300' level
	1550'-1700' level		1000' level
	1350'-1500' level		750'-800' level

There is, then, on this upland mass, a summit surface at 1700-1750', below which is found a level at 1500-1650' found on spurs and in the upland cols. The western-most of the southerly projecting spurs bears a flat area at 1400-1450' which appears to represent a lower-lying remnant of the higher level at 1500-1650', rather than part of a separate lower level.

The Cape Anguille Mountains present a steep front to the east above the Codroy Pond defile, which feature is attributable to vigorous downcutting consequent upon recent rejuvenation and in part to erosion by ice or by glacial meltwater (see fig. 5.4). This part of the mountains was studied in the area around the horizontal control point at 1759' above Codroy Pond (fig. 5.5). That summit is attained from the highway with no break of slope recognizable as an erosion level. However, from the slopes leading up to it, various levels can be seen along the eastern edge of the mountains to the northeast. Across the incised valley of Morris Brook the eastern edge of the mountains is topped by a level area at 1500'. It appears to slope northwestwards away from the eastern margin of the mountains whereas streams whose valleys disrupt the continuity of this level flow southeastwards (see fig. 5.5). Less extensive high level spur flats occur on the southeastern side of the hills northeast of Morris Brook at 1400-1450'. These slope towards the brook but it was not possible to discern their longitudinal slope.

The 1759-foot summit is extremely flat and has a gentle slope to the northwest. It is matched by a similar summit to the southwest across the intervening valley, and these two summits appear to represent remnants of a high-level erosion surface at above 1750'. The gentle slopes of these remnants, if not accountable for by grading of formerly steeper slopes, suggest that they were formed by streams heading some distance east of the present eastern edge of the Cape Anguille Mountains. Yet the occurrence of remnants of this level atop the upland mass to the east, suggests that a drainage divide passed from the neighbourhood of these summits on the Cape Anguille Mountains to those on this mass and thence east to the Long Range.

Northwest of the 1759-foot summit a broad flat area, recognizable as an erosion level, occurs on the interfluve between the through valleys of Paul's Gulch and Morris Brook - Shoal Point Brook. It slopes northwest from a faint break of slope at its back at 1700' to 1600' and has equivalents in the summit level described northeast of Morris Brook and in levels at similar elevations occurring on spurs projecting southeast and southwest from the 1750-foot summit level across Paul's Gulch (see fig. 5.6). Looking north from the back of the 1600-1700 foot spur level northwest of the 1759-foot summit, down the valley of Shoal Point Brook (fig. 5.5) an erosion level is seen very well preserved on spurs on either side of this valley, especially on the southwestern side. These spur flats are at 1400-1500' and find their equivalents in the broad interfluve level between Shoal Point Brook and French Brook (see fig. 5.5). Fragments of a still lower erosion level found on a few small spurs at 1250' would seem to be correlative with the level at similar height found more widespread on major spurs in the southwestern field area.

(iii). The Paul's Gulch Area

Overlooking the broad upland valley at the head of Paul's Gulch from the southwestern slopes of the 1759-foot summit, similar features to those

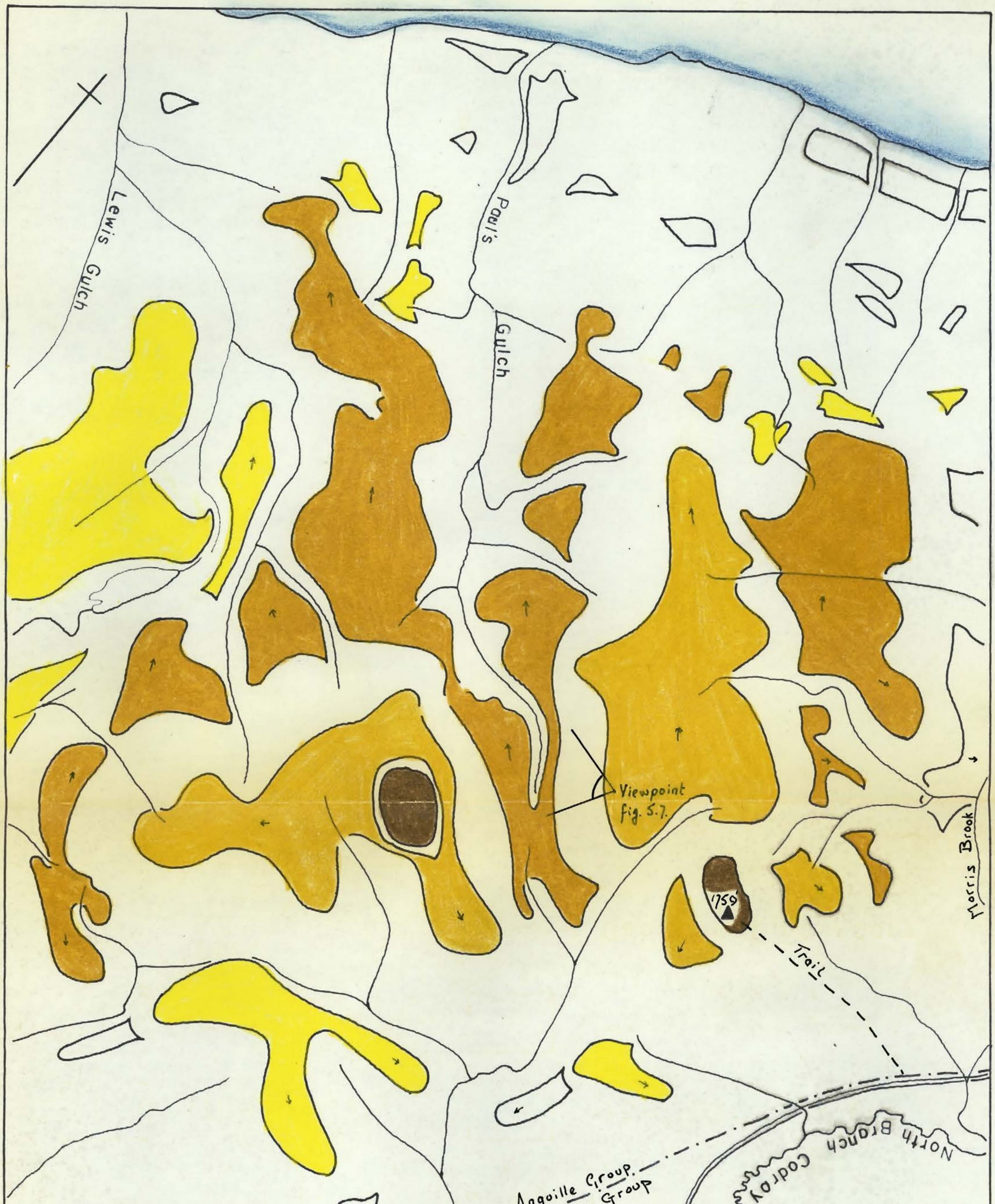


Fig. 5.6. The Erosion Levels of the Pauls Gulch area, Cape Anguille Mountains

1750+' level

1350' - 1500' level

1550'- 1700' level

1200'- 1300' level

Geological boundary

just described are seen. Mention has been made of a summit surface atop the 1750-foot summit across Paul's Gulch and of a 1600-1700 foot level sloping southeast, southwest and west from this high level remnant. To the northwest, the spur projecting from the 1750-foot summit bears another very marked and extensive flattening at between 1500' and 1400', over a distance of two miles. This interfluve level is correlative with valley-side spurs on the northeast side of Paul's Gulch, which pass up-valley into straths on either side of the valley and finally into the broad upland valley floor at the head of the Gulch (see Fig. 5.6 and 5.7). The present stream has cut a small gorge into this valley floor as a result of recent rejuvenation. Thus, in this area, the modes of occurrence of this level are close to the classical examples: interfluve level pass up-valley into valley-side spurs and then into valley straths and finally into an upland valley floor.

Remnants of a lower erosion level are seen well developed on spurs to the west of this major spur at between 1200' and 1300' (see fig. 5.6) and also on spurs along the eastern front of the Cape Anguille Mountains.

(iv). The Northeastern Area.

The next field area is the northeasternmost section of the mountains, between the through valley and the edge of the mountains overlooking the valley of Highlands River (fig. 5.5). The northeastern rim presents a steep front similar to that in the southwest, conforming to the lithological boundary between the Anguille Group and the Codroy Group. The Anguille Group in this area are coarser grained than in the southwest. Conglomeratic facies outcrop in steep freefaces crowning the northern slopes of the mountains.

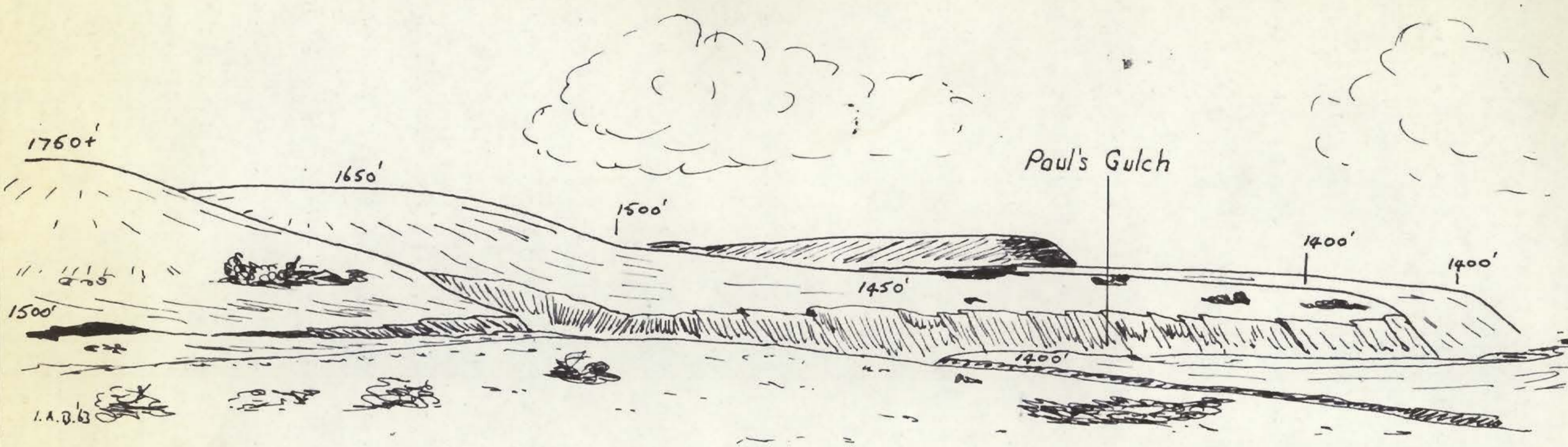


Fig. 5·7. Paul's Gulch from the northeast, 1400-1500-foot surface on interfluve, strath & valley-floor

The conglomerate has phenoclasts of rounded milky and pink quartz pebbles set in a matrix of coarse rounded and subrounded quartz sand grains. These rocks dip to the southeast at high angles.

The summits of this part of the mountains is seen in two broadly rounded hills, one at above 1550' and the other at above 1500', lying respectively to east and west of a flat-floored col at 1400' (see fig. 5.5). The surface of these two hills is covered with the dark-hued dwarf conifer-juniper-scrub vegetation, reflecting good drainage compared to the boggy upland valley floor between them (see fig. 5.8). A small pond lies in the floor of the col and streams drain from it to the north and south. That flowing south does so across a gently sloping erosion surface at 1400-1250' before entering a deeply incised valley which leads the brook through the eastern side of the mountains to join a stream flowing northwards into Highlands River. The 1400-foot level seen in the col and on either side of the latter brook is also found bevelling a spur projecting northwards from the 1550-foot summit on the latter's eastern flank (see fig. 5.9) and wrapping around the flanks of the 1500-foot summit. Towards the southwest, across the valley leading south from the pond, two spurs with flattenings upon them at 1300-1450', project southeast and northwest from a gently rounded summit area which rises above 1450'. The spurs themselves are part of the 1300-1450 foot erosion surface (seen in the col and flanking the summits on each side of it), and it is possible that the 1450-foot summit itself is a part of the same level. Distortions of true perspective are the cause of this uncertainty. On the map of erosion levels in this area (fig. 5.5) the summit itself is represented as a part of a higher erosion surface, that

Figure 5.8. 1400' surface in floor of col, with remnant of 1550-1700' surface in distance, northeast Cape Anguille Mountains looking west.

Figure 5.9. 1400' surface on spurs, northeast Cape Anguille Mountains. Long Range Mountains in distance.

seen atop the two isolated summits to the northeast. On the eastern side of the 1550-foot summit in the northeast occurs a narrow spur flat at 1200-1300' which is equivalent to a broader flat on the western side of the mountains at 1150-1250' (see fig. 5.5).

Thus, from this description of the scenery of the northeastern end of the Cape Anguille Mountains, it appears that there is a summit erosion surface at 1500-1550' atop isolated hills. Below it is another level at 1300-1450' on the flanks of these hills, in the cols between them and either over the whole of a broad interfluve to the southwest or at least on two spurs projecting northwest and southeast from a low remnant of the summit level. Below this, another level at 1150-1300' occurs on two spurs on the east and west sides of the northeast end of the mountains.

(v). Summary.

In the Cape Anguille Mountains, erosion surfaces have been recognized at the following elevations:- 1750' plus, 1550-1700', 1350-1500', 1150-1300', 1000-1100', 850-950', 750-850', 500-600'. All of these surfaces appear to be of subaerial origin and are undeformed. The position that the upper surfaces occupy in the landscape changes to both northeast and southwest of the highest area above Codroy Pond. In that area the 1750' - plus surface occurs on only two summits and is surrounded on three sides by the 1550-1700' surface which is well developed on northwest sloping spurs. To the northeast and southwest the 1550-1700' surface occupies a summit position above the 1350-1500'surface which is found well-developed on spurs sloping mainly northwest, but also to the southeast and towards the major valleys. This surface also occupies a spur and valley-side bench position in the central area below the remnants of the 1550-1700' surface. In the far southwest the 1350-1500' surface is seen atop two isolated summits overlooking Codroy village, above spurs belonging to the lower 1150-1300' surface. This latter surface is also found on spurs and valley-side benches below that at 1350-1500' throughout the mountains. The 1150-1300' nowhere occurs on summits. In the southwest, where it forms a prominent element in the landscape, it lies above well-developed spurs levels belonging to the 1000-1100' surface.

Little can be said about drainage evolution during the earlier stages represented by the upper most surfaces, since so little evidence is found of the direction and slope of the drainage which produced them. The solution of the problem of the origin of the through-valleys which cross the Cape Anguille Mountains is linked with the broader problem of the drainage evolution throughout the mountains. No solutions to either problem can be offered here. All that can be said is that there is little evidence of easterly-flowing drainage during the upper two stages - 1750'-plus and 1550-1700' - since all but one remnant of those surfaces slope to the northwest. This absence could possibly be due to subequent destruction of the traces of such drainage. Certainly, drainage southeastwards to a proto-Codroy at the earliest stages seems likely, if only for the reason that the non-existence of such drainage at those times is difficult to explain.

Drainage upon the upper surfaces appears to have been well-adjusted to differences in lithology within the Anguille Group, since the present pattern is markedly rectangular, in conformity with the strike of the rocks. Successive rejuvenations have caused the streams to extend their valleys into the heart of the mountains, becoming increasingly well-adjusted to minor flexures and faults in the process.

Erosion surfaces on the western side of the Cape Anguille Mountains have lower longitudinal gradients than those on the eastern side. This is due to the fact that westerly-flowing streams were longer, since they flowed to a base-level far out in the Gulf of St. Lawrence. Those dissecting the eastern edge of the mountains flowed to the Codroy depression, a much shorter distance. The existence of a major drainage line across the floor of the Gulf of St. Lawrence at an earlier period of lower sea level is very well evidenced by the present submarine topography of the area. The course of this ancestral "Laurentian River" may have been directed by a southeastward-curving projection of the thrust fault, known as "Logan's Line", which marks the eastern front of the Appalachian Mountains of eastern Canada. MacNeil (1956) has shown this fault trending southeastwards across the Gulf of St. Lawrence, and he tentatively projects it through Cabot Strait.

Figure 6.1. The 1200-1300' surface bevelling the southern rim of the Long Range Mountains. Looking north from near Cape Ray. Cook Stone at left.

Chapter Six: The Long Range Mountains

The Long Range Mountains of western Newfoundland comprise two geological units: the Lower Palaeozoic igneous and metamorphic complex, east of the reverse fault from near Cape Ray in the far southwest of the island to the shores of White Bay, and the Precambrian granite and granite-gneiss area of the Northern Peninsula. Field study of the latter area was not possible due to inaccessibility, but the former area was studied in several places along the western margin. The interior of these mountains are largely inaccessible to a solitary worker beset by logistical difficulties.

(i) The Table Mountain Area, Port aux Basques.

The first view gained of Newfoundland, when approached from Cape Breton Island via the ferry, is that of the southern end of the Long Range Mountains, whose level tableland surface is abruptly cut off by a steep wall at about two miles from the coast. This view typifies the scenery of western Newfoundland. This level tableland surface at the rim of the mountains is at 1250-1300 feet (see fig. 6.1) and access to it can be gained by way of a track following the eastern side of a steep ravine whose mouth is situated opposite the twin paps of Sugar Loaf (1005'). From the disused radar station at Cook Stone (see fig. 6.2), it can be seen that this level penetrates westwards towards the Long Range front in the form of a broad, shallow upland valley, between spurs belonging toa higher level at 1350-1450 feet (fig. 6.2), upon which Cook Stone itself is situated. Similar relationships are seen on the eastern side of the ravine. In the far southwest of the Long Range the summits of the higher level at above 1450' are cut off by the precipitous wall

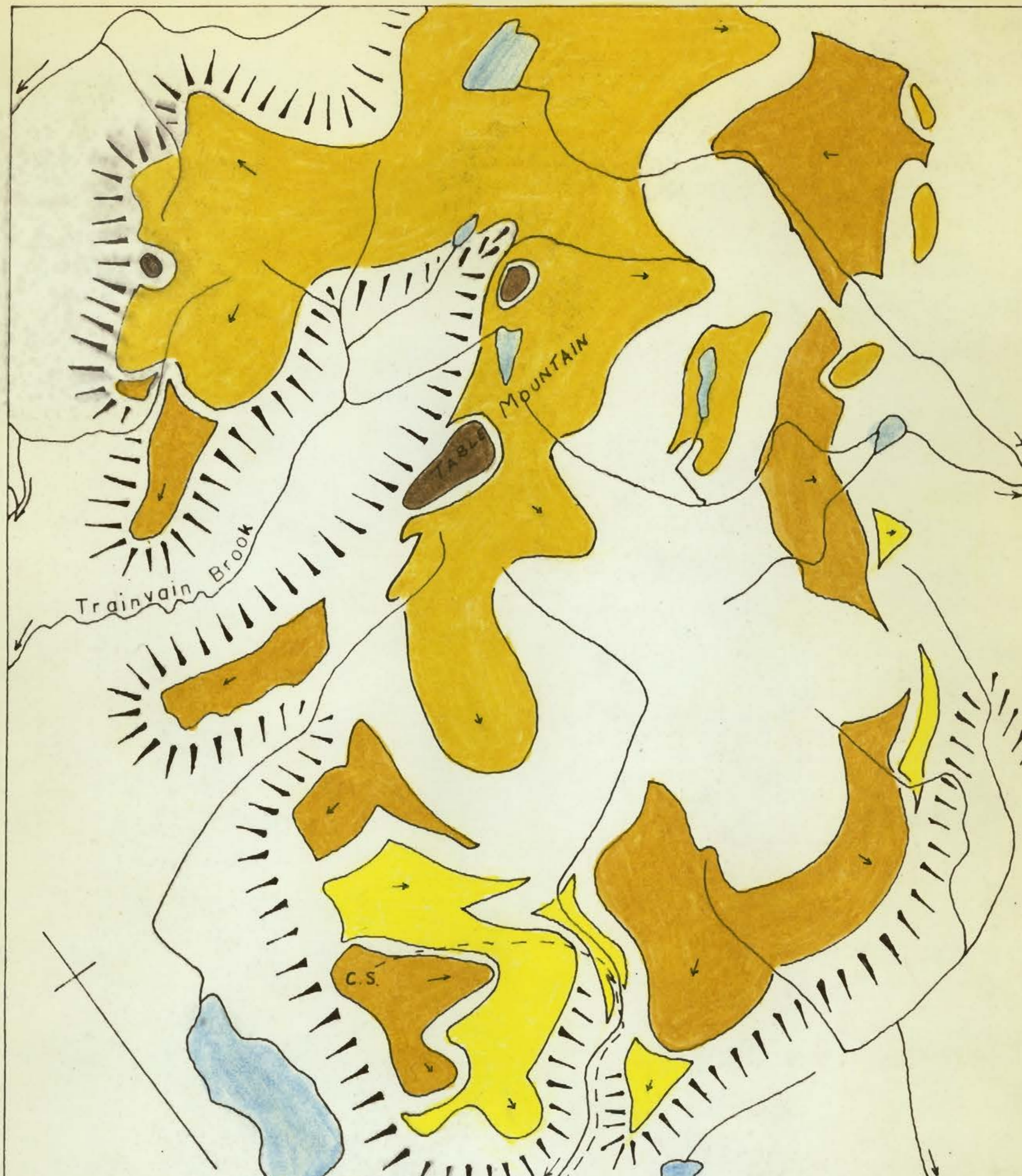


Fig. 6.2. The Erosion Levels of the S.W. Long Range

legend.

1750-2000'

1350-1500'

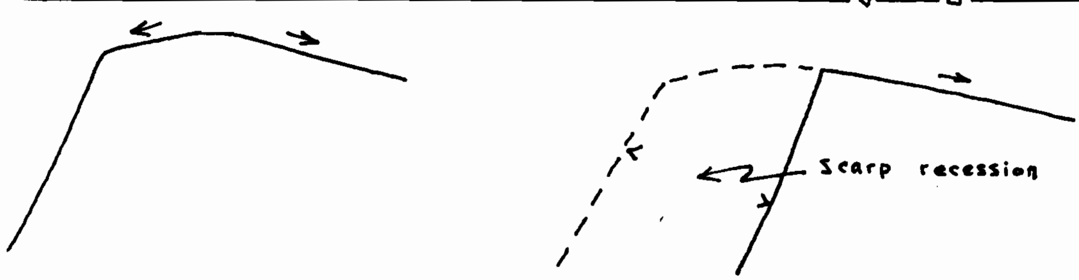
1550-1700'

1200-1300'

0 1/2 1 2 miles

of the scarp and the surface slopes southeast with the upland valley below it. This has also been beheaded by recession of the scarp face which has given rise to a shallow saddle in the crest line of the scarp. Across this saddle, towards the north, another remnant of the higher level is found at above 1550', smoothly sloping east into a small north-south col between it and the higher ground of Table Mountain. The general picture of the back slopes of these summits can be seen in the photo. (fig. 6.3). The erosion surface atop them is seen again towards the north on spurs projecting westwards from the Table Mountain mass and the higher areas to the north of it (fig. 6.2). The western extremities of these spurs have a slope towards the Long Range Scarp which truncates them, indicating that scarp recession has not proceeded as far as it has further south in eliminating these westerly sloping parts of the spurs. This also implies that they were cut by streams flowing west off the Long Range. At the time of formation of the higher level, drainage must have been to east and west from a divide near the present western edge of the Mountains (see figs. 6.4 & 6.5).

Fig. 6.4. Production of an eastward slope on Long Range spur.



East of the ravine dissecting the southern rim of the Mountains the 1350-1500 foot level is more extensive, forming a broad level on the long spur which projects south from the Table Mountain mass. It also continues around the southern rim of the mountains (see fig. 6.2). Slopes in this area east of the ravine are very gentle, making it difficult to discern

Figure 6.3. The 1350-1500' surface sloping eastwards on back slopes of spurs above Long Range Scarp (right). Degenerate "felsenmeer" of angular fragments of acid igneous rock in foreground. View south.

significant slope breaks. It is only by comparison with adjacent areas that a coherent pattern emerges. Thus, at the back of the 1150-1350 foot erosion surface, a break of slope marks it off from the 1400-1500-foot level. This slope break (fig. 6.6) is more marked than any in the area and may be due to structural factors.

From the broadly rounded, tor-crowned summit of Table Mountain at 1900' there project several spurs with level areas at 1600-1750', the most extensive of which occurs west of the stream draining the southern slopes of the mass. On its western side it rises above remnants of the lower level at 1400-1500', while on the eastern side of the mass smaller remnants slope southeast with the drainage. Table Mountain itself appears to be composed of rocks more basic in composition than the types seen in the scarp face. The summit area is a broad erosion surface at 1800-1900' which is separated from the lower level just described by a bluff 50-100' high. On the north of the mass the precipitous walls of Trainvain Canyon (fig. 6.7) rise almost to the summit. The summit surface is matched in areas to the northeast and north, where it becomes more extensive, particularly south of Little Codroy Pond.

The Long Range Mountains, between the southern rim and Little Codroy Pond, are surmounted by a summit level which appears at 1800-2100'. In places it appears as two levels. The main drainage lines trend northwest and southeast from this, the former to join Little Codroy River or to flow directly to Cabot Strait, and the latter tributary to southwest-flowing drainage elements which dominate the lower ground to the east of the mountains.

Figure 6.5. 1350-1500' surface on spurs above Long Range scarp. Looking north across Trainvain Canyon. 1800-2100' summit surface in distance (right).

Figure 6.6. Break of slope between 1350-1500' and 1550-1700' surfaces; possibly structural. Looking east from Long Range front across valley on 1200-1300' surface.

This lower terrain is dominated by N.W.-S.E. and N.E.-S.W. structural lines (see end plate 4), either foliations in the gneisses and/or faults in them and other igneous and metamorphic rocks. It is striking how streams flowing southeast from the Long Range, in which structural control is not readily apparent, make a sharp turn to join southwest-flowing streams, which are strongly influenced by structure. The slope of the land surface east of the highest areas of the Long Range changes from east and southeast on the eastern side of the high summits, to south, where the structurally controlled drainage pattern is dominant. This implies that this control has been important throughout the development of the land surface east of the Long Range, and has been reinforced of late in response to rejuvenation of the drainage.

While drainage to the southeast at an early stage seems well evidenced, drainage to the south and west seems equally well in evidence in the remnants of erosion surfaces at 1600-1750' and 1350-1500', which slope radially outwards from the Table Mountain mass. The high degree of development of these levels suggests that the steep scarp of the Long Range was absent from the landscape at those times. Thus, as with the streams draining the northwestern slopes of the Cape Anguille Mountains, the rivers which cut these levels must have been graded to base levels far from that of present streams. This base level must have been in Cabot Strait in the case of the rivers in the extreme southwest, but towards the north these rivers must either have joined a proto-Codroy flowing southwest or flowed northwest to join rivers flowing to St. George's Bay. The question of drainage evolution will arise again later, after discussion of the Long Range further north.

Figure 6.7. Looking west down Trainvain Canyon from slopes of Table Mountain. Codroy estuary in distance (right).

Figure 6.8. Looking south along Long Range front above South Branch, showing 1300' surface on spurs above Stephens' Brook and cuesta-like summits of 1900' surface.

(ii) The Long Range Above South Branch Codroy Valley.

This field area lies in the Long Range above the village of South Branch which is situated near the confluence of North and South branches of the Codroy River. Access to the mountains is gained along hunting and logging trails which are cut through thick woodland covering the Long Range scarp. From the vantage point of the scarp crest above South Branch (fig.6.8), looking along the Long Range front, several flats can be seen bevelling the spurs and the plateau tops to the southwest and northeast. A narrow flat at 1250-1400' bevels the front of all spurs projecting west from the mountain front, and is also seen as valley side flats and valley head benches in the valleys dissecting the front. The photo (fig. 6.8) is taken from a bevel on a spur and a complementary feature occurs across the deeply incised valley. The equivalent benches can be seen on the side of the valley and, as the valley widens upstream, the valley floor itself remains preserved as part of the same erosion level. Similar relationships are seen in the valley to the northeast of this which runs from the amphitheatre-like feature below the 1857' summit. The spurs on each side of this valley bear remnants of the 1250-1400' surface while the great amphitheatre of the upland section of the valley bears beautifully preserved benches at two levels, the lowest of which is equivalent to this spur level (fig. 6.9). Less extensive parts of a lower level at 1050-1150' can be seen bevelling certain spurs along this section of the Long Range front near Little Codroy Pond.

Although the area inland from the highest summits was seen only from a distance, it is evident from field and map observation that the 1250-1400' surface is widespread in that area. The Little Codroy, Isle aux Morts and South Branch Codroy rivers all head on a level, lake-dotted upland at

1300-1400', the valleys being set into this at 1250' (see end plate 3). The area around these headwaters appears to have been a well-developed pre-glacial valley, because of its wide open form and broad dimensions, and also because of the alignment of the headwaters of South Branch with that south-flowing stretch of Isle aux Morts River. The present divide between south-coast and Codroy drainage wanders indeterminately across this valley. That such drainage as is represented by this old valley was established towards the southeast, east of the highest summits of the Long Range agrees well with the conclusions reached in the area further southwest.

So far on the Long Range front, levels at 1050-1150' and 1250-1400' have been identified. From the highest summit remnants along these mountains there project spurs between shallow upland valleys, which are bevelled at between 1550-1700'. Particularly well seen are those developed around the 1857' summit and its counterpart to the northeast (fig. 6.10). Most of the upland valley floors between the highest summits belong to this erosion surface and these valleys are continuous with those in which signs of the next lowest erosion surface are found. This suggests that easterly flowing drainage has been established at least since the 1550-1700' stage.

The summit areas are of relatively small extent in this area, unlike that south of Little Codroy Pond. Looking southwest along the mountains, cuesta-like hills with a gentle east slope and a steep west slope are set back about one mile from the scarp face and bear remnants of an erosion surface at 1850-1950' (fig. 6.8 left). The form of these hills suggests structural control, and possibly they are composed of minor thrust slices

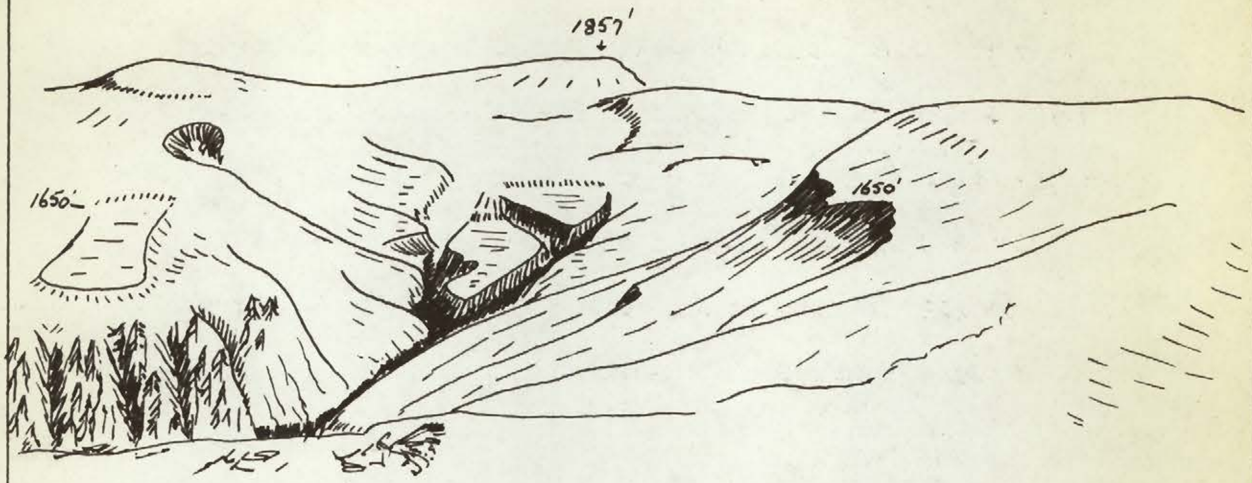


Figure 6.9. High-level valley benches at the head of a ravine cut into the Long Range scarp near South Branch.

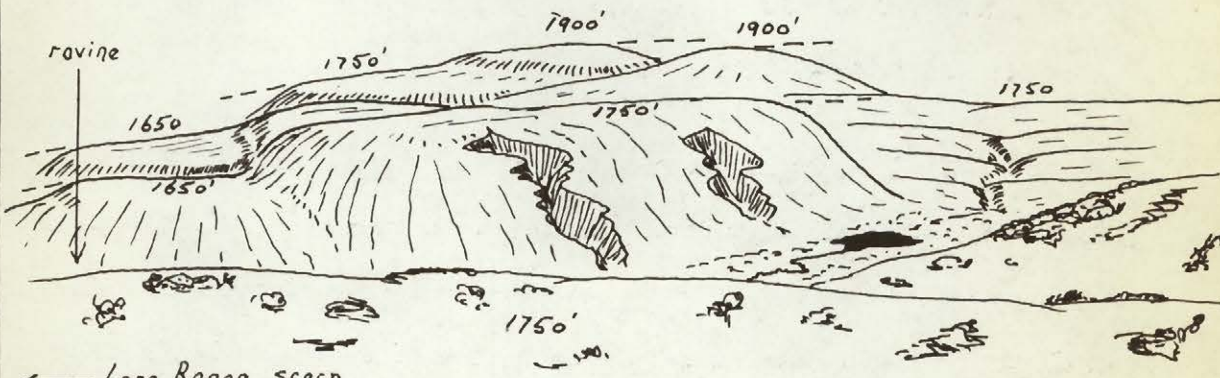


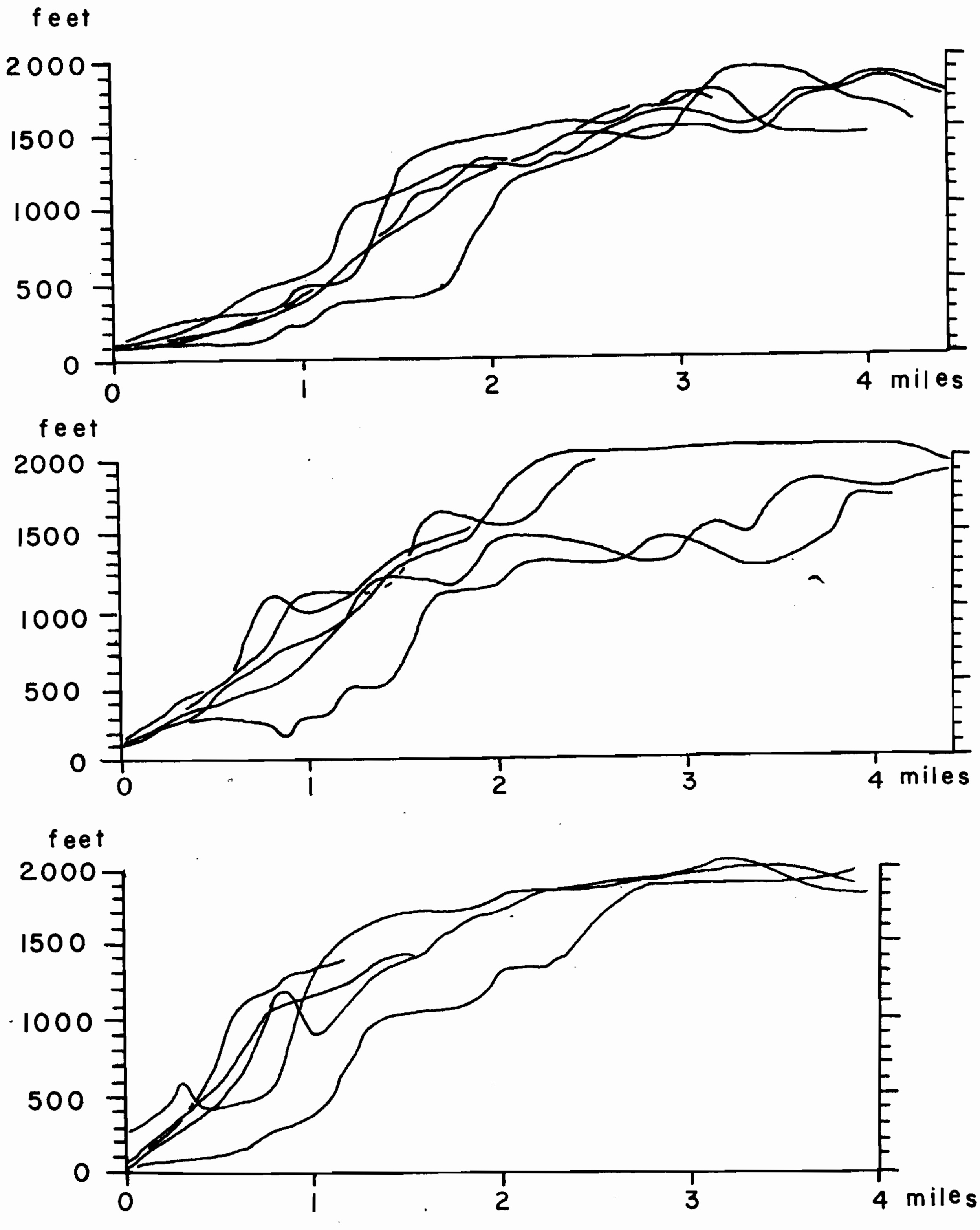
Figure 6.10. Relationships of erosion surfaces around the head of a ravine cut into the Long Range scarp.

moved towards the west during the Taconic orogeny. The summit at 1857' is another such remnant, as is the summit at above 1900' to the northeast.

(iii) The Development of the Codroy Depression.

In the sections of the Long Range so far discussed several well developed erosion surfaces exist. These are seen on the spur profiles constructed from the 1:50,000 maps (fig. 6.11). The lowest level, at 1050-1150', is matched by a similar level in the Cape Anguille Mountains, bearing similar relationships to other surfaces. The next highest surface is that at 1250-1400', found on spurs and valley sides. Figure 6.12 shows this surface on spurs of the Long Range front above South Branch. West of the height of land this level, like the one below it, slopes westwards with the drainage, but to the east where it is better developed, it has an easterly or southeasterly slope. Remnants of a similar level in the Cape Anguille Mountains slope predominantly to the northwest, but there are significant remnants which slope towards the Codroy depression. This implies that drainage southwestwards along this depression was established at least as early as the 1250-1400' stage. In the Long Range a surface at 1550-1700' has a slope west from the highest summits and also to the east. That well marked levels appear at this height along the eastern edge of the Cape Anguille Mountains with a dominantly northwest slope has previously led to the conclusion that the drainage system responsible for the formation of this surface flowed northwestwards from a divide east of the present edge of those mountains. The presence of a similar surface with the same slope in the Long Range has suggested the possibility of drainage being continuous from those mountains to the Cape Anguille Mountains at this atage. From a

Fig. 6.11. Longitudinal profiles of spurs of the southwestern Long Range



study of the relevant profiles (end plate 5) and geological structures, this possibility diminishes on several counts. Firstly, the distance from the remnants of this level on the Long Range northwest to those on the Cape Anguille Mountains is approximately ten miles. If the slope of this surface on the Long Range is projected westwards for this distance, the surface is carried below the remnants on the Cape Anguille Mountains. Secondly, the Codroy depression is such a marked structural feature, bounded by lines of weakness initiated in the late Palaeozoic, that it is difficult to explain why it did not influence drainage trends, even at the time of formation of the upper erosion levels. Obviously, at those times the depression could not have been bounded by such steep walls. How, them, can the absence of remnants of the 1550-1700' surface along the eastern Cape Anguille Mountains be explained? Firstly, the absence is not total, since one easterly sloping remnant was noted in the highest parts of the mountains and can be accounted for by postulating an easterly swing in the trend of the drainage divide to cross the upland mass at the north-east end of the depression and thence to the Long Range. One part of the explanation may be that the development of the embayment in the east side of these mountains had proceeded so far that the surface was reduced below the general level of the 1550-1700' surface and has since been further lowered, thus eradicating all traces of it. Another contributory factor may have been the development of southwest-flowing streams parallel to the Codroy, aided by structural weaknesses, which caused the destruction of remnants of the surface and the recession of the divide which bore other remnants of it far back into the centre of the mountains during later stages.

Figure 6.12. Looking south along Long Range scarp above Codroy depression at South Branch, showing 1250-1400' surface on spurs.

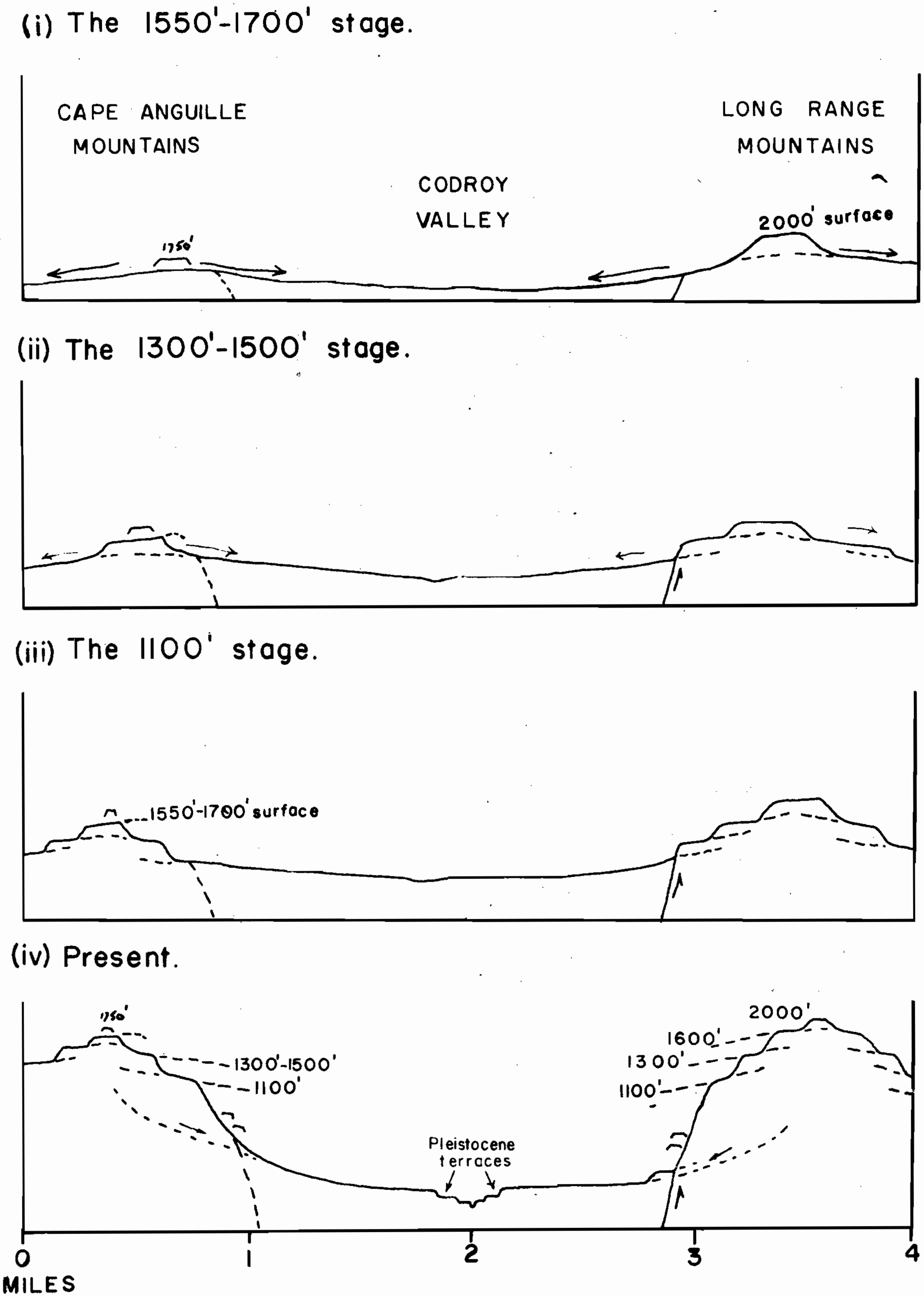
Also, the eastern front of the Cape Anguille Mountains may have receded west as the Long Range front appears to have done so to the east, under the attack of streams successively rejuvenated at later times, leading to the destruction of easterly sloping parts of the 1550-1700' surface. The evolution of the Codroy depression has been tentatively set out in fig. 6.13.

(iv). The Flat Bay Brook Area.

The Long Range Foreland is backed by the northeast trending Long Range scarp until the latter is crossed by the Valley of Fischell's Brook. Northeast of this valley the Long Range Mountains project westwards towards St. George's Bay in a salient of anorthosite rock. The edge of this mass trends first north-by-west along its unfaulted margin from the Long Range scarp, then north-by-east along the faulted margin which crosses Flat Bay Brook, and then northeast along the contact with the drift-covered area around the head of St. George's Bay to its contact with the Long Range igneous and metamorphic rocks near the valley of Bottom Brook.

Flat Bay Brook is the major stream crossing this anorthositic salient. Like other rivers of the St. George's Bay system, it rises far to the east of the Long Range mountains front, in this case some twenty-five miles, on a divide marked by a line of summits at about 1800-2000', rising 200-300' above the 1500-1600' surface of the High Central Plateau. From this divide streams belonging to the Lloyds and Victoria river systems also flow east and northeast to the Atlantic Ocean. For about ten miles from the divide Flat Bay Brook flows in a wide upland valley across the plateau, but then becomes incised to a depth of 500-700' for the remainder of its course across the mountains to the restricted Long Range Foreland,

Fig. 6.13 The Development of the Codroy Trough.



between the mountain front and St. George's Bay. The incised reaches of the river and its tributaries are much influenced by the structural control of faults and other weaknesses in the igneous and metamorphic rocks of the Long Range complex and in the anorthosite mass. The drainage pattern is markedly angulate as a result of this control. Many stream junctions are at an angle of about thirty degrees and many smaller streams directed by minor faultlines flow in the reverse direction to that of the major streams. The interfluves are extremely level in long-profile and bear extensive remnants of erosion surfaces far into the heart of the Long Range. The northeasterly prolongation of the Long Range faultline is not reflected in any physiographic discontinuity, since rocks on either side of it are of similar resistance to denudation. This fact has implications for the development of the Long Range faultline scarp farther southwest and will be discussed in the following section dealing with the St. George's Bay river system.

The erosion surfaces were investigated on the interfluves to the north and south of Flat Bay Brook above that section of the valley between the fault to the west and the confluence with Lookout Brook, near the hydro-electric power house (see fig. 6.15). The general appearance of the interfluve to the south is seen in fig. 6.14. On the southern interfluve which is higher than the northern, an erosion surface is extensively developed at elevations between 1050-1300', sloping northwestwards with the streams dissecting it. Between Three Brooks and Flat Bay Brook the surface rises westwards to the junction with a higher surface at above 1400', but this latter surface lies outside the area of field study. Remnants of the 1050-1300' surface are found on the northern interfluve only as summits in

Figure 6.14. Looking south across Flat Bay Brook (bottom) from Steel Mountain (see fig. 6.15), showing 1050-1300' surface on interfluve and lower valley benches. Mammillated topography reflects anorthosite bedrock.

the central parts and are only found surmounting interfluves further east. Thus, the western extremity of this surface lies along a northeast to southwest line reflecting the trend of the coastline in this area. Far more extensive over the northern interfluve is a lower erosion surface, at elevations between 900-1050'. This occurs in the same situation as the higher surface on the southern interfluve. In the area south of Flat Bay Brook it occurs only as high valley-side benches below the main interfluve surface (see fig. 6.13). The highest valley-side benches occurring north of the brook are parts of a locally developed erosion surface at 750-800'. Benches at this height are quite extensive, even in the eastern parts of the area shown in fig. 6.13, and the level summit of Cairn Mountain also belongs to this surface. On the south side of Flat Bay Brook valley benches at this height are generally poorly developed, the exception being a bench found west of the debouchment of Three Brooks from the southern interfluve. The only other recognisable erosion surface found in this area is at 500-600', bevelling the western extremities of the northern and southern interfluves. One remnant is found north of the main summit of Steel Mountain, and the other on a spur which is crossed by the fault-line which bounds the anorthosite mass to the west. These remnants are so flat that a marine origin was first thought possible. This conclusion cannot wholly be discounted since the only evidence to the contrary, the lack of other recognisable remnants in the vicinity, may be due to their destruction or modification by glacial erosion and deposition.

In the valley of Flat Bay Brook, incised 700-800' into the plateau, minor constructional terraces of glacio-fluvial sands and gravels occur on

Fig. 6.15. The Erosion Levels of the Flat Bay Brook area.

Legend.

1050 - 1300' level

950 - 1050' level

750 - 850' level

500 - 600' level

fault

solid - drift boundary

MILES

0 ½ 1 2

the valley sides. Below them the flood plain of the brook is up to half a mile wide, and the brook flows across it in wide irregular meanders with many shoals and braided reaches. Nowhere in this section is the river more than about six feet deep in summer.

(v) The Development of the St. George's Bay river system.

The rivers which flow west into St. George's Bay - Highlands River, Crabbes Brook, Barachois Brook, Robinson's River, Fischells Brook, Flat Bay Brook, Little Barachois, Southwest and Bottom Brooks - rise on a divide 8-25 miles east of the front of the Long Range Mountains. This divide (see end plate 7) trends southwest to northeast across the level, lake-dotted upland of the High Central Plateau at 1500-1600'. For long stretches it is indeterminate, sometimes passing through a lake from which drainage is to both west and east. In places it rises up on to a line of higher summits at 1800-2000', standing above the plateau which, as Twenhofel (1940) pointed out, represent outlying remnants of the 2000' summit plateau which surmounts the Long Range Mountains further west. From this divide, the rivers follow sub-parallel courses to St. George's Bay. The individual drainage basins are markedly elongate as a result and the broad pattern may be called pinnate, although, in the mountain section, the pattern is angulate.

Although the courses of the rivers may have been used as general guides in mapping the outcrops of the various rock units of the Long Range igneous and metamorphic complex, from Riley's (1962) map it appears that the boundary between the bodies of truly igneous granitic rocks and the more widespread outcrop of the gneissic rocks of the complex is a common situation for long reaches of the streams crossing the Long Range Mountains. The rivers cross the regional structural trend, but are much influenced by faults and minor structures, such as limeations in the gneiss and the trend of the gneissosity. In crossing the Long Range faultine no change of direction in the rivers' courses is seen. The exception is Fischell's Brook which, as it approaches the faultline bounding the Long Range complex proper and the anorthosite to the west, swings southwest, parallel to the faultline and then, on crossing it, swings northwest along the southern boundary of the anorthosite mass. Possibly, capture has effected the peculiarity of this river's course.

The rivers cross the Long Range faultline or thw estern boundary of the anorthosite mass on to the folded and faulted Carboniferous strata of the Foreland (see end plates 2 and 7) and entirely disregard these structures in their courses to St. George's Bay. They appear wholly to be consequent upon the surface of the Foreland, and it si through analysis of the relationships of erosion surfaces on either side of the Long Range faultline that the approach to the problem of accounting for the transection of the Foreland structures should be made.

The three generally accepted means by which drainage may come to transect pronounced geological structures are (a) cpatures induced by headward erosion (b) superimposition from a thickness of rock previously masking the structures, and (c) antecedence.

Acceptance of an explanation by capture must depend upon a favourable disposition of stream courses and geological structures over a wide area. In the Long Range Foreland the stream courses and the major interfluves all trend at right angles to the geological structures. Also, unless they have been masked by a cover of glacial drift, strike valleys are completely absent. These objections, as well as that raised by the great extent of the area of transverse drainage, rule out acceptance of river capture as an explanation of the transection of the Foreland structures.

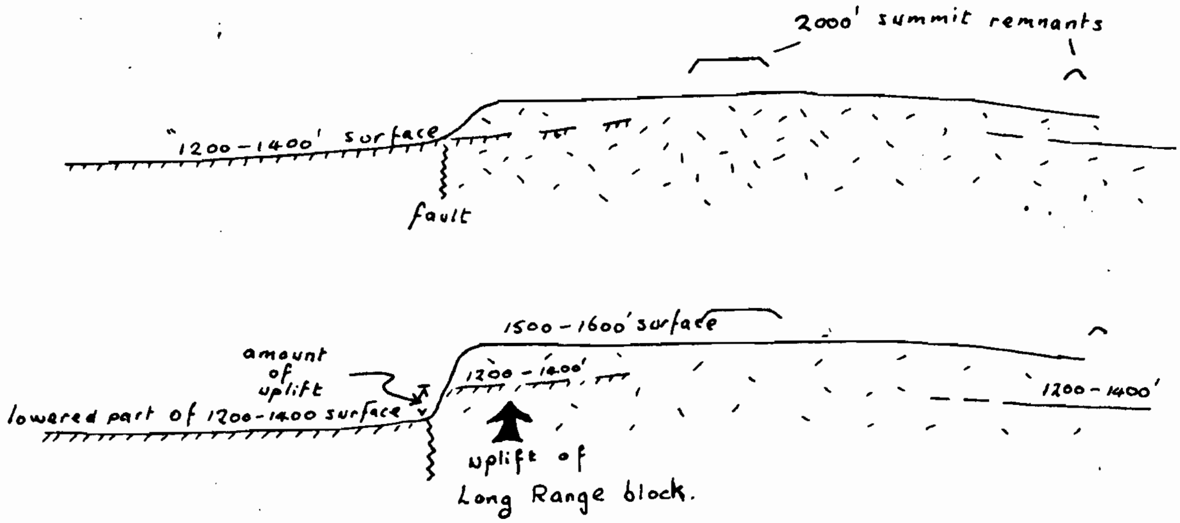
Arguments against accepting superimposition of the drainage from a cover rock are less conclusive. No evidence of rocks younger than those in which the structures are found has so far been revealed in the area. This is not a serious objection since, in the Piedmont province of the Middle Atlantic States, this absence has not hampered general acceptance of Johnson's (1931) theory of regional superimposition of Appalachian drainage. It is known that during the Permian and Triassic periods no significant marine transgressions occurred in northeastern North America, and it is unlikely that a cover rock of this age ever existed in this area. Similarly the Jurassic period does not appear to have been marked by transgressive episodes. Good reasons for believing that Cretaceous rocks never covered any part of Newfoundland were given in chapter three, and positive eustatic changes of sea level during the Cenozoic were not of sufficiently wide extent to cover present-day land areas of the Atlantic seaboard of northeastern North America. Therefore, the lack of practical and theoretical evidence of a cover rock young enough to have been stripped from the Long Range Foreland in relatively recent times militates against acceptance of superimposition of that drainage.

Antecedence remains as the third means by which transection of structures by drainage lines may arise. When a river or a whole drainage system is antecedent, this means that the drainage was in existence before the structures; it is antecedent to the structures. In the case of the Long Range Foreland drainage this imples that it is a pre-Carbo-Permian system since the structures were formed during the "Appalachian Revolution". This was also the period of earth movement in which the Long Range fault was initiated, according to Betz (1943). It is unlikely that any drainage system would have been able to maintain its course across these structures because they are so pronounced (the Long Range fault has a throw of approximately 3000'). Also, if the drainage has existed at least since the Appalachian Revolution, then it would certainly have reached an adjusted state by this time. No later date can be accepted for the movements against which the rivers held their courses, since this presupposes that the drainage was in existence across the Appalachian structures before these movements occurred. Thus, antecedence, in the normal sense, cannot be entertained as an explanation of the transection of Foreland structures.

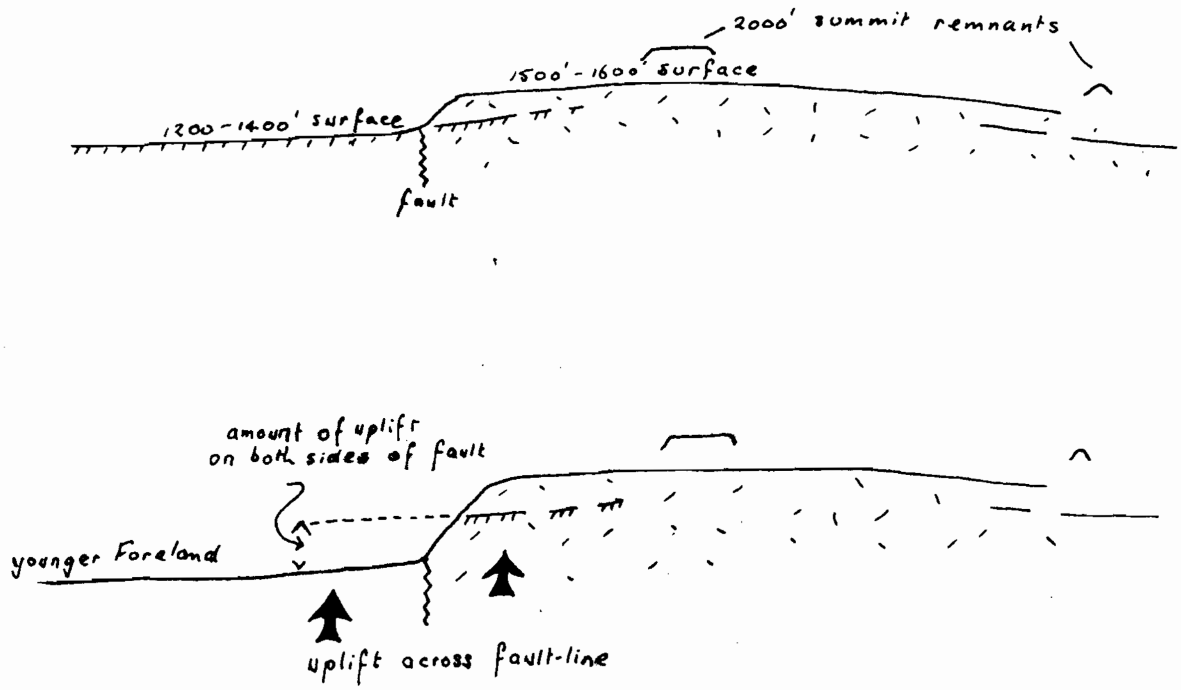
As previously stated, erosion surfaces are found in the Long Range Mountains east of the scarp at about 1800-2000', 1500-1600', 1300' and 1100'. The latter surface occurs only as minor benches on westerly-projecting spurs and in the valleys of the St. George's Bay river system in their mountain sections. The question arises as to whether earth movement in the area on each side of the fault has been of equal or unequal amount. In other words, is the Long Range scarp a faultline scarp or a regenerated fault scarp? If the relief differential between the Long Range Mountains and the Foreland is the result of upward movement of the Long Range block along the faultline relative to the Foreland, then the Foreland surface should be equivalent in

Figure 6.16. Two Possible Mechanisms of Evolution of the Long Range Scarp & Foreland

(a). Regenerated Fault Scarp.



(b). Fault-Line Scarp



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age to the youngest erosion surface in the Long Range formed before the uplift occurred. In this case, before the uplift, an erosion surface would have extended across the Foreland, transgressing the faultline into the harder rocks of the Long Range complex. The faultline would have been reflected in a scarp whose height would be equivalent to the difference in elevation between this surface and the one above it, approximately 200-300' (fig. 6.16a). Following movement along the faultine, the scarp would be at approximately its present height above the Foreland, 600 feet, implying that the uplift was about 300-400'.

Two factors hamper acceptance of the origin of the Long Range scarp as a feature regenerated by the process outlined above. Firstly, similar or greater relief differentials are found along the unfaulted contacts of the Anguille and Codroy groups around the nose of the Cape Anguille Mountains anticline and along the unfaulted boundary between the anorthosite and the Carboniferous rocks north of Fischell's Brook. Independent movement of the Cape Anguille Mountains anticline, at least, seems improbable, so that the relief differential can be accounted for wholly by differences in resistance to denudation of rocks on each side of the boundary. Secondly, as the Long Range fault is traced northeastwards, out of the area where it forms the boundary of the mountains into the area where it merely separates the Long Range complex from the anorthosite body to the west, it is not marked by any physiographic discontinuity. Erosion surfaces are developed evenly across it since there is little difference in the resistance of the rocks on either side. It, thus, seems unnecessary to invoke differential movement along the faultline to explain the Long Range scarp.

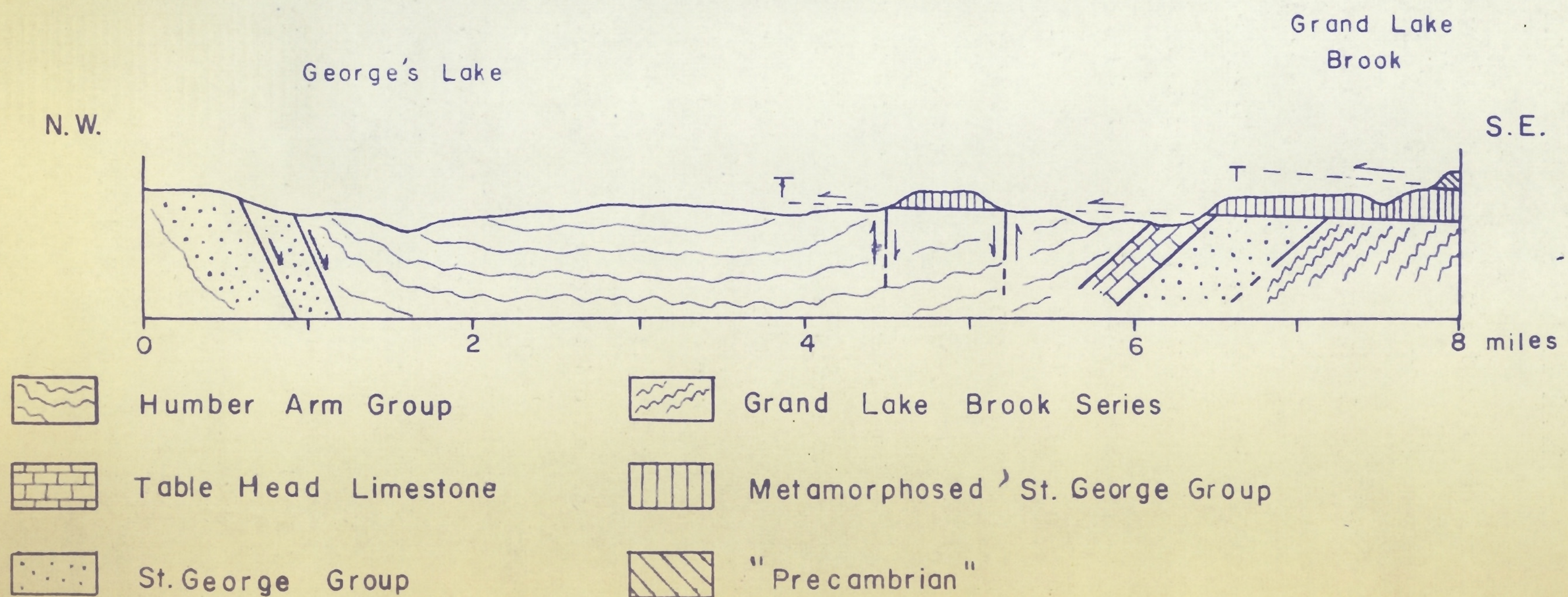
The alternative explanation is that uplift has occurred equally on either side of the faultline and the present scarp is merely a reflection of differences in resistance to denudation. If this is the case, then the Foreland surface is not related to a surface in the Long Range Mountains (see fig. 6.16b). Following an uplift over the whole area on both sides of the faultline of the same order as that postulated in the previous case, the Foreland would have been lowered to its present level relative to the mountain front by rivers whose erosive power was increased by the fall in base level. The superior resistance of Long Range rocks would ensure their maintenance at a high elevation. By this process the scarp would, thus, be a faultline scarp. However, regeneration of old faultlines during periods of earth movement is a well documented tectonic phenomenom and so one might ask why this did not occur in the case of the Long Range fault before the faultline scarp hypothesis is accepted.

Whatever the evolution of the Long Range scarp, the question of how the drainage across the Foreland attained and maintained its westerly direction across the structures in the Carboniferous rocks is still open. The distribution of the erosion surfaces at 1800-2000', 1500-1600' and 1300' in the Long Range Mountains and their slopes imply that drainage from these mountains has been to both east and west during their formation. It is possible that some outside factor(s) not connected with the present surface and drainage has contributed to the attainment and maintenance of westerly courses in the rivers. It is possible, though not likely, that structures in the Carboniferous rocks, which have been removed following successive uplifts of the area, were less pronounced than those seen in the present outcrops and that drainage was more easily established across them. Also, a steadily falling base level would have kept the rivers actively downcutting during the stripping of these rocks, enabling them to disregard structures of greater intensity when these were excavated. Neither of these explanations can be offered with any degree of certainty, however, and the solution of the problem must await more detailed work.

Chapter Seven: The Humber Arm - St. George's Bay Dissected Plateaux Area

The area bounded by St. George's Bay, Humber Arm, the Long Range front and the eastern edges of the plutonic bodies of the Lewis Hills and the Blow me Down Mountains (region VIa, end plate 1) is underlain by lower Palaeozoic rocks of two major types (see chapter 3, end plate 2), the carbonate and the clastic terranes. These rocks are believed by some to represent the unmetamorphosed equivalent of rocks found in the Long Range igneous and metamorphic complex, and by others to be in part allochthonous and to have found their way to their present positions through large scale horizontal displacements (Rodgers & Neale, 1963). Knowledge of the geology and structure of this area has resulted largely from the work of Walthier (1949) and the lithology and structure has recently been summarized by Riley (1962). Walthier recognized two thrusts in the area between the Blow me Down Mountains and Corner Brook; one carrying "Precambrian" (now thought to be lower Palaeozoic) gneisses westwards over Cambrian meta-sediments, and another, carrying Cambrian and Lower Ordovician sediments over the Middle and Upper Ordovician. Major folds were recognized west of these thrusts, and the region is broadly synclinal in structure. One of Walthier's sections west to east across the area (fig. 7.1) shows this structure. Crustal disturbance occurred during the Middle or Upper Ordovician, in the post-Ordovician - pre-Mississippian, and in the post-Pennsylvanian periods. East of the plutonic bodies of the Lewis Hills and the Blow me Down Mountains, the sedimentary rocks have been folded around these masses by pressure from the east, during movements which must thus have occurred after their emplacement. A large overthrust was recognized by Walthier, bounding the

Fig. 7.1. Geological Section across the George's Lake Area. (after Walthier, 1949)



Lewis Hills pluton to the west, and swinging eastwards around the southern end of the mass.

The most widespread erosion level between the Long Range and the plutonic bodies, and south of the latter, occurs at between 1100' and 1200'. It is best developed east and west of the central George's Lake depression, where the land surface drops below that level in lower erosion levels and drift-covered lowlands. It is also well seen on the carbonate rocks of the Port au Port Peninsula in the White Hills, upon Table Mountain, Port au Port on similar rocks, and on other areas of Ordovician carbonate rocks, such as that between Romaine and Blanche brooks, north of Phillips Brook and on clastic terrane fringing the plutonic bodies. The occurrence of this level, therefore, is not related to lithology or structure in the Lower Palaeozoic rocks. The surface was studied in two areas: Table Mountain, Port au Port and the area west of George's Lake, in the valley of East North Brook.

Table Mountain is an anticlinal structure in Ordovician carbonate rocks, aligned slightly east of north, extending for a distance of ten miles from above the settlement of Port au Port to the high ground overlooking the lower valley of Fox Island River. The steep, western face of the mountain is a structural slope formed by the steep limb of a monocline. The limestone beds are seen sharply bending over on the brow (fig. 7.2), from an almost horizontal attitude which is reflected in the level surface of the mountain, to a dip of about 60 degrees. The whole surface of the mountain is not structurally controlled, however, since the rocks further

Figure 7.2. Monocline in St. George Group limestone, west face of Table Mountain, Port au Port.

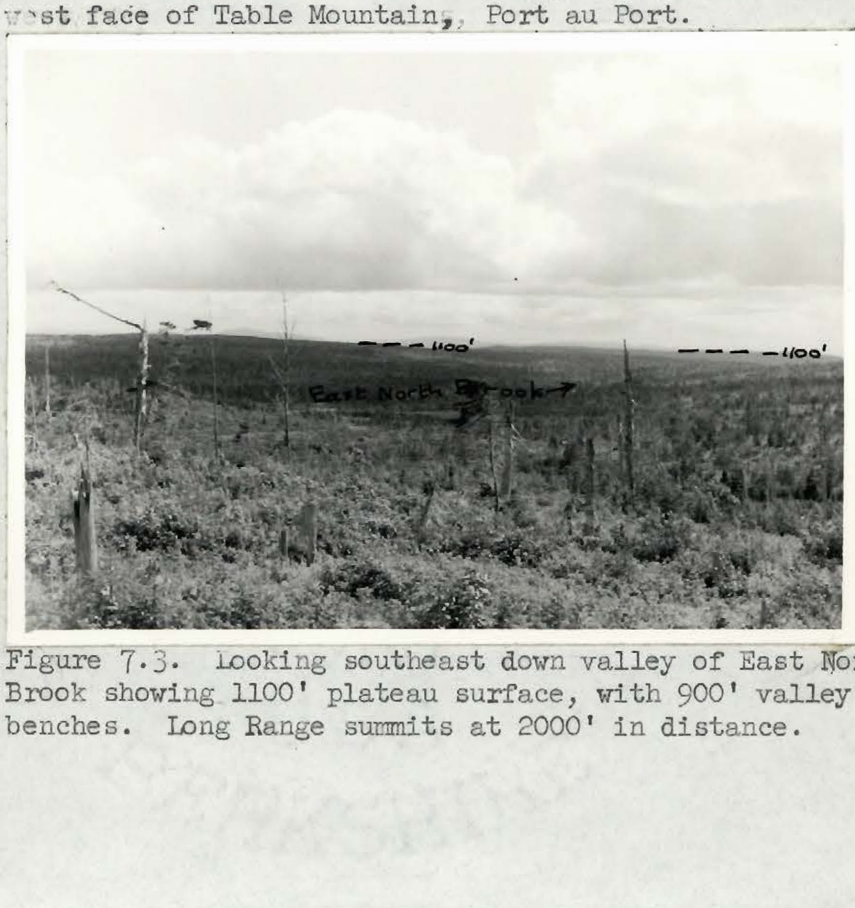


Figure 7.3. Looking southeast down valley of East North Brook showing 1100' plateau surface, with 900' valley benches. Long Range summits at 2000' in distance.

east of the western edge are folded into an anticline towards the northern part and a syncline in the southern part. The surface of Table Mountain is remarkably level between 1100- and 1200'. Only in the south and north does the land rise above this level as gently rounded monadnocks on this surface. From the vantage point of the Table Mountain summit the view eastwards shows other areas conforming to the general level at 1100-1200', such as the northeast to southwest-trending limestone ridges between Table Mountain and the Indian Head Range. This 1100-1200' erosion surface fringes the Indian Head Range and rises up into it along the valleys penetrating the upland, so that its back edge may stand at 1350'.

The Indian Head Range comprises a north to south trending ridge of Precambrian acid intrusive rocks, extending from the head of St. George's Bay some sixteen miles northwards into the heart of the lower Palaeozoic sedimentary area (end plate 1). By virtue of the superior resistance to denudation of the intrusive rocks of the Range, the surface rises to above 1800'. Summits which reach this elevation lie above a well-developed flatter area at about 1500'-1600', which displays all the characteristics of a subaerial erosion surface.

The other area examined in this wide outcrop of Lower Palaeozoic rocks is that northwest of the settlement of Gallants, along and above the valley of East North Brook. Access is gained by private roads belonging to the Bowaters Pulp and Paper Corporation. The lower two to three miles of East North Brook, above its confluence with Harrys River, flows across a smooth, flat, flood-plain developed upon drift deposits, associated with

Fig.7.3. The Erosion Levels of the East North Brook area

Legend.

1350'

1150'-1250'

1050'-1150'

850'-1000'

(?) 800'

750'

Drift border

Esker delta & esker

anticlinal axis

synclinal axis

0 ½ 1 2

MILES

the main area of drift deposition in the Harrys River lowland. Upstream from its confluence with West North Brook, East North Brook flows in a valley incised into the St. George Group limestones, which are of exactly similar nature to those seen on the Port au Port Peninsula and on Table Mountain. This valley is steep-walled and is incised 500' into the general plateau level. The margins of the plateau around the Harrys River lowland are at a maximum elevation of 1000-1100' and a partial peneplain or strath-like flat can be reconstructed from remnants of this level found on spurs and valley-side benches (fig. 7.3). Parts of the major valleys, as yet unaffected by recent rejuvenation, bear this level down to 850' as local relief on this surface. On the broad plateau area between East and West North Brooks, this 1000-1100' erosion surface lies below a higher and better developed surface at 1150-1250' on the interfluves. The junction between these two surfaces is well seen looking down the valley of East North Brook, above the right-angled bend near the boundary of Humber West and St. George's administrative districts (fig. 7.4). Figure (7.3) shows the extent of the erosion surfaces in the area, their relationships to the drainage and the relationship of both to the structure.

Speculation on the evolution of the drainage pattern in the area of the Lower Palaeozoic outcrop is unwarranted when so little detailed field work was possible. However, a few observations are noteworthy. The major drainage lines, such as Harrys River - George's Lake, Serpentine River, Fox Island River, Romaines Brook, Cooks Brook, all appear to be structurally guided (see end plate 7). At the time of formation of the highest erosion surface, that at above 1350' found on the major divide crossing the area west to east in its central parts, drainage would have been finely adjusted to structures, and would thus have flowed predominantly northeast and southwest. The central depression probably was occupied by a stream from earliest times and this may have had its source far outside this area. The alignment of the depression with the valley of the lower Humber River (end plate 7) suggests a possible source at an early date. The Serpentine trench must certainly have been occupied by a major drainage element since earliest times, but the question as to the direction of this element is difficult of solution on so little morphological evidence. At the time of formation of the highest surface in the sedimentary area the drainage divide must have crossed the area roughly east to west, from the Lewis Hills to the Long Range, and it seems possible that the Serpentine River would have established its present direction by that time. Northward drainage to Humber Arm must also have been established, since remnants of the highest surface at above 1350' also slope in that direction. Thus, the initial drainage of the area was probably northeast to southwest along the line of the central depression and may have had its source in the present line of the lower Humber River. Subsequent evolution involved adjustment to structures in the Palaeozoic rocks, particularly the Humber Arm line of weakness and the Serpentine Trough. This adjustment probably was enhanced by the vigour of streams flowing west and north. Since uplift along the major fault line of the west coast considerably shortened these streams. More will be said of drainage evolution in this area in chapter eight.

Regarding the correlation of these erosion surfaces in the Lower Palaeozoic sedimentary lowland with those of nearby regions, it seem likely that the surfaces at 1100-1200' is the same as that surface so well developed over the interfluve south of Flat Bay Brook at comparable elevations (fig. 6.15). The lower levels in the Ordovician sedimentary area (see fig. 7.3) would then be correlative with those seen on the lower parts of that same interfluve and as valley-side benches in Flat Bay Brook.

Chapter Eight: The Bay of Islands Mountains

The Bay of Islands Mountains comprise four roughly circular mountain areas, developed upon the plutonic mafic and ultramafic bodies of the Bay of Islands Igneous Compex (Smith, 1958) (end plates 1 and 2). These mountains, together with those of the southern part of the Great Northern Peninsula, are the highest in Newfoundland, and the highest point in the island is found in the southernmost of the masses, the Lewis Hills, at 2673'. The most detailed work on the geology of this complex is that by Smith (1958), whose findings are summarized in chapter three. The general physiography of the four mountain masses, the Lewis Hills, the Blow me Down Mountains, Mt. St. Gregory-North Arm Mountain Mass and Table Mountain, Bonne Bay has been described in chapters one to four.

The mountains present a steep scarp on all sides (end plate 3) due to differences in resistance to denudation of the igneous and sedimentary rocks which are in unconformable and faulted contact around the mountain edge. Surmounting this scarp is a remarkably even surface at about 2000', which was considered by Twenhofel (1940) to represent the general level of the surface of these mountains. This was because views obtained of them, either from a distance on the ground or from an airplane, did not show the true relief of the mountains' surface. When the surface is viewed from the interior of the mountains it soon becomes apparent that the earlier impression was an oversimplication.

The mountains were examined in the following localities: -

(i) Between the mouth of the Serpentine River and the 2673' summit of the Lewis Hills.

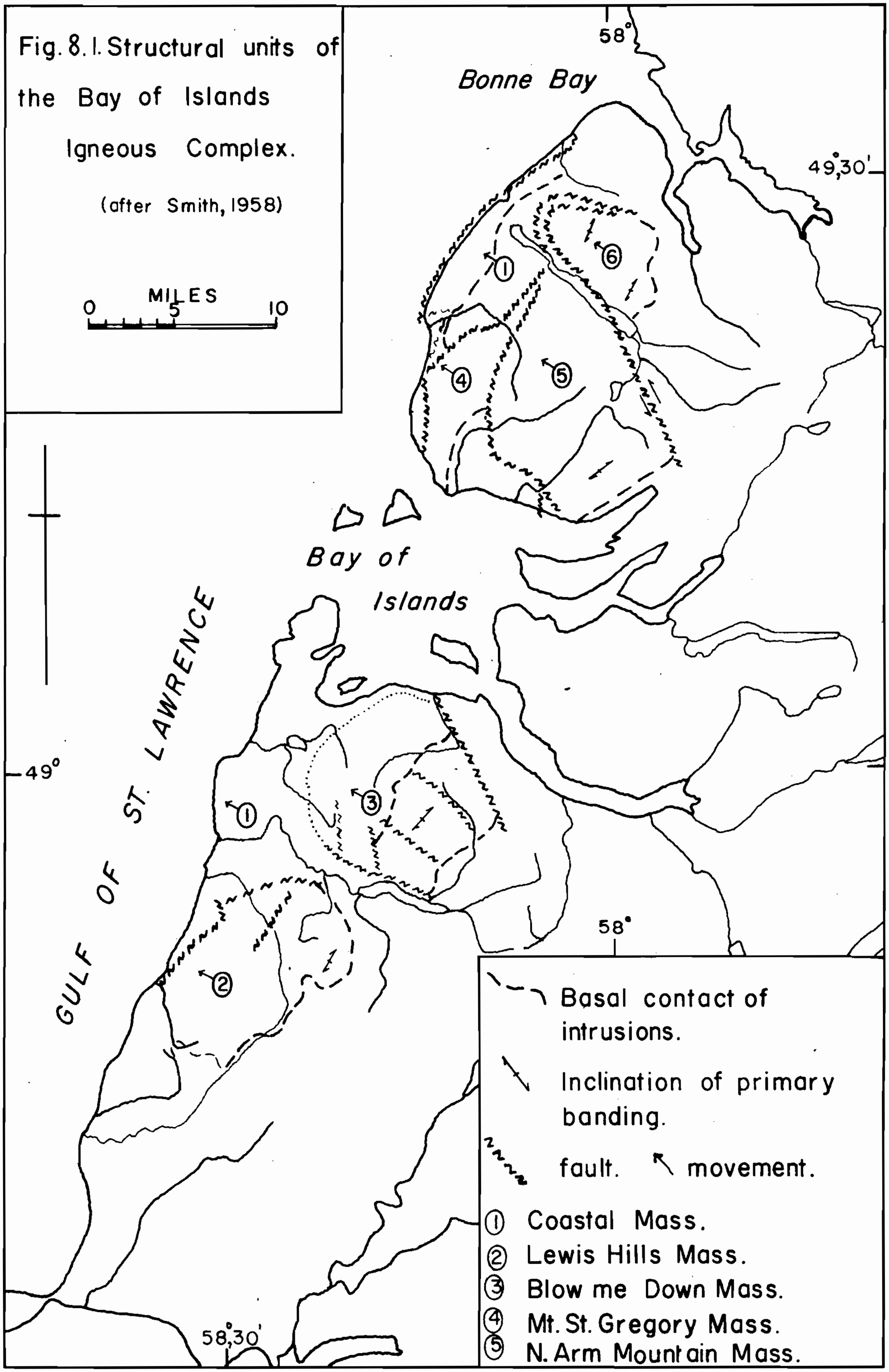


Fig. 8.1. Structural units of the Bay of Islands Igneous Complex.

(after Smith, 1958)

Figure 8.2. The northern face of the Lewis Hills, looking southeast along Serpentine trough from fans of "Sand Desert" at mouth of Rope Cove Canyon.

(ii) Around the head of Lewis Gulch in the south of the Lewis Hills, westwards to Bluff Head and east to Big Level.

(iii) The northern part of the Blow me Down Mountains.

(iv) Table Mountain, Bonne Bay between the Gulch and the Trout River Ponds.

(i) <u>The northern and central Lewis Hills</u>.

The Serpentine River leaves the Serpentine Lake faulted trough between the Lewis Hills and Blow me Down Mountains, and turns sharply south-west then west, to flow to the Gulf of St. Lawrence across a level, drift-covered lowland below 200', which is underlain by Ordovician sedimentary rocks of the Humber Arm Group. Three miles to the south of the river's course in this section the bold northern scarp of the Lewis Hills rises to 2000' from a low-angle fan, the back of which is at about 300' and which slopes from the scarp foot to the north at an angle of less than 2 degrees (fig. 8.2). Three rock types are exposed in the scarp face: volcanic rocks of the Humber Arm Group, and the gabbroic and the serpentinised rocks of the igneous pluton. The serpentinised rock stands cut strikingly since it is bare of all but the lowliest vegetation, due to the toxic effect of magnesium minerals. The yellow ochre colour also is in marked contrast to the darker shades of adjacent rocks. The profile form of the scarp also reflects rock type; in the serpentinised rock the scarp is smooth, less steep, has fewer bare outcrops and a greater talus accumulation than that in the gabbroic and volcanic rocks, reflecting its greater susceptibility to weathering, both mechanical (mainly frost shattering) and chemical. The effect of lithology upon the form of the slopes is well seen in the sides of Rope Cove Canyon (see fig. 8.3.).

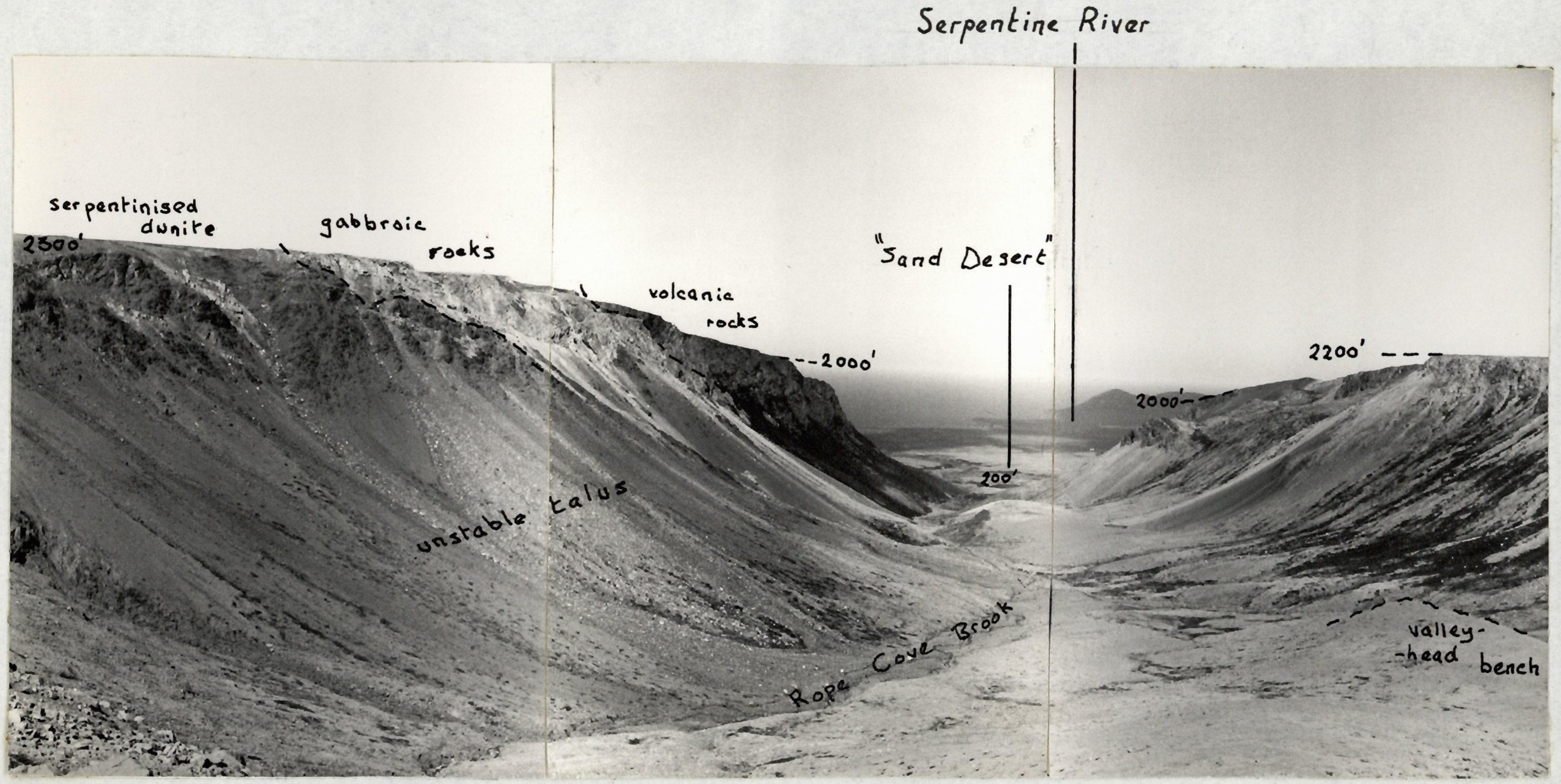


Figure 8.3. Looking down Rope Cove Canyon from headwall at c. 1700', showing influence of rock-type on slope profiles. Serpentine River 7 miles distant.

The first major flat area surmounting the scarp of the Lewis Hills in the north is at 2000-2100'. This extends back into the mountains on spurs on either side of Rope Cove Canyon. Flat areas at this elevation are widespread throughout the Lewis Hills, usually occurring as valley-side flats and minor summits. The widespread distribution and modes of occurrence of these flat areas suggest that they are parts of a subaerial erosion surface (fig. 8.4). To the east and west of Rope Cove Canyon the spurs show a break of slope at the back of these last flat areas above which are found similar features, better developed, at about 2150-2300'. The distribution and modes of occurrence are similar to the lower surface, but this surface is more widespread, particularly in the central parts of the mountains, around the fringes and on the summits of the southwest to northeast-trending hills, from Big Level in the south to the head of Rope Cove Canyon. A higher set of flat areas occurs at 2350-2500' in such areas as Big Level, the area fringing the monadnock whose summit is at 2673', and the summit to the northeast of this (fig. 8.4). This is similar to the lower surfaces described above, but its distribution is more limited. These three highest erosion surfaces were the only ones examined closely in the field, although others occur at lower elevations and can be seen in sweeping views of the mountains obtainable from the 2673' summit (fig. 8.5) and on the contoured maps. One of the most prominent of these lower levels is seen in humerous flat areas found at 1800-1900', mainly on spurs but also on valley sides; similar situations to those in which the higher levels are found, implying a similar origin for them. The only other evidence of erosion surfaces in the central and northern parts of the Lewis Hills is.

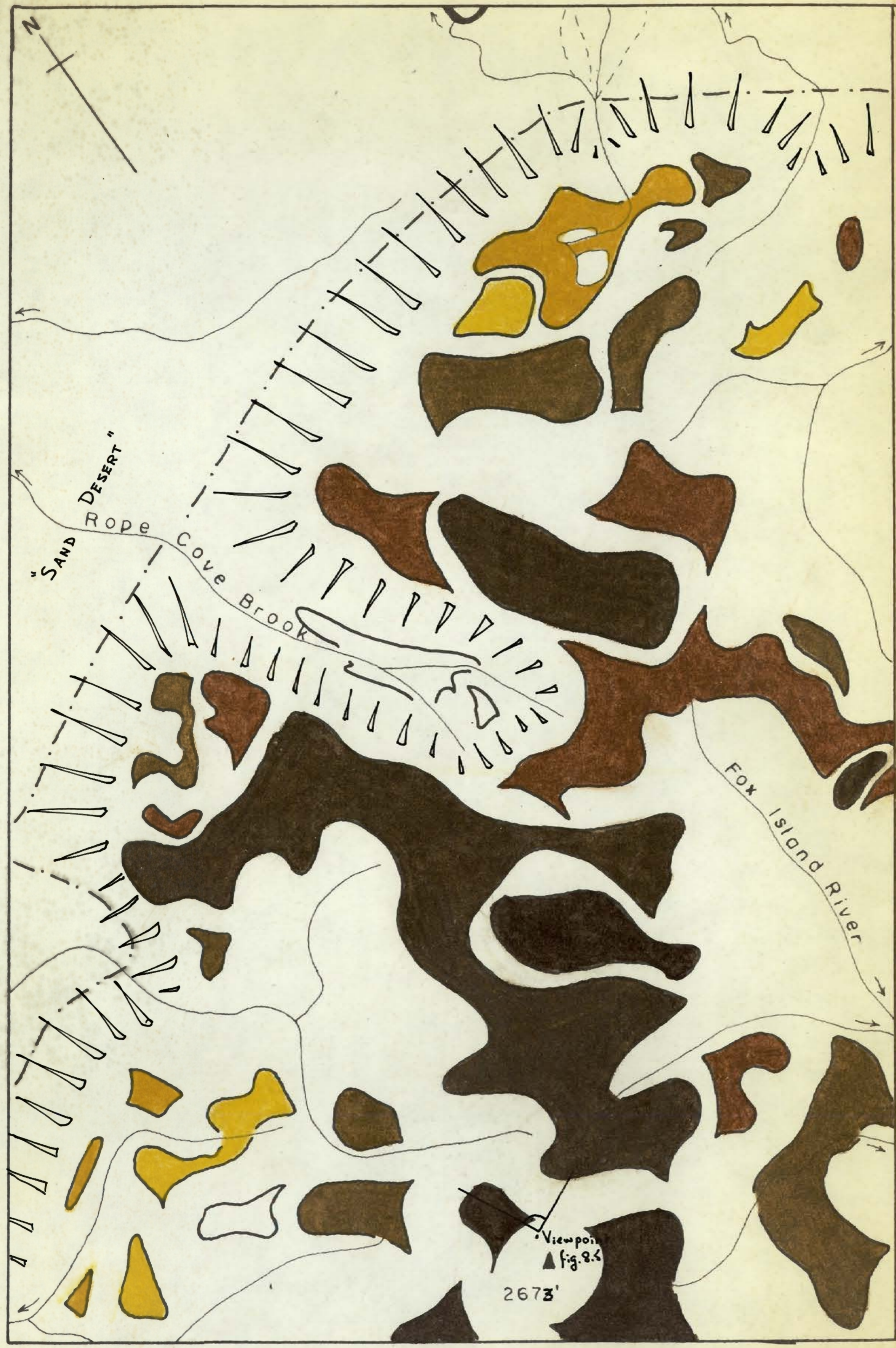


Fig. 8.4. The Erosion Levels of the Central Lewis Hills.

Legend.

2350'-2500'

2150'-2300'

2000'-2150'

1800'-1900'

1600'-1700'

1350'-1500'

other levels

boundary of Lewis Hills 'pluton'

0 1 2

MILES

Figure 8.5. Looking north from 2673' summit across Lewis Hills. "Felsenmeer" of weathered gabbro blocks in foreground. Gulf of St. Lawrence at left.

found at about 1400'. Flats at this height are found only around the edge of the mountains on narrow spurs. In most places around the mountain edge the scarp falls from about 1800' to less than 500' and, thus, most traces of erosion surfaces between those heights have been removed following recent uplift. Better evidence of the lower erosion surfaces is found in the next area to be discussed.

(ii). The Lewis Brook area.

Lewis Brook dissects the southwestern end of the Lewis Hills mass, flowing northwards to the Gulf of St. Lawrence in a valley incised to a depth of 400', on the western slopes of Big Level. Access to the valley is relatively easy along a road which runs from the village of Point au Mal to the bridge across Fox Island River, and then by trail leading from there to a disused asbestos mine in the upper part of the valley (fig. 8.6). The valley now only partly occupied by Lewis Brook is a north-south through valley, in part structurally guided. Upper Ordovician rocks of the Humber Arm Group, which surround the plutonic mass of the Lewis Hills, are prolonged southwards up the valley in a narrow tongue reaching as far as the mine. Rocks in the valley sides are of the gabbroic and serpentinised types common throughout the pluton. On the western side of the lower valley outcrop undifferentiated gabbroic rocks of the same type as those described above occurring on the west side of the entrance to Rope Cove Canyon. The dominant structures of the area are (a) the Coastal Fault (Smith, 1958) of the Lewis Hills mass, along which that part of the igneous complex has been thrust to the west for a distance of about six miles, thus disrupting its former continuity with the Blow me Down Mountains mass and (b) the east-west

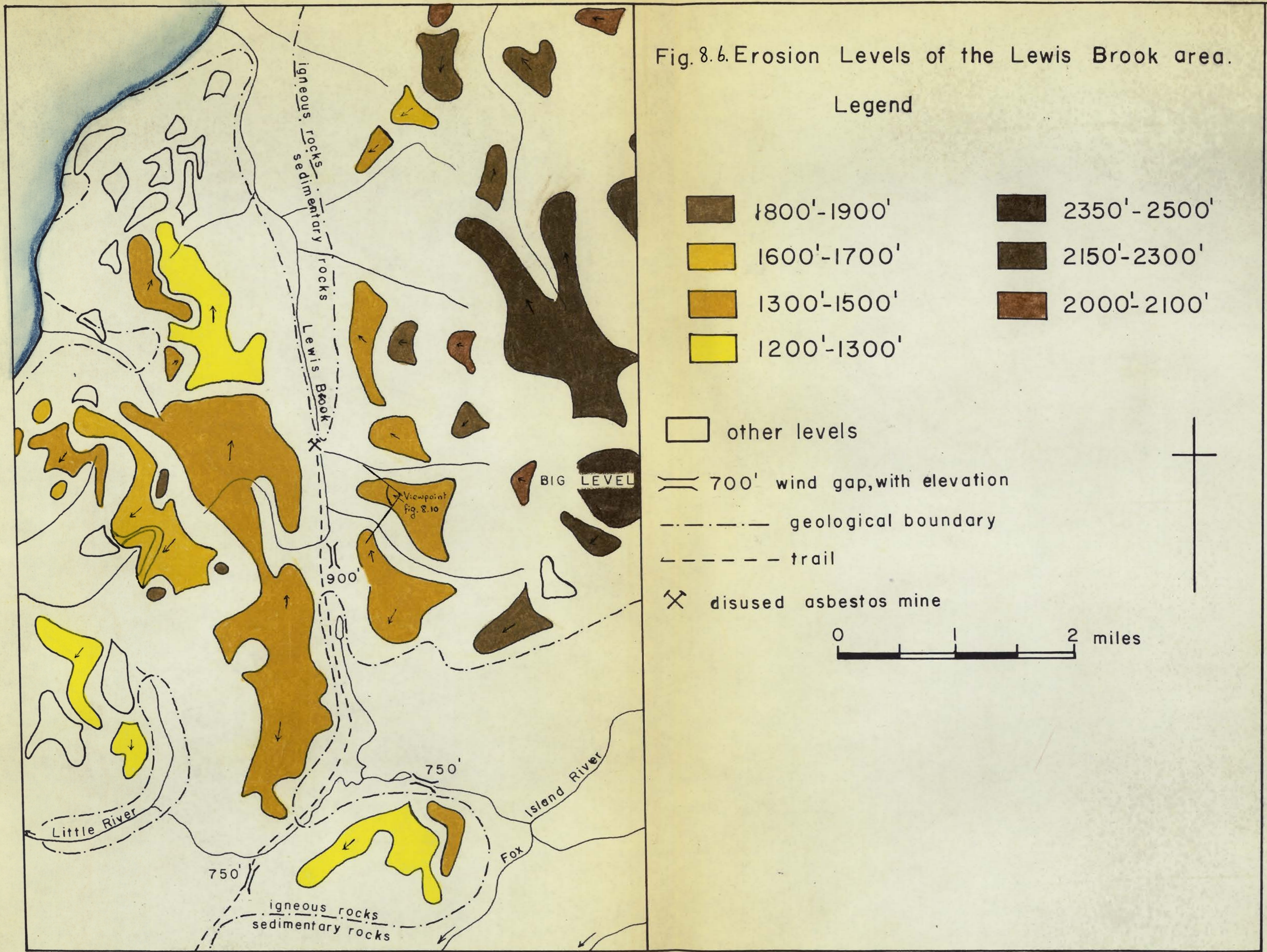


Fig. 8.6. Erosion Levels of the Lewis Brook area.

Cache Valley syncline followed by the lower reaches of Little River.

The highest erosion surface in this area is represented by the remarkably flat summit of Big Level at about 2400' (fig. 8.6 & 8.7). This surface is the same as that at 2350-2500' described in the region to the northeast. A small spur remnant of the 2150-2300' surface is seen on the southern slopes of Big Level and the col to the north of that summit is also a part of this surface. Small flat areas at 2000-2100' are seen on the higher valley sides east of Lewis Brook, and similar features at 1800-1900' are poorly, but significantly, found in the same situation. These two groups of flats are parts of the erosion surfaces seen at those elevations further northeast in the Lewis Hills. On the west side of the Lewis Brook valley three small summits, two at above 1800', and one at above 1750', are correlative with the more widespread remnants of the 1800-1900' surface east of the valley. These summits stand above the eastern edge of a southwesterly sloping flat area at 1600-1700' (fig. 8.8). The latter level is an erosion surface of subaerial origin, as is attested by its slope with the drainage lines across it and the variable height of its back edge. This surface appears not to have an equivalent in elevation on the east side of the valley. The eastern boundary of this 1600-1700' surface is marked by a steep drop to a broad strath in the valley of Lewis Brook at 1300-1500'. The relationships of the three surfaces (1300-1500'. 1600-1700' and 1800-1000') is seen in figure (8.9).

The 1300-1500' valley strath is the most well-marked and striking of the erosion levels in this area. It occurs at the same elevation on

Figure 8.7. Summit of Big Level at 2450', looking northwest. Degenerate gabbro "felsenmeer".

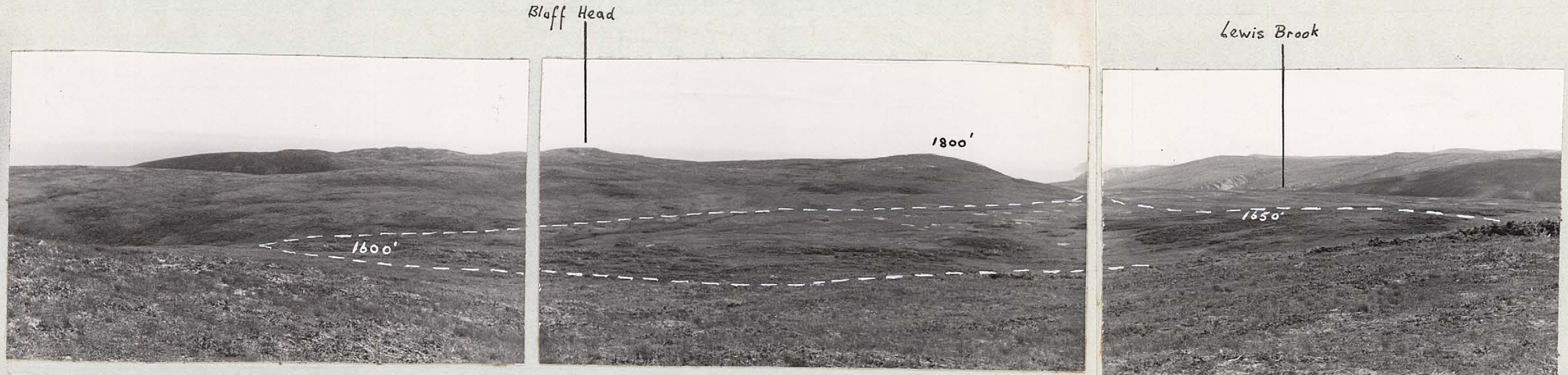


Figure 8.8. The 1600-1650' upland valley surface west of Lewis Brook, looking northwest.

each side of the valley, but generally is more restricted in its development on the eastern side. Above the head of Lewis Brook, in the main valley, remnants of the strath are at their widest and have very low transverse and longitudinal slopes (Fig. 8.10). Down the valley, the remnants become narrower and these slopes increase. This is not the expected situation if the strath had been cut by a north-flowing stream, apart from the fact that the surface slopes to the north in the area north of the col at 900' (fig. 8.6). South of this col the surface remains well developed but has a southerly slope. This imples that, at the stage when this surface was cut, a drainage divide must have passed across the valley from the small remnants of the 1800-1900' surface west of the col to the vicinity of Big Level to the east.

West of Lewis Brook, lower surfaces than the 1300-1500' valley strath are seen. The wide remnant of the strath west of the mine drops steeply from 1450' at its front edge to 1300' at the back of a lower flat area which slopes northwards to 1200' (figs. 8.6 & 8.10). This steep drop marks a geological boundary in the same way as that at the back of the strath. This boundary is that between the gabbroic rocks upon which the strath is developed, and the serpentinised rocks across which the 1200-1300' valley level has been formed. The cyclic significance of the lower level is not certain but, since remnants are not found on the east side of the valley, it is likely that the lower valley level at 1200-1300' belongs to the same stage as the higher one and was developed on the less resistant serpentine rocks. The back of the lower level coincides very closely with the boundary

Figure 8.9. The relationship of the 1800', 1600' and 1300-1500' surfaces at the back edge of the latter, west of Lewis Brook.

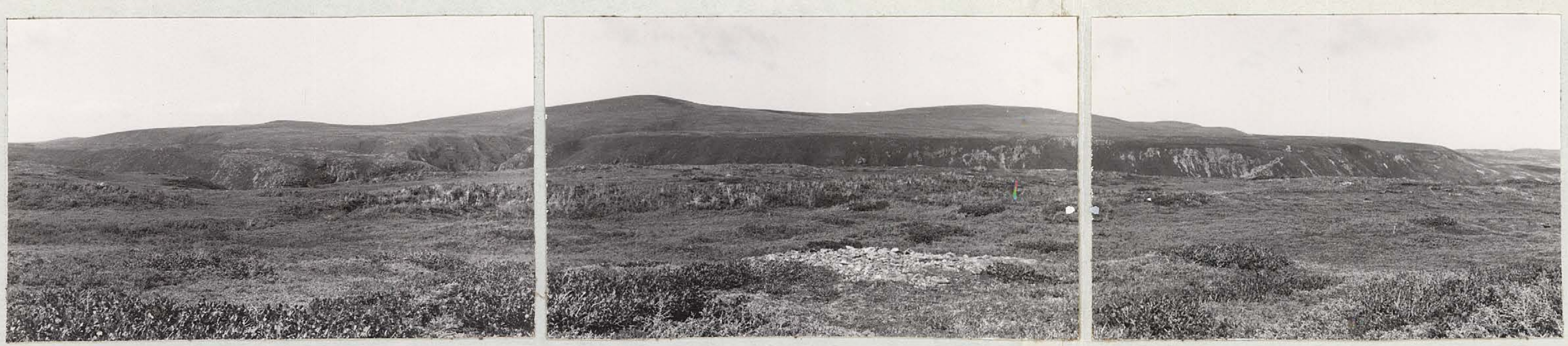


Figure 8.10. The 1300-1500' valley strath above Lewis Brook gorge, looking west from the lower slopes of Big Level.

between the gabbro and the serpentinised rock. The only evidence contrary to this conclusion comes from the hill area south of Cache Valley where the higher level is seen atop a hill along its eastern edge and the lower level is seen below it to the west. Since there are no mapped geological differences in this area, it is possible that the two levels have cyclic significance there.

Minor spur flats west of Lewis Brook, particularly those at about 1100' definitely are remnants of fluvially eroded surfaces in the old valley of the brook, but are not of major significance (fig. 8.6).

The findings from the Lewis Hills may be summarized as follows:-
(i) they comprise a regular oval-shaped upland mass rising from sea level to 2673'.

(ii) erosion surfaces of subaerial origin are found upon them at 2350-2500', 2150-2300', 2000-2150' to the east of Lewis Brook and at 1800-1900', 1600-1700', 1350-1500', and 1200-1300' to the east of the brook.

(iii) no evidence of tilting of these surfaces is found; they appear equal in extent and stage of development on all sides of the mass, and at roughly comparable elevations.

(iv) drainage evolution appears to have been simple, involving progressive adjustment to structural weaknesses, such as the Cache valley syncline, the Fox Island River fault and the lithological boundaries between the igneous and sedimentary rocks, of an initial pattern which was predominantly radial.

(iii) The Northern Blow me Down Mountains.

The Blow me Down Mountains present a very similar appearance to that of the Lewis Hills. They are of the same lithology and structure, but in plan their form is more rectangular than the elongate Lewis Hills to the southwest. These mountains were examined only in the northeast part above Blow me Down Brook and in the northwest above Mine Brook, but views from these areas were wide enough to afford evidence for some general conclusions, in addition to the detailed study of the two field areas.

The northern edge of the Blow me Down Mountains overlooks the Bay of Islands in an impressive scarp, which in one place rises to 2100' from sea level, and which has a general elevation of 1700-1800'. In the northeast, the scarp overlooking the settlement of Frenchman's Cove is capped by a gabbroic layer of superior resistance to denudation than the serpentinised rock of the lower part of the scarp. The gabbro gives rise to steep, craggy free faces up to 200' high, showing many signs of active retreat under the action of frost shattering. Gabbroic debris weathered from these crags, and from the wide amphitheatre-like gully-heads in the scarp face, litters the smooth lower scarp slope formed in the softer serpentinised rocks. These latter rocks weather easily along closely-spaced, curvilinear and rectangular joints. This weathering is facilitated by the solution of pneumatolytic minerals which were deposited along the cooling joints during the later stages of plutonism. When this rock is exposed on the scarp slope it is usually in the upper parts as craggy interruptions in the general smooth slope profile which was described in the centre of the island as "castellated and bristly" by the Scots traveller, W.E. Cormack, in 1822. In time these interruptions are eradicated by weathering of the serpentinised rock to rounded and platy fragments of gravel size which accumulate in a talus. This talus extends itself upslope to "smother" the outcropping crag. The talus slope composed of the fine fragments is highly mobile and is extremely difficult to traverse on foot. Where the upper slopes are composed of the gabbroic rocks, a constant supply of weathered debris is furnished to the talus slope, which maintains a constant angle during retreat. However, when the whole slope is in the serpentinised rock, the free face at the scarp top is soon eradicated by the merging of the actively developing waxing and waning slopes from above and below, respectively. The contrast in slope profiles between the gabbro-capped scarp and that wholly in serpentinised rock is shown diagramatically in figure 8.11 below.

FIGURE 8.11

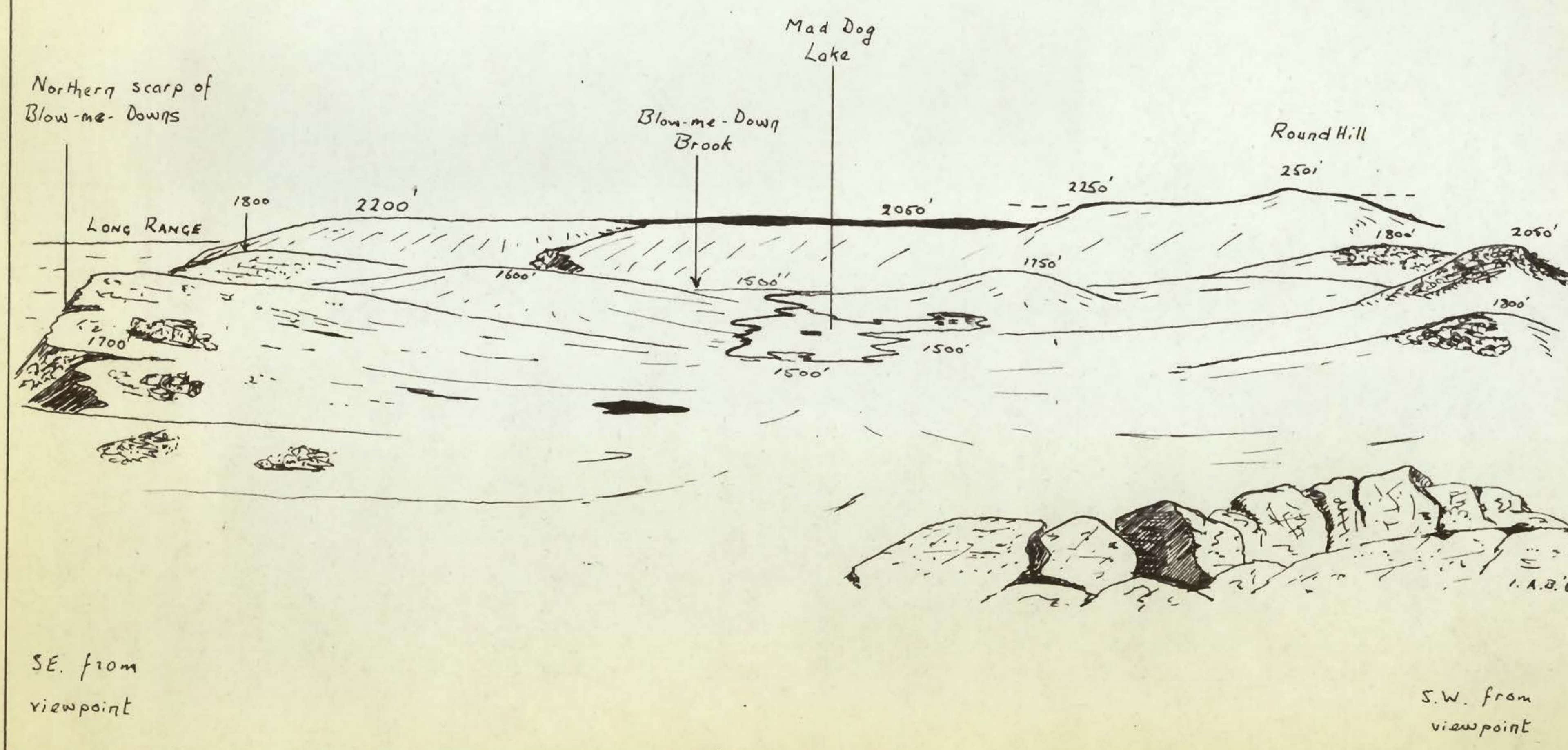
The Influence of Rock-Type on Hillside Slope Profiles, Northern Blow me Down Mountains

The effect of the differences in resistance to weathering of the gabbroic and the serpentinised rocks on the form of the level surface of the mountains is also marked. On the gabbroic rocks, slopes are steeper, relief is greater, a thin soil cover is developed and some vegetation is present, sometimes in the form of small coniferous trees. On the serpentinsed rock slopes are low, vegetation is virtually absent and the surface is littered with a felsenmeer of weathered debris of all grades, from large blocks many feet in diameter to gravel size.

The general form of the erosion surfaces and their modes of occurrence are similar to those in the Lewis Hills, but they are less symmetrically arranged. The highest part of the Blow me Down Mountains is in the southwest of the massif, and the main physiographic lineament is at a lower elevation, trending east to west from the highest summit across its southern part. Only a small part of the summit area of Round Hill (2501') forms a remnant of an erosion level at above 2400'. Surrounding this remnant on three sides is a more extensive level area at 2250-2350' which is most likely equivalent to that extensive interfluve area at 2150-2200' in the southeast of the massif. Towards the west, this interfluve area curves around the head of Sims Brook and a step is visible between this last level and a lower one at 2000-2100'. This level is seen widespread throughout the massif. Other levels seen at elevations equivalent to those in the Lewis Hills are found at 2000-2100', 1800-1900', 1650-1750', 1500-1600' and 1300'. Minor levels are seen around the head of Mine Brook in the northeast of the mountains at 1200-1300' and 1000-1100', corresponding to levels found at those elevations in the Ordovician sedimentary area to the east.

In the northern Blow me Down Mountains the local development of the erosion surfaces, and the views obtained of those forming the horizon, enable the pattern of surfaces over the whole of these mountains to be

Figure 8.12. View Across Northern Blow-me-Down Mountains From the Northern Rim



well seen. The view south over the mountains from the northern rim is seen in the sketch in figure 8.12. The skyline is formed in the southeast by a level-topped spur at about 2000' and is replaced towards the south by a higher level, referred to previously as forming the main lineament of the physiography in the southern part of the massif, at about 2200'. The summit of Round Hill at 2501' can be seen in the distance. In the middle distance, in the southeast, the main surface is that atop the northern scarp of the mountains at about 1750'. Because of the retreat of the scarp, this surface has been truncated, so that now it slopes southwards away from the scarp crest. Formerly, it would have continued to rise for some distance further north, and then would have fallen gently towards the drainage line passing through the Bay of Islands. This surface, then, slopes southwards towards Mad Dog Lake which occupies a broad, flat-floored area, appearing as an interior lowland in the encircling ring of higher ground. This is a part of an erosion surface at 1500-1600', which is poorly developed as a strath on the southern side of the mountain section of the Blow me Down Brook valley. To the west of Mad Dog Lake the land slopes up steeply to the summits of several isolated hills, which lie at 1700-1800', and which are parts of the erosion surface seen atop the northern scarp. The peak of Blow me Down Mountain itself, at 2134', is probably a part of the surface on the interfluve seen to the southeast across Blow me Down Brook see figure 8.12.

The drainage pattern of the Blow me Down Mountains is very similar to that of the Lewis Hills. In fact the two massifs may be considered as


Fig. 8.13. The Erosion Levels of the Table Mountain area, Bonne Bay.
Legend.
2300'-2350'+
2100'-2250'
1900'-2100'
1750'-1900'
1550'-1700'
1400'-1500'
1250'-1350'
1100'-1150'
Fault
igneous - sedimentary border
0 ½ 1 2
MILES
Trout River Pond
Viewpoint fig 8.15
Winterhouse Brook
GULCH FAULT
South Arm (Bonne Bay)
N

mirror-images. They both have drainage to the Serpentine trough, a long-established drainage line, and both have a broadly radial pattern. In both massifs, one side is bounded by a deep fault-guided valley - Fox Island River in the Lewis Hills and Blow me Down Brook in these mountains - and in both areas river capture on the fringes of the massifs has caused some disruption of the original radial pattern.

(iv) <u>The Table Mountain Area, Bonne Bay</u>

The Table Mountain massif of the Bay of Islands Mountains has a roughly rectangular plan form. It is bounded on the north, west and south sides of rectilinear or curvilinear faults and on the east by an abrupt north-south contact with the Ordovician sedimentary rocks. The block is nine miles long from southeast to northwest and has a maximum width of 4 miles from southwest to northeast. The predominant rock outcropping is the serpentinised type found elsewhere in these mountains, and a smaller body of gabbroic rock outcrops in the northwestern end of the massif (end plate 2). Because of the predominance of the serpentinised rock, the mountain's surface, apart from the levelness imparted to it by denudation, has an evenness due to the smoothing effect of the plentiful amount of weathered debris which covers the surface (fig. 8.14).

The summit of Table Mountain is at above 2350' which represents the upper elevation of the uppermost erosion surface at 2300-2350' (fig. 8.13). At its widest, this surface extends east to west for two miles and has a barely perceptible downward slope in all directions.

Figure 8.14. The summit plateau of Table Mountain, Bonne Bay, at 2300'. Degenerate "felsenmeer" of serpentinised dunite. Horizon two miles distant. View southeast.

Figure 8.15. Looking southeast along Trout River Ponds faulted trough from southern edge of Table Mountain at 2000'. Lake at sea-level. Jointed serpentinised dunite at left. Lower surfaces on volcanic rocks at right.

Surrounding this summit surface, and separated from it by a slight break of slope about 50' high, is a lower erosion surface occurring mainly on spurs, radiating from the summit area between deep canyons, which mark the effects of erosion consequent upon recent rejuvenation. This surface lies at 2100-2250' and occasionally rises to 2300'. The next lower level is at 1900-2150', but the latter elevation is reached only by the back of one flat area, which occupies a somewhat anomalous position. This anomaly appears to be due merely to the exigencies of erosion surface development, since geological factors cannot be invoked. Towards the northwest, the surface of the Table Mountain massif decreases in elevation and no remnants of erosion surfaces so far noted are seen. The spurs projecting northwest from the summit of the mountain have a stepped appearance in profile and the elevations of these steps can be correlated all around the massif. Below the 1900-2150' surface the next noticeable flattening occurs at about 1750-1900'. Isolated summits in the extreme northwest of the massif bear remnants of this level above upland valley floors at 1700', 1450-1500' and 1300-1400'. The latter level also occurs on spurs projecting from the 1750-1900' summits. The lowest noticeable erosion level in the Table Mountain area is at 1200'. Remnants of this level are scarce, but equivalents are found across the Trout River Ponds faulted trough in remnants of an erosion surface at 1250-1350', well developed upon metamorphosed volcanic rocks in the St. Gregory Highlands. (fig. 8.13).

In summary, the Table Mountain massif exhibits several erosion levels (shown in fig. 8.13), some less distinct from those above and below than others, but all fitting into a recognisable pattern. The surfaces are found at the following elevations: 2300-2350', 2100-2250', 1900-2150', 1750-1900', 1450-1500' and 1200-1250'. Below the lowest level, the massif is bounded by steep scarps, commonly along fault lines, such as along the Trout River Ponds fault system which rise from sea level to 2000' (see fig. 8.15).

(iv) Summary.

The question arises as to whether the difference in elevation between the surface of the plutonic massifs and the surrounding sedimentary areas is the result of erosion acting differentially on the harder rocks of the plutons and the softer sedimentary rocks, subsequent to the recent wholesale uplift of the massifs and the surrounding sedimentary areas, or whether the plutons have moved upwards along their borders, which are often faulted, leaving them standing above the surrounding sedimentary areas. Clues to the solution of the problem are found in a study of the erosion surfaces. The highest erosion surface found in the sedimentary area is at 1150-1300'. The only areas where remnants of this surface are found in the massifs are in the northwest of the Blow me Down Mountains, in the area west of Lewis Brook in the southwestern Lewis Hills, and in the upper parts of Fox Island River in the southeastern Lewis Hills. Elsewhere, the massifs are bounded by steep scarps which, on their eastern sides, rise abruptly from the 1150-1300' surface on the sedimentary areas to the 2000' surface of the mountains. The lack of evidence for the development of the 1150-1300' surface in the mountains suggests that these have been uplifted separately and by a greater amount than the surrounding sedimentary areas. The bulk of this vertical movement must have taken place in geologically recent times, shortly before the Quaternary, since the deep valleys cut into the mountain edge appear too youthful for their cutting to have been initiated at an earlier date. While such an uplift adequately accounts for the relative elevation of the mountain surfaces at the present, the movements which terminated the development of each period of erosion and which gave rise to the morphologic discontinuities were of a lesser magnitude. It has previously been suggested that these discontinuities, represented by the vertical intervals between the several erosion surfaces, were the result of periodic minor isostatic adjustments to the relief of load following denudation. Such a mechanism has been described by Gunn (1949), and applied by Baulig (1935) and Schumm (1963).

Drainage evolution in the area north of the Bay of Islands appears to have been much the same as that to the south, between the bay and St. George's Bay. Throughout the area (region VI, end plate 1), up-standing mountains lie to the west of dissected plateaux and are separated from each other by transverse faulted troughs. It is likely that these lines of weakness have been the sites of the main drainage lines throughout the period in which the present land forms developed. At the time of formation of the upper erosion surfaces (generally above about 1750') it is likely that a major drainage divide traversed the length of the Bay of Islands Mountains and extended into the Northern Peninsula, and that drainage flowed both northwest and southeast from this divide. No major land upheaval occurred during this period, as is evidenced by the small vertical intervals between surfaces in the highest group. With progressive adjustment of the drainage to structure, particularly to the transverse zones of weakness, the divide was moved to the east as northwest-flowing streams captured the headwaters of those flowing southeast. This process has continued so that now the northwest-flowing streams head far to the east of the line of highest summits in the plutonic mountains. Smaller scale adjustments to structures in the plutonic and sedimentary rocks have given the present pattern.

Finally, before leaving the Bay of Islands Mountains, the erosion surfaces may be correlated between each physiographic unit as shown in table III below.

TABLE III

Correlation of Erosion Surfaces in the Bay of Islands Mountains

Lewis Hills	Blow me Down Mountains	North Arm Mountain - Mount St. Gregory Highlands	Table Mountain
2350-2500'	2400-2500'	2250-2300'	2300-2350'
2150-2300'	2250-2350' 2100-2250'	2100-2200'	2100-2250'
2000-2150'	2000-2100'	1950-2050'	1900-2150'
1800-1900'	1800-1900'	1750-1850'	1750-1900'
1600-1700'	1650-1750'	1650-1750'	?
1300-1500'	1500-1600'	1350-1500'	1450-1550'
1200-1300'	1300'	1200-1300' 1100-1200'	1200-1250'

Chapter Nine: The Long Range Mountains of the Northern Peninsula

Although no detailed field work was possible in the Northern Peninsula, views obtained from a distance and study of the topographic maps enable a few general points to be made. The peninsula extends from north of Bonne Bay, some 270 miles north-northeastwards to the Strait of Belle Isle, which separates the island of Newfoundland from the mainland of Labrador. Like a spine running almost the whole length of the peninsula are the Long Range Mountains. They are composed of Precambrian high-grade metamorphic gneisses and schists and are bounded by pronounced faults, on the western side where a scarp separates the mountains from the low coastal plain, and on the east where the remarkably straight coastline reflects a major fault (end plate 4). The crest of the mountains occurs above the scarp on the western side, sometimes set a little way back from the mountain edge, and decreases in elevation northwards from above 2600' north of Bonne Bay to 1800' near the northern end of the range.

The present drainage divide of the peninsula follows the crests of the highest summits to the west, with sharp easterly-directed re-entrants along the broad, shallow upland valleys. From this divide, streams flow roughly southeastwards in wide, mature valleys, sometimes slightly overdeepended by glacial erosion, and their lower reaches have been incised by a wave of rejuvenation which has passed up them. The upper valleys of these streams are at about 1400-1500', at which height a broad, flat-floored col separates them from a short, shallow, upland valley section belonging to west-flowing streams, which become very deeply incised in crossing the 2000-' high scarp. Vigorous glacial erosion has overdeepened these canyons, for example Western Brook Pond canyon, which has walls rising 2300 feet from near sea level. The highest erosion surface is that atop Gros Morne (2651') seen from the summit of Table Mountain in figure 9.1. The surface here occurs at 2450-2600' and is sensibly flat for a distance of half a mile on the summit of the mountain. The levelness of this summit may, in part, be due to the effects of flat-lying Cambrian sediments, but elsewhere the surface is just as level bevelling the Precambrian rocks. To the north the eastward (inland) extent of the surface widens until, on the interfluve between Bakers Brook and Western Brook Ponds, it is seven miles wide. The surface slopes in all directions from its highest parts which are set back from the western edge of the mountains.

Below this summit level, which is at approximately the same elevation as that seen in the Bay of Islands Mountains, other erosion surfaces occur. They are less widely developed in the western parts of the mountains and have their maximum extent further east. They also slope in all directions when they occur on summits and are found associated with valleys cut by east and west flowing streams. These surfaces are at 2250-2350', 2200-2300', 2100', 1800-1900', on spurs which project in all directions from the summit areas, and become increasingly well developed towards the east on the "backslope" of the mountains. On the west, the scarp truncates the surfaces. Their high degree of development suggests that they once extended some distance to the west of the fault line and

Figure 9.1. View northeast from northern edge of Table Mountain, Bonne Bay, to Long Range Mountains of the Northern Peninsula. Gros Morne (2651') in left centre. "Felsenmeer" of serpentinised dunite.

have been truncated by recent vertical movements along it. Thus, in this case, as on the margins of the Bay of Islands Mountains, the height differential between the mountain surfaces and the low or intermediate areas surrounding them is due more to recent diastrophism than to differential erosion on either side of the lithological boundary. Another analogy with the Bay of Islands Mountains is the drainage evolution. With uplift along the western faultline, the activity of west-flowing streams was invigorated and these proceeded to cut back into the mountain front, but have not succeeded in capturing east-flowing streams probably because the surface of the mountains was tilted to the east, thus allowing these streams to maintain their easterly flow. Diagramatically, the evolution tentatively proposed for the western parts of the Northern Peninsula Long Range Mountains may be summarized as in figure 9.2.

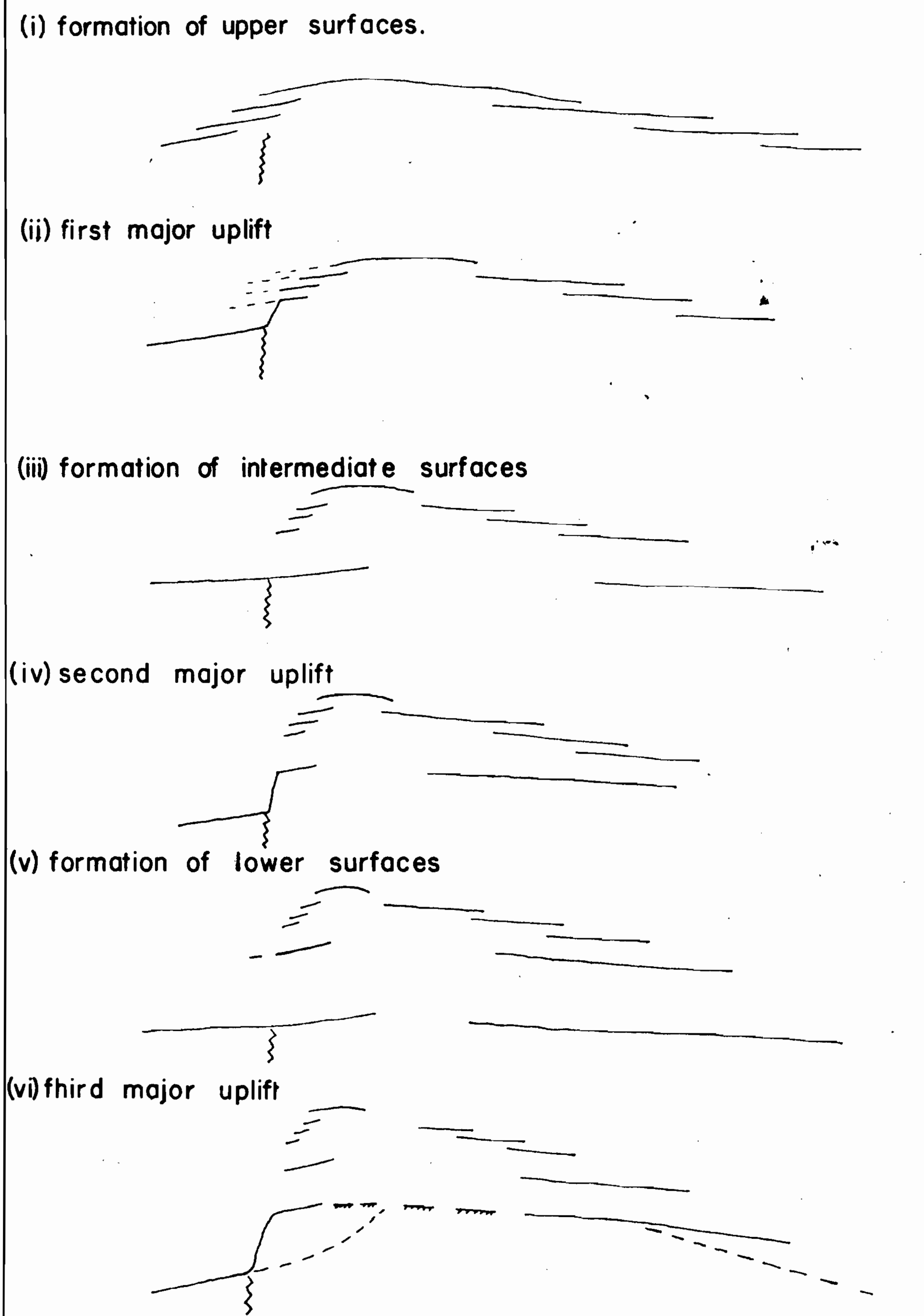


Fig.9.2 Suggested Geomorphic Evolution of N. Peninsula.

I.A.B.

PART THREE

Chapter Ten: Summary of Erosion Surfaces and Conclusions Regarding their Number and Origin.

In reviewing the field evidence of erosion surfaces in western Newfoundland, it has been concluded that those surfaces found can best be explained by a multicyclic evolution of a landscape undergoing denudation by river action and slope wastage in periods of relative crustal quiescence, interrupted by many shorter episodes of crustal uplift, themselves grouped into discrete time periods, which were in turn separated from each other by periods of larger scale uplifts due to different causes.

The main objects of a study of this kind are:

(1) to ascertain whether or not erosion surfaces are present and, if so,

(ii) to enumerate these surfaces and the number of cycles or partial cycles represented.

(iii) to ascertain the origin of the surfaces.

(iv) to ascertain their absolute or relative age.

(v) to correlate the surfaces, where possible, with those found in other areas.

Objects (i), (ii), and (iii) have been satisfied in the foregoing exposition of the evidence and will be dealt with here in that order. Discussion of subject (iv), concerning the ages of the surfaces is presented in chapter eleven, together with proposed correlations with other areas.

As far as the recognition of erosion surfaces is concerned, it has already been noted that this has an element of subjectivity. However, subjectivity of approach is not in itself an objection to the findings of such an enquiry, since every scientific investigation is undertaken with some preconceived idea of what might be found. When the results expected do come from the enquiry, this is not to say that the latter has no scientific value or that it was undertaken on false premises. When results other than those expected come from the enquiry, the hypothesis which it tested must be revised.

The writer undertook the preliminary cartographic work and the field work fully expecting erosion surfaces to be present in western Newfoundland because of the situation of the area with respect to other areas lying within the same geological and tectonic province of eastern Canada, and possessing broadly similar physiographic traits, in which the existence of erosion surfaces previously has been demonstrated. This expectation was fulfilled, as the foregoing review of the evidence in the field area shows. Had it not been fulfilled, the absence of erosion surfaces so well represented in neighbouring area would have demanded explanation. This explanation would have presented a problem of a truly geographical nature: the explanation of areal differentiation over the earth's surface. The demonstration of the existence of erosion surfaces in western Newfoundland constitutes a contribution to the geography of the Atlantic Provinces of eastern Canada, in that these surfaces are a unifying factor over a large area of the earth's surface: a factor physical in essence but also with an important bearing on the human geography of the region.

Concerning the enumeration of the surfaces and the cycles represented, the former were at first thought to be embarrassingly numerous. A geomorphologist who is investigating a new problem, or re-investigating an old one, must be assiduous in observing and reporting what actually is to be seen on the surface of the earth. Too often, geomorphological hypotheses are based on faulty observation or inaccurate or insufficient description. Although the description in itself may be somewhat tedious (not the least to the reader), its importance can not be underestimated. Therefore, in reporting in detail the surface features of the earth, the proliferation of erosion surfaces need not necessarily arouse suspicion. The justification for the number, the small vertical intervals and the relatively great height ranges of the surfaces has already been given.

The importance of recognising a pattern in a landscape must again be stressed for its significance in interpreting the initially confusing proliferation of erosion surfaces. Patterns of different scales are perceptible. On the broadest scale, the pattern is very plain; it is one of generally level surfaces at high elevations. Narrowing the field of view, a more complicated pattern emerges. At this scale, the land surface can be generalised to conform to three major groups of erosion surfaces. The highest group of surfaces is that including the summit surfaces of the Long Range Mountains and those of the Bay of Islands Mountains. In the

Long Range Mountains between Cape Ray, in the extreme southwest and the shores of Grand Lake (region Ia, end plate 1), this group of surfaces usually has only one member, - the 1800-2100' surface, although too brief observation may have simplified the true picture. Remnants of this group of surfaces in this area are usually too small in extent to enable the determination of a general slope. Most remnants are flat so that any attempt at a reconstruction of drainage lines at the time of the formation of this surface can only be based upon the distribution of the remnants and must be tentative.

The mountains of the Bay of Islands Igneous Complex display a physiographic unity by virtue, not only of their geology and structure, but also of the occurrence here of erosion surfaces falling within this highest group. In each of these massifs there are four members of the group (table III). They are distinct from the lower group of surfaces, since the latter are not significant for the general form of the surface of the mountains. These surfaces of the highest group show no signs of having been deformed by diastrophism. This is contrary to expectations, since they have been uplifted to their present elevations by movement along the well-defined fault-line, which is reflected in the south-southwest to north-northeast trend of the coastline of this and most other parts of western Newfoundland. It would be expected that such a movement would have resulted in the back-tilting of the erosion surfaces, but no tilting is evidenced. These surfaces appear to have been uplifted vertically, implying that movement has occurred, not only along the coastal faultine, but uniformly, aided by regeneration of the well-marked faultines bounding the massifs around much of their perimeter (see map of geological structures in the Bay of Islands Igneous Complex, (fig. 8.1). Such a movement explains not only the great height of the scarps bounding them, but also the abrupt truncation of erosion surfaces on all sides of the massifs, and the fact that the highest erosion surface in the dissected plateau area to the east of the massifs penetrates the latter in only a few places.

The highest group of erosion surfaces is also found in the Northern Peninsula Long Range Mountains (region I, end plate 1). Although no detailed field work was possible in this area, examination of the relationships of the highest erosion surfaces on the topographic maps gives no reasons for believing that they have been tilted during uplift to their present high levels. The explanation of this lack of tilting is more difficult than in the Bay of Islands Mountains, since it necessitates postulation of the uniform uplift of the whole structural block of the Northern Peninsula. More detailed study is required in this area before this problem can be solved.

The next lowest group of erosion surfaces in western Newfoundland is represented in the southwestern parts of the Long Range Mountains (Region Ia, end plate 1) by the predominant plateau level upon which the remnants of the higher group of surfaces stand. This surface is seen at 1600-1800' to the east of the line of highest summits and is less extensively developed as small platforms to the west of them, both above the Codroy depression and the Long Range Foreland. The slopes of this surface in the southwest of this region suggest that drainage at the time of its formation was towards the east and southeast, east of the highest summits, and westwards towards the Codroy depression, to the west of these summits. This group of surfaces is also seen on the Cape Anguille Mountains in the small areas of the 1750' surface and the more extensive 1550-1700' and 1300-1500' surfaces. Drainage evolution in these mountains has been considered together with that in the Long Range above the Codroy depression (see fig. 6.12) and involves a progressive adjustment to major structural elements.

Traced towards the northeast, this group of surfaces becomes better developed over the western parts of the Long Range Mountains, as in the area around Flat Bay Brook (fig. 6.13) and the western end of Grand Lake.

Drainage evolution in the Long Range Mountains east of the Foreland has already been discussed, but few definite conclusions reached as to its course. The problem of accounting for the transection by the west-flowing river system of the folds and faults within the Carboniferous strata of the Long Range Foreland has been attacked by considering each of the possible causes - capture, superposition and antecedence - in the light of the scanty field evidence and that from the geological literature and topographic maps. None of these possibilities has been found adequate to explain the origin of that drainage.

The middle group of surfaces is represented only by its lower members in the dissected plateau region between the Long Range and the Bay of Islands Mountains. The 1350-1500' surface is preserved on the major east to west divide across the region and the 1150-1300' surface is better developed on the interfluves of both north and south-flowing rivers, and on the plateau area in the southwest of the region, such as Table Mountain. The valley surfaces belong to the lowest group. Around the fringes of the plutonic massifs to the west, this group of surfaces is poorly represented. In most places, the scarps rise from between sea-level and 1200' to the 2000' plateau edge, with no signs of peripheral levels. However, in the Lewis Brook area, surfaces between 1200-1800', occurring as straths and summit levels, belong to the middle group. Similarly, south of the Trout River ponds, this group is represented by surfaces between the same heights.

The lowest group of surfaces comprises those below about 1000'. These are found only in the valleys throughout the field area. Surfaces belonging to this group have been described from the southwestern Cape Anguille Mountains (and are found all around the fringes of these mountains), the Flat Bay Brook area (and are found in the valleys of all the major west coast rivers) and the dissected plateau region between Humber Arm and St. George's Bay.

No evidence that the erosion surfaces of western Newfoundland are of marine origin has been found. The following characteristics of the surfaces lead to the conclusion that they are the product of subaerial denudation during periods of relatively stable base-level: -

(i) All the surfaces slope with the drainage lines.

(ii) They have a height range greater than that normally associated with wave-cut platforms.

(iii) In many places, the surfaces are found in situations where marine erosion would not have been sufficiently effective to have cut them.

(iv) The elevation of the back of each erosion surface remnant, at the foot of the break of slope separating it from a higher surface, is not constant. A difference of 50-100' is found over distances measurable in tens of miles.

(v) No marine deposits were found upon the surfaces.

(vi) Where isolated remnants of a surface occur, separated from the main part of the surface, they do not occupy positions which lead to consideration of them as former sea stacks or islands, standing above a surface of marine erosion.

(vii) The vertical interval between two surfaces, represented by the bluff separating them, is constant at about 50-100'. If this bluff is of marine origin, its height would be expected to vary with the rock type in which it is cut, as do modern cliffs.

No definite conclusions have been made as to the initial form of the erosion surfaces. Use of the term "peneplain" has purposely been avoided, except when reference has been made to the conclusions of previous workers. It is not the writer's intention here to discuss, in the light of the field evidence, the many mechanisms of landscape evolution under various tectonic and climatic conditions, proposed in the hypotheses of, for example, W.M. Davis, W. Penck, L.C. King and J. Büdel. Knowledge of the tectonic and climatic conditions of the Mesozoic and

Tertiary in northeastern North American is inadequate to justify an inductive approach to the problem of the mechanism of landscape evolution during those times, and the landscape itself must be studied in greater detail before it will yield valuable clues. Modification of old landscapes by agencies other than those responsible for their development is an important complication. In Newfoundland, particularly, the effects of periglacial weathering and mass-movement on the pre-glacial landscape are unknown.

Chapter Eleven: The age of the erosion surfaces and correlations with other areas.

Erosion surfaces can be dated relatively or absolutely. This applies to members of a group of surfaces or to the different parts of the same surface. The age of a surface can be determined with reference to (a) the time of initiation of its development, (b) the time of completion or arresting of this, (c) the time span covering the entire development, or to (d) the actual age of the surface in its entirety or in a particular locality. The age of one surface relative to those above and below it, with reference to the time of its initiation, can be determined merely by assuming that the lower parts of the surface (if its undeformed) were formed earlier than similar parts of surfaces below it and later than those parts of surfaces above it. With reference to the time of completion, similar assumptions can be made regarding the relationship of the uppermost parts of a surface to similar parts of surfaces below and above it.[1.] With reference to the time span covering the entire development of a surface, the same assumptions can be made regarding the entire surface relative to those below and above it. These relative age determinations refer to (a), (b) and (c) above. The relative age of a surface in any locality may be determined more accurately if it is developed across deposits the actual age of which is known. The surface can then be assumed to be younger than the deposits.

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1. Although, in the case of a peneplain, development proceeds simultaneously all over the surface, peneplanation is first accomplished in the parts of the surface nearer the base-level, these being the lower parts of upraised undeformed peneplains. In the case of pediplains, development of the surface proceeds with the initiation of a scarp which retreats into the land surface above, leaving a surface below it, the age of which increases with distance from the scarp face.

Determination of the actual age of erosion surfaces, or parts of them, must rely upon geological, rather than morphological, evidence. The most direct method of absolute dating can be applied when the surface is overlain by dateable deposits. The surface must then have been formed immediately prior to the deposition of the overlying materials (sediments or lavas). Also, where dateable deposits overlie an erosion surface, and can be shown to be autochthonous, i.e. intimately associated with the formation of the surface, as with some laterites and duricrusts, then the surface in the locality where the deposits are found must have been formed immediately prior to, or simultaneously with it.

Periods of subaerial erosion on land are represented by marine deposition off-shore. Unconformities in off-shore sequences must therefore mark the termination of a period of erosion of the adjacent land area, and the initiation of a later erosional period. The detritus brought to the sea during this later period is then deposited above the unconformity. Unconformities which are dateable with reference to overlying deposits thus give a minimum age for the completion of a period of subaerial erosion surface development. Correlation of the relative positions of erosion surfaces and unconformities, or the projection of the latter inland on to the former thus give the actual date of the termination of erosion surface development.

In western Newfoundland erosion surfaces can most usefully and easily be dated in relative terms, with reference to their position within the general scheme erected as a result of this study, and absolutely, with reference to the known and hypothetical geological histories of the land areas and adjacent sea-floors of northeastern North America.

It has been concluded that the erosion surfaces found are of subaerial origin, at least above 600' (study of surfaces below this height was not undertaken, so this should not be taken as implying that surfaces below that height have a different origin, although this possibility cannot be ruled out). It has also been concluded that they developed during a number of periods of relatively stable base-level, which were separated from each other by shorter periods of falling base-level. This hypothesis imples that the higher a surface is, the older it is, relative to those below it. Therefore, in western Newfoundland, the group of surfaces found in the Bay of Islands Mountains and the Northern Peninsula above about 2000' are the oldest in the area. These surfaces do not appear to be equivalent in age to the surface(s) at 1800-2000' in the Long Range Mountains to the east and southeast, since this latter surface finds its equivalent in a surface (or surfaces) at similar elevations in the higher areas where the upper group is found. So with other surfaces, their relative age decreases with decreasing height within the general scheme.

As regards absolute dating, the erosion surfaces of western Newfoundland must be younger than the latest orogenic movements which affected the area, i.e. the Appalachian earth movements of immediately post-Pennsylvanian age. They must also be older than the latest major negative change of base-level, which resulted in the elevation of even the youngest of the subaerial erosion surfaces found. Previous students of the physiography of northeastern North American (e.g. Cooke, 1931;

Alcock, 1926,1928) have suggested that this great elevation occurred in the late Pliocene, since the valleys which were deeply incised as a result did not evolve beyond a youthful stage before being glaciated in the succeeding Pleistocene period. The erosion surfaces were thus developed between the Carbo-Permian period and the latter part of the Tertiary era.

Deposits available for the absolute dating of erosion surfaces in Newfoundland have not yet been revealed. However, the recent discovery of Cretaceous clays in the Shubenacadie valley of Nova Scotia (Stevenson & McGregor, 1963) incised into the "middle surface" of Bird (personal communication, 1964), which likely is equivalent to the 1550-1700' surface of the Cape Anguille Mountains in the writer's field area (see chapter five), suggests an age of that surface greater than those deposits if the latter are <u>in situ</u>. Other evidence which points to an age of the uppermost surface(s) of the Atlantic Provinces generally, greater than that expected by comparison with the Appalachians of the Atlantic seaboard of the eastern United States, where the sequence of surfaces and their ages is fairly well known, comes from the conclusions reached in chapter three from Engelen's (1963) hypothesis of continental margin evolution. It has been suggested that the graben formation in the continental margin, which occurred in the early Cretaceous off Newfoundland and spread southwest to reach the Bahamas by the late Miocene, was an expression of isostatic compensation for loading of the margin with continental detritus. Uplift of the continental land areas would have accompanied this subsidence, probably simultaneously. Thus, from Engelen's figures for the magnitude of this movement, $\pm$ 3.8 kms., it is very likely that the present land areas of Newfoundland and much of the Atlantic Provinces would have been uplifted above the maximum limit of the Upper Cretaceous marine transgression which affected areas farther southwest.

Supporting evidence for this conclusion comes from a recent paper by Flint (1963). Flint notes that the Upper Cretaceous Raritan and Magothy sediments underlying Long Island, New York, are non-marine, probably estuarine, and represent, according to Veatch, the transported products of a long period of weathering of crystalline rocks. The estuarine character of these sediments suggests the close proximity of a shoreline, contradictory to Johnson's (1931) postulation of an Upper Cretaceous marine transgression 125-200 miles across the New England seaboard, to account for the superimposition of the southeasterly-flowing drainage. The evidence of Cretaceous non-marine sediments in Nova Scotia (Stevenson and McGregor, 1963) corroborates this conclusion in that area. Further, Flint presents three lines of evidence which cast doubt upon the hypothesis that the drainage of Connecticut was originally consequent upon a cover of coastal-plain sediments and has been superimposed from this on to the structures in underlying crystalline rocks. Firstly, he comments on the fact that between the end of the Trias and the beginning of Raritan sedimentation in the Upper Cretaceous, a period of about 75 million years, more than 5000' of Triassic sediments were removed from considerable areas of Connecticut. The drainage system which accomplished this removal would surely have become adjusted to structures in pre-Triassic rocks as these were revealed. A similar argument can be presented in the Maritime Provinces. Secondly, Flint notes that the similarity in the direction and slope of the well-defined Fall Zone surface of Connecticut and the surface beneath the coastal-plain sediments suggests that post-Cretaceous erosion has not achieved much in the way of altering the form of the Fall Zone surface, and that major rivers crossing it occupy much the same positions as in pre-Raritan (Upper Cretaceous) times. The rivers must, therefore, have existed before the Upper Cretaceous and were not superimposed. Thirdly, rivers crossing the Fall Zone surface are adjusted to a substantial degree to underlying structures and thus occupy valleys which were originally part of the Fall Zone (pre-Upper Cretaceous) surface.

Flint also notes that streams do not show a marked increase in gradient as they pass over the Fall Zone in Connecticut, as they do farther south in the Middle Atlantic States. This, he says, may be due to glacial erosion, but could possibly be due to an "earlier date of uplift and stripping of the coastal-plain strata from the Fall Zone than in the states farther south", thus "allowing a longer time for readjustment of stream profiles" (Flint, 1963). This corroborates Engelen's (1963) conclusions from the continental margin evidence to the effect that graben formation in the continental margin (and concomitant uplift of the adjacent land areas) occurred in the early Cretaceous off Newfoundland and spread farther soutwest during later Cretaceous and Cenozoic times.

Therefore, from evidence provided by the geologic and tectonic history of the continental margin of northeastern North America, and from the morphology of an adjacent land area whose geomorphic history is likely to have been similar, in its broad aspects, to that of the Atlantic

Provinces, it seems probable that the land surface of the latter area has undergone subaerial erosion throughout at least Cretaceous and Cenozoic times.

Some earlier physiographic studies in the Atlantic Provinces have concluded that the earliest erosion surface found in that area is of early Tertiary age since, if it was of earlier date, it would not be so well preserved. This argument has been used by Alcock (1928) for the well-preserved peneplain surface(s) between 3700 and 4300 feet atop the Shickshock Mountains of the Gaspe Peninsula, Quebec. Goldthwait (1924), however, postulated completion by the close of the Cretaceous of the cycle which produced his "ancient plain" or "Atlantic Upland" (p. 40). This latter date would seem to be more in accord with the conclusions reached above and can be extended to include the highest erosion surface(s) of western Newfoundland.

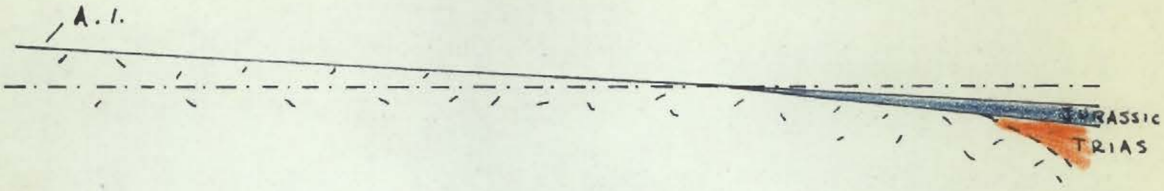
Drainage on this early land surface is generally agreed to have been towards the south and southeast from a divide in the north of the region, possibly on the north side of the present St. Lawrence estuary, and traces of this consequent drainage are seen at present in the transverse courses of the major rivers of the Maritime Provinces, such as the upper St. John. In western Newfoundland, traces of southeasterly-directed drainage at an early stage have been found in the southwestern Long Range Mountains and reference has been made to drainage both to northwest and southeast from a divide along the Bay of Islands Mountains and the Northern Peninsula at the time of the formation of the highest erosion surfaces in those areas. Later adjustment of drainage to geological structures, which in Newfoundland trend approximately at right angles to the proposed early drainage, has eradicated any trace of such drainage from the present pattern. The existence of a major drainage line across the now submerged floor of the Gulf of St. Lawrence flowing through Cabot Strait seems well evidenced. In western Newfoundland most streams would, during earlier stages, have been tributary to this "Laurentian River".

With the conclusion reached that the uppermost erosion surfaces of western Newfoundland probably are of greater age than surfaces occupying similar positions in the Appalachian Mountains of the Middle Atlantic States, the actual ages of erosion surfaces in the former area must now be considered, and comparisons made of geomorphic evolution in both areas.

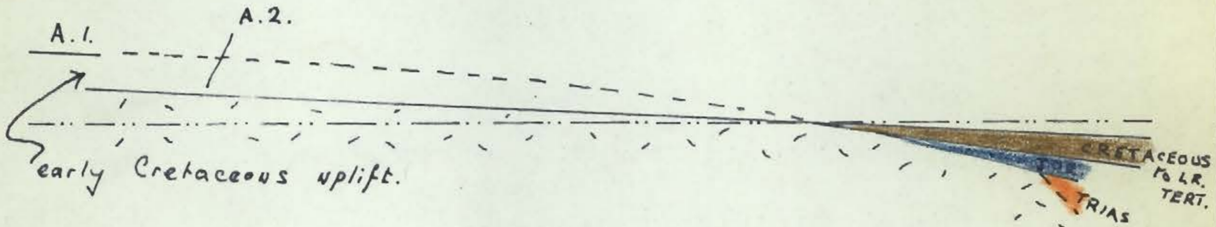
Assuming, from conclusions previously reached, that Newfoundland was uplifted in the early Cretaceous, a land surface must have been produced by denudation in the period between the end of the "Appalachian Revolution" of Carbo-Permian times and the time of this uplift. Call this surface A.I. (fig. 11.1). The development of a land surface of low relief in the Middle Atlantic States would have progressed uninterrupted through early Cretaceous times, from the end of the Carbo-Permian earth movements, since no early Cretaceous uplift is evidenced in that region. Call this surface B.I (fig. 11.2 I). Its development would have continued until it was submerged beneath the transgressive Upper Cretaceous seas (fig. 11.2,II). This is the surface which, in the eastern United States, is known as the

Figure. II.1. . Proposed Evolutionary Scheme for Landsurfaces of Atlantic Provinces of eastern Canada.

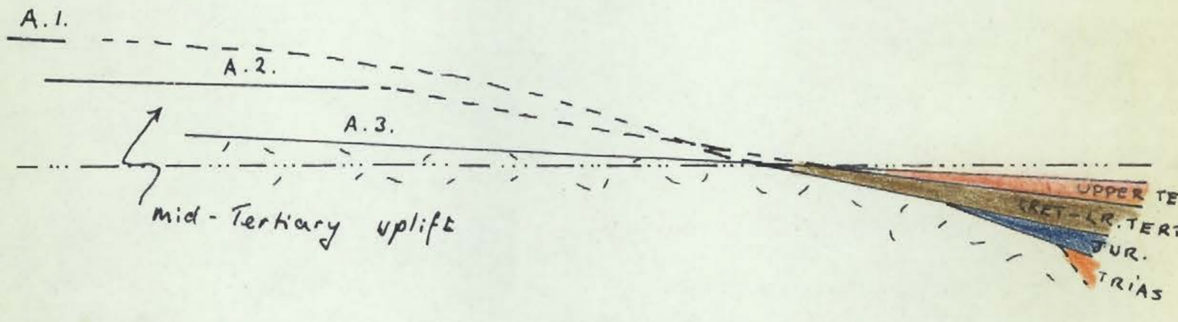
I. Development of pre-Early Cretaceous landsurface.



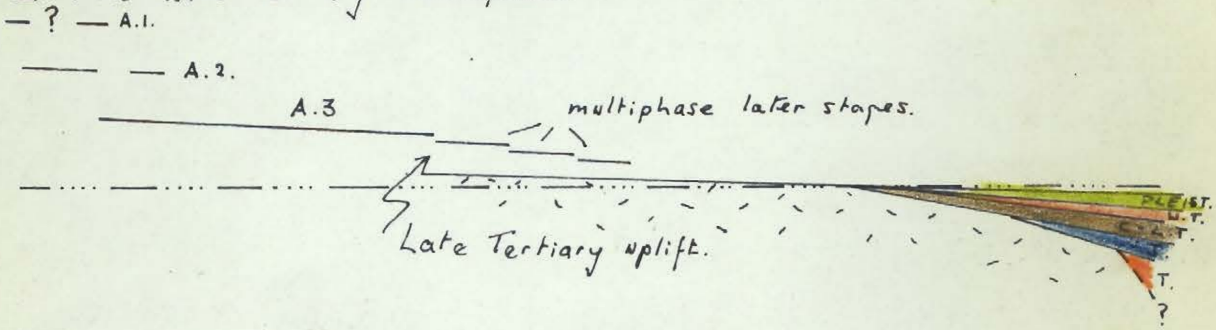
II. Development of Cretaceous - Early Tertiary landsurface.



III Development of Late Tertiary landsurface.



IV. Post-Late Tertiary Development



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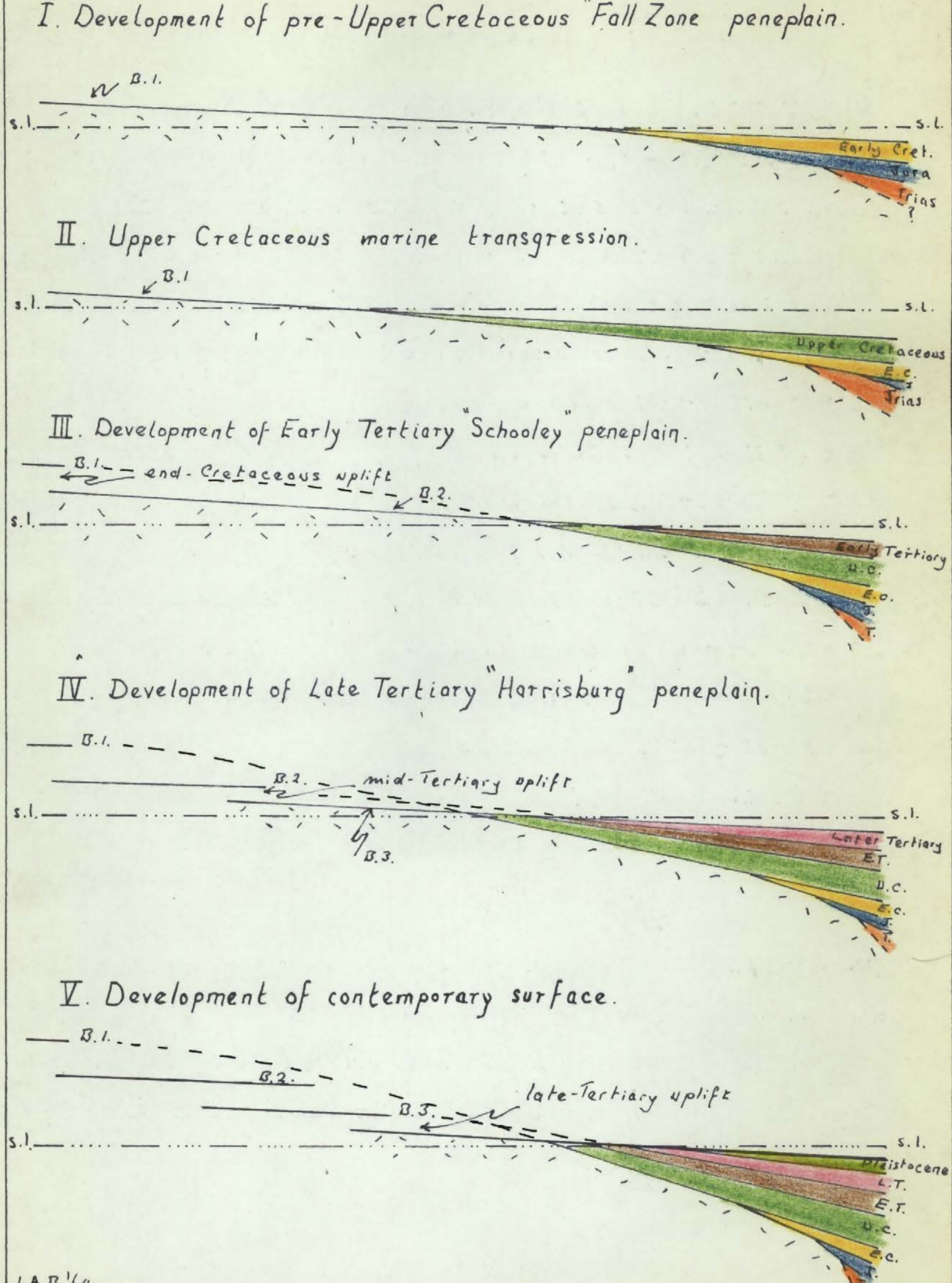
Fall Zone Peneplain. Brown (1961) has correlated this surface with the "Upland Surface" at above 3000' in the Catskill Mountains below which the younger Schooley Peneplain is developed.

In Newfoundland, the proposed early Cretaceous uplift initiated a new cycle of erosion and the post-Appalachian - pre-early Cretaceous surface, A.I., was upraised. This new cycle progressed through the later Cretaceous period, but no certain date can be fixed for its termination by uplift or a fall in base-level. In the eastern United States uplift occurred at the close of the Cretaceous to expose the Upper Cretaceous sea floor to a new cycle of denudation which resulted in the formation of the Schooley Peneplain (fig. 11.2 III). This uplift may possibly have been due to similar causes as the early Cretaceous uplift of Newfoundland. This latter uplift has previously been suggested to be contemporaneous with the graben formation in the continental margin, which progressed southwestwards during the later Cretaceous and early cenozoic. Since graben formation occurred immediately prior to the Upper Cretaceous off the northeastern United States, it seems likely that, by the close of the Cretaceous it would have reached the Middle Atlantic Continental margin and would have been accompanied by land uplift in that area, an uplift which caused the withdrawal of the Upper Cretaceous sea.

Lack of evidence of a similar end-Cretaceous uplift in Newfoundland forces the assumption of continued land surface development throughout the period following the early Cretaceous uplift, until relative uplift in the mid-Cenozoic arrested this development. No direct evidence of this uplift

Figure 11.2. Landsurface Evolution in the eastern United States (after D.W. Johnson, 1931).



is found in Newfoundland, but there is a strong argument for the eustatic nature of a mid-Cenozoic fall in base-level which affected the eastern United States. If this fall arrested the development of land surface A.2 (fig. 11.1, III) in Newfoundland, which began following the early Cretaceous uplift, then the development of A.2 was contemporaneous both with the Upper Cretaceous marine transgression of the eastern United States <u>and</u> the development of the Schooley Peneplain, which was also arrested by the mid-Cenozoic fall in base-level. The Schooley Peneplain can be called B.2. in this scheme (fig. 11.2, III), but it must be remembered that its development extended only over the latter part of the period in which surface A.2 was developed in the northeastern Atlantic seaboard.

The mid-Cenozoic fall in base-level initiated a new cycle in the eastern United States, which resulted in the formation of the Late-Cenozoic <u>Harrisburg Peneplain</u> (B.3, figure 11.2, IV) on weak rocks in the Great Valley west of the Appalachian Blue Ridge and over the Piedmont Province bordering the Coastal Plain. If this fall in base-level was eustatic, similar effects may be expected in Newfoundland and the Maritime Provinces. This Late Cenozoic cycle probably would not have sufficient time to develop a complete erosion surface (A.3, fig. 11.1, III) on any but the weaker rocks. This cycle would have been arrested by the pronounced uplift which is generally agreed to have affected the whole area of the Atlantic Provinces and the eastern margin of the Canadian Shield in the late Pliocene.

In summary, these conclusions, which have been arrived at by inductive reasoning are as follows:

(1) In Newfoundland (and probably the entire Atlantic Provinces), pre-early Cretaceous erosion cycle(s) produced surface(s) A.I (fig. 11.1, I).

(2) Early Cretaceous uplift initiated a new cycle which proceeded through the Cretaceous and early Tertiary to produce surface A.2 (fig.11.1,II).

(3) In the eastern United States erosion continued uninterruptedly from pre-early Cretaceous to pre-Upper Cretaceous times, producing surface B.I., the Fall Zone Peneplain (fig. 11.2, I).

(4) Surface B.I. in the eastern United States was submerged beneath the Upper Cretaceous marine transgression and covered by these deposits (fig. 11.2, II).

(5) End-Cretaceous uplift in this latter area initiated the Schooley cycle upon the Upper Cretaceous sea floor and surface B.2 was produced (fig, 11.2, III).

(6) A mid-Cenozoic fall in base-level throughout the eastern North American seaboard initiated a new cycle which produced surfaces A.3 and B.3 in Newfoundland and the eastern United States, respectively (fig. 11.1, III and fig. 11.2, IV).

(7) End-Tertiary uplift initiated a new cycle which proceeded to the stage of youth in the mountains of the Atlantic Provinces before Pleistocene glaciation of the region and produced the Somerville Peneplain in the eastern United States (fig. 11.1, IV and fig. 11.2, V).

In this scheme it can be seen that there were three major erosion cycles in Newfoundland. The following surfaces were produced:

I. A post-Appalachian to pre-early Cretaceous surface, A.I, equivalent in age to the Fall Zone Peneplain of the eastern United States.

II. A Cretaceous - early Cenozoic surface, A.2, the later-formed parts of which are equivalent in age to the Schooley Peneplain.

III. A later Cenozoic surface, A.3, equivalent in age to the Harrisburg Peneplain.

Erosion surfaces are found in western Newfoundland at the following elevations:

2350-2500'

2100-2300'

1900-2150'

1750-1900'

1550-1700'

1250-1500'

1100-1350'

1000-1150'

850-1000'

750-850'

It has long been thought that an erosion surface exists above the Schooley peneplain (s.s.) in the eastern United States and its equivalent in the New England states and the Quebec Appalachians. In the areas of the Appalachians where the denudation chronology was first established by Johnson (1931) this higher surface has been correlated with the Fall Zone Peneplain of pre-Upper Cretaceous age (Brown, 1961). Traced through the Appalachians as far northeast as Newfoundland, the Schooley peneplain can be expected to be better developed since, from the conclusions arrived at above, subaerial erosion continued into the early Cenozoic, while it was arrested by the Upper Cretaceous marine transgression farther southwest.

The question now arises as to which surface in the northern Appalachians is equivalent in its position within the denudation chronology to the Schooley. Goldthwait (1913) referred to a Cretaceous peneplain found throughout the Appalachians from Alabama to New Brunswick which was raised and warped in the early Cenozoic (end Cretaceous?) Johnson (1931) believed, with Davis, that the Schooley and New England peneplains are the same and are of Cenozoic age. Atwood (1940) correlated the oldest and highest surface of the New England - Acadian region with the New England peneplain of Davis and noted monadnocks above this in the White and Green Mountains. He referred the general level of those mountains and that of the Shickshock and Long Range Mountains to the Schooley surface. A comprehensive review of erosion surfaces in the northern Appalachians has been made by Sangree (1953).

It appears, from the concensus of opinion, that the erosion surface found throughout the northern Appalachian region of the eastern United States and the Atlantic Provinces of eastern Canada at approximately 2000' is the equivalent of the Schooley peneplain of the central

Appalachians. As previously mentioned, its development continued until it was arrested by the mid-Cenozoic fall in base-level which probably was eustatic. This surface (A.2, fig. 11.1, II) may be represented in western Newfoundland by the 2000' summit surface of the Long Range Mountains and the spurs and valley benches at this elevation in the higher mountains of the Bay of Islands and the Northern Peninsula. It also appears correlative with the "Carleton Surface" of Bird (personal communication, 1964) at 2050-2300'.

In western Newfoundland, this study has demonstrated the existence of a group of erosion surfaces at between 2000' and 2500'. This has not been correlated with the Long Range summit surface, since the equivalent of this has been found below it. This raises the question of whether or not this highest group of surfaces represents the pre-early Cretaceous surface whose existence was earlier postulated. If this is so, the difference in elevation between the higher group and the Long Range surface is not as great as would be expected, considering that the uplift which separated the two cycles was that early Cretaceous uplift which is herein postulated to have raised the Newfoundland land area above the influence of the Upper Cretaceous marine transgression. In the Gaspe Peninsula, the 3000-4300' surface atop the Shichshock Mountains lies 2000-2300' above the 1500-200' surface which may be correlated with the Long Range and Carleton surfaces. The Shickshock summit surface may therefore represent the pre-early Cretaceous surface. The fact that, in the Gaspe Peninsula, the surface at 1500-2000' is peripheral to the higher one may probably be accounted for both by the relatively great distance from that area to base-level compared to areas where the surface is found as the upper summit level, and by the greater initial elevation of that area. It does not seem that in Newfoundland the pre-early Cretaceous surface is to be found in the present landscape. This could possihly be due to its destruction during the process of adjustment of the drainage from the consequent south - or southeast-flowing pattern to the pronounced southwest to northeast-trending structures during the Cretaceous-early. Cenozoic cycle. That streams in Newfoundland are usually well adjusted to structures may be a reflection of the great length of this cycle in which the Long Range summit surface was formed.

What, then, is the position of the surfaces between 2000' and 2500'? They could be considered to have formed during this Cretaceous-early Cenozoic cycle, at an earlier date than the 2000' Long Range summit surface, and have had all traces removed from above the latter. In this case, the Cretaceous-early Cenozoic cycle must be extended back in time to include the development of the 2000-2500' surfaces. Bird's "Carleton Surface" lies at 2050-2300' and has been here correlated with the Long Range summit surface.

What of lower surfaces? Surface A.3 (fig. 11.1, III) was formed during the later Tertiary and is likely to be represented in Newfoundland by the well-developed plateau surfaces between 1200' and 1700'. These surfaces occur at 1150-1350', 1250-1500' and 1550-1700'. Since these surfaces are so well developed across hard rocks, it may be questioned whether the cycle in which they developed is equivalent to the Harrisburg cycle of the eastern United States. In the latter area, during this cycle, rivers succeeded only in cutting surfaces in weaker rocks. Possibly, greater relief and greater erosive power of the rivers, greater rainfall or the assistance of pronounced structural weaknesses not found in the American area, can account for this. Surfaces within this group are found all over the Maritimes. Particular reference can be made to the equivalence of the 1550-1700' surface in the Cape Anguille Mountains of southwestern Newfoundland to the 1525-1700' surface which Bird (personal communication, 1964) recognises in the Cape Breton Highlands, Nova Scotia. In this area, as well as in others throughout the Maritimes, lower surfaces of this group above 1000' find equivalents in western Newfoundland. The 1200-1700' surfaces therefore, appear to be equivalent in age to the late Cenozoic Harrisburg peneplain of the eastern United States.

The place of the lower group of surfaces at 750-850', 850-1000' and 1000-1150', must now be considered. These were also affected by uplift of a few hundreds of feet and, thus, their development would seem to belong in the late Cenozoic cycle. This is possible when it is considered that they are best developed in the areas of softer Ordovician sedimentary rocks and appear only as minor valley benches on harder rocks, for example, in the valley of Flat Bay Brook. They were certainly formed prior to the last Pleistocene (Wisconsin) glaciation since they have been modified by glacial erosion in recent times. They could not have formed during Pleistocene interglacials, since such features would be cut in glacio-fluvial drift rather than in solid rock.

In summary, then, both the highest group of erosion surfaces between 2000' and 2500' in the Bay of Islands Mountains and the Northern Peninsula Long Range Mountains and the 2000' Long Range summit surface in those mountains to the southeast appear to have been formed during the Cretaceous - early Cenozoic cycle. In the late Cenozoic cycle the group of well-developed plateau surfaces at between 1300' and 1700' were formed and, in the later stages of that cycle, prior to the major uplift of the area at the close of the Tertiary period, the lowest surfaces between 750' and 1150' were formed.

The erosion surfaces found in western Newfoundland appear to be separated by vertical intervals which are small, relative to the height ranges between them. In order to test the validity of elevation as a criterion for correlation of the surfaces, a simple statistical technique was applied. In Table IV are listed the elevations of the surfaces, their height ranges, their average elevations and the vertical intervals between these average elevations. The arithmetic mean of the height ranges is 180' and that of the vertical intervals between the mean heights of the surfaces is 175'. Thus, as the variation among the surfaces is greater than that between them, elevation alone does not appear to be a valid criterion for erosion surface correlation. Elevation, however, was not the only criterion used. The important additional criterion of the continuity of the surfaces in the field area has been used to good effect. Indeed, throughout the investigation, the position which a surface occupies in the landscape, that is its mode of occurrence and its relationship to surfaces above and below it, has been found to be a most useful criterion for correlation.

(all figures in feet)

TABLE IV

Statistical Analysis of Erosion Surfaces in Western Newfoundland

Surface	Height range	Average elevation	Vertical interval
2350-2500	150	2425	
			225
2100-2300	200	2200	
			175
1900-2150	250	2025	
			150
1750-1900	150	1825	
			200
1550-1700	150	1625	
			250
1250-1500	250	1375	
			150
1100-1350	250	1225	
			150
1000-1150	150	1075	
			150
850-1000	150	925	
			125
750-850	100	800	
			125

Arithmetic mean of height ranges ....180'

Arithmetic mean of vertical intervals ...175'

The largest vertical interval between the average elevations of the surfaces is 250', although the upper limit of one surface and the lower limit of the surface above never differ by more than 50', and often are at the same elevation, or may even overlap.[2] Even between the surfaces which are thought to have developed before and after the mid-Cenozoic uplift there is a difference in elevation no greater than the average vertical interval. This may cast doubt upon the validity of assigning the surfaces above and below about 2000' to two different erosion cycles. However, the major criterion used in fixing the position of the uppermost of the lower surfaces (the 1550-1700' surface) within the denudation chronology is the position of this surface in the landscape, both its mode of occurrence as widespread plateaux and its relationship to the 2000' summits and the less widely developed surfaces below.

The relatively small vertical intervals between the surfaces has previously suggested that periodic isostatic uplift has been the predominant mode of tectonism which has interrupted periods of erosion surface development. That such small vertical intervals characterize the whole suite of surfaces suggests that periodic isostatic response occurred at least from the beginning of the Cretaceous (if the conclusions reached as to the age of the surfaces are valid), until the late Cenozoic uplift.

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2. It must be remembered that the figures given for the elevations of the surfaces represent those elevations between which the surfaces lie throughout the field area and not in any one locality, where the height range may be 100' less.

This study has demonstrated the existence in western Newfoundland of a suite of erosion surfaces between 2400 feet and 700 feet. While no attempt has been made to explain the detailed exogenetic and endogenetic processes active in their formation, the evidence herein presented will, it is hoped, form the basis for a sounder conceptual approach to the study of these features.

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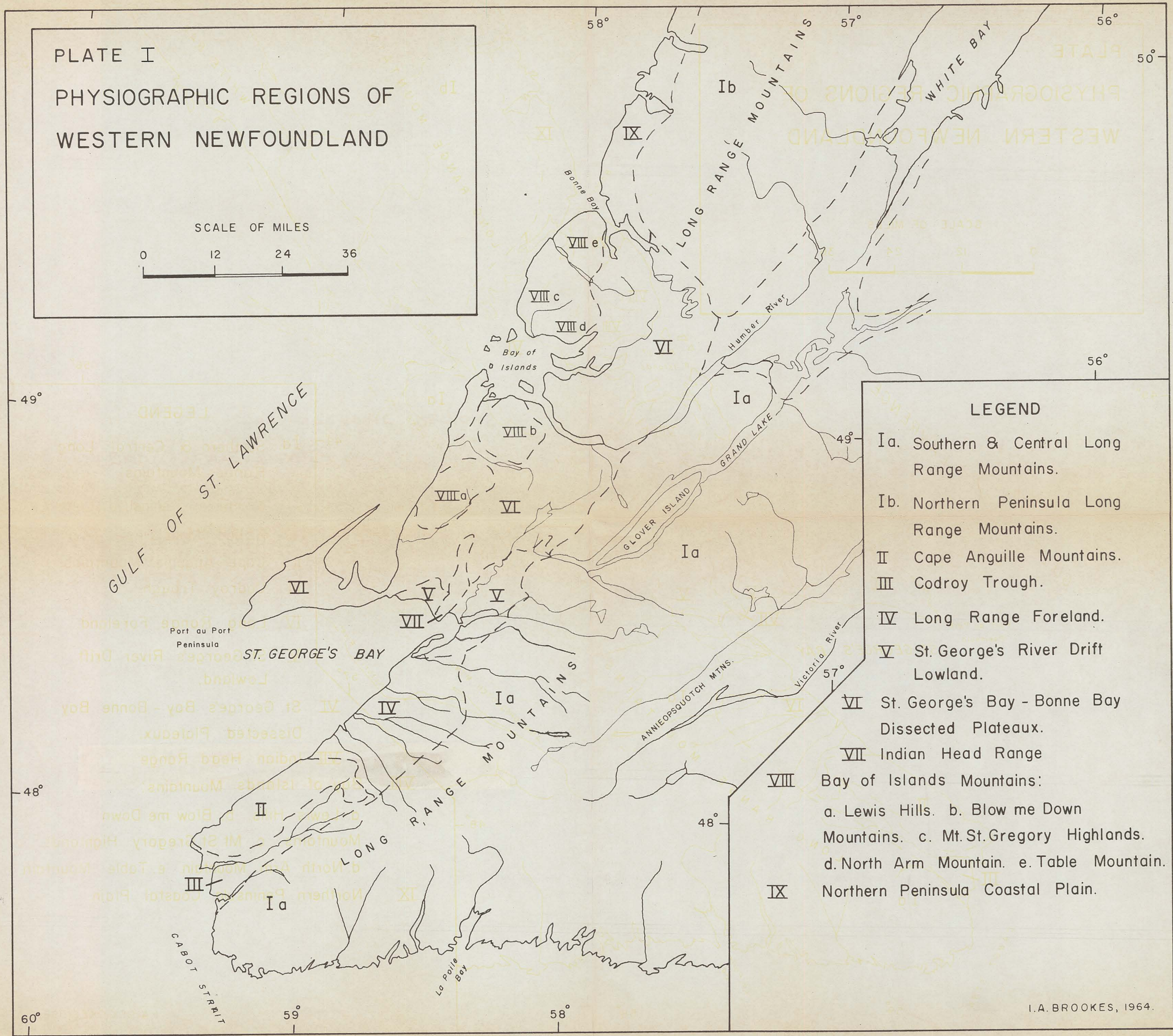


PLATE I
PHYSIOGRAPHIC REGIONS OF
WESTERN NEWFOUNDLAND
SCALE OF MILES
0
12
24
36
LEGEND
Ia. Southern & Central Long Range Mountains.
Ib. Northern Peninsula Long Range Mountains.
II Cape Anguille Mountains.
III Codroy Trough.
IV Long Range Foreland.
V St. George's River Drift Lowland.
VI St. George's Bay - Bonne Bay Dissected Plateaux.
VII Indian Head Range
VIII Bay of Islands Mountains:
a. Lewis Hills. b. Blow me Down Mountains. c. Mt. St. Gregory Highlands. d. North Arm Mountain. e. Table Mountain.
IX Northern Peninsula Coastal Plain.
GULF OF ST. LAWRENCE
LONG RANGE MOUNTAINS
WHITE BAY
Bonne Bay
Bay of Islands
Humber River
GRAND LAKE
GLOVER ISLAND
Port au Port Peninsula
ST. GEORGE'S BAY
ANNIEOPSQUOTCH MTNS.
Victoria River
CABOT STRAIT
La Poile Bay
Ia
Ib
II
III
IV
V
VI
VII
VIII a
VIII b
VIII c
VIII d
VIII e
IX
50°
49°
48°
60°
59°
58°
57°
56°
I.A. BROOKES, 1964.

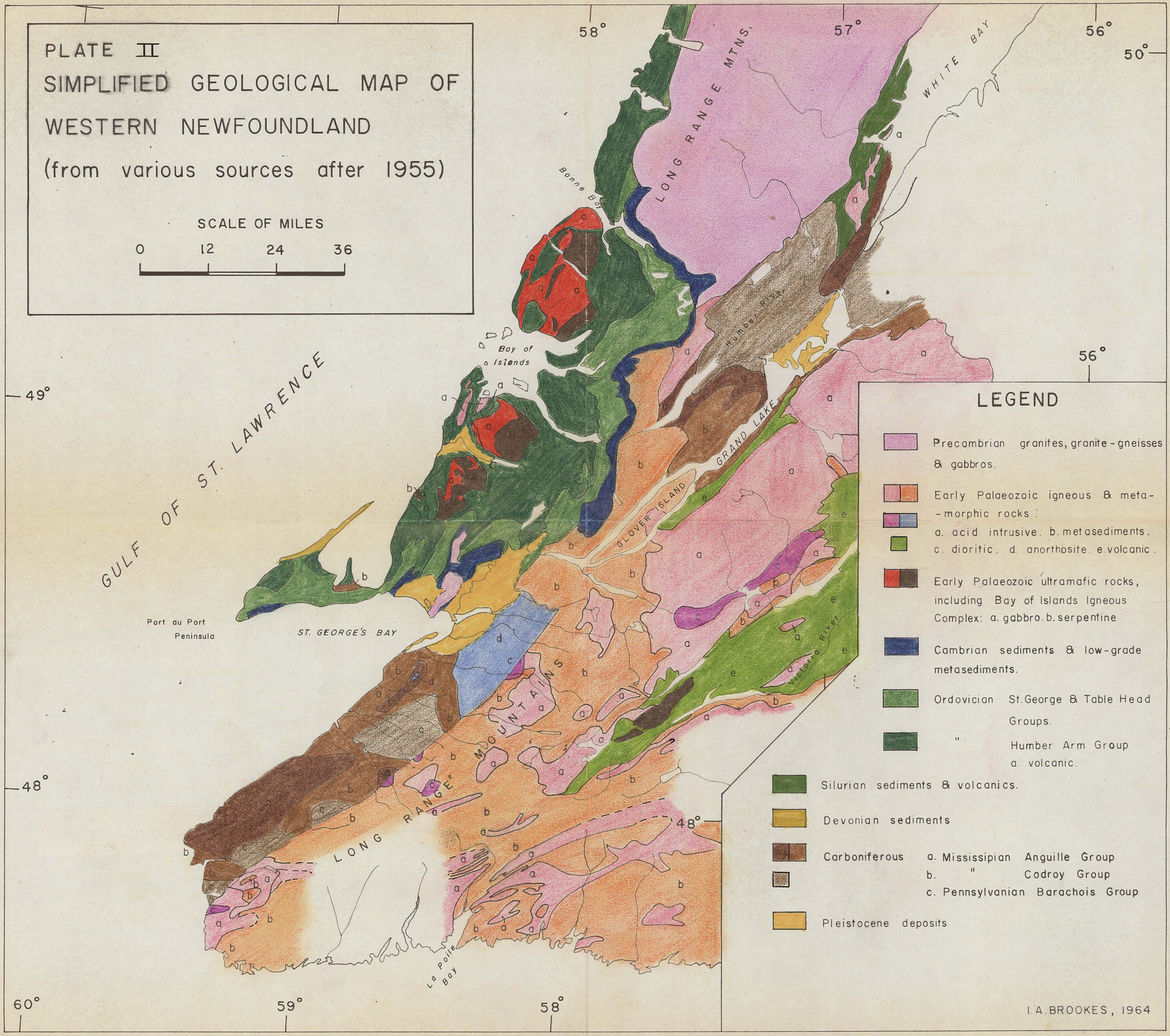
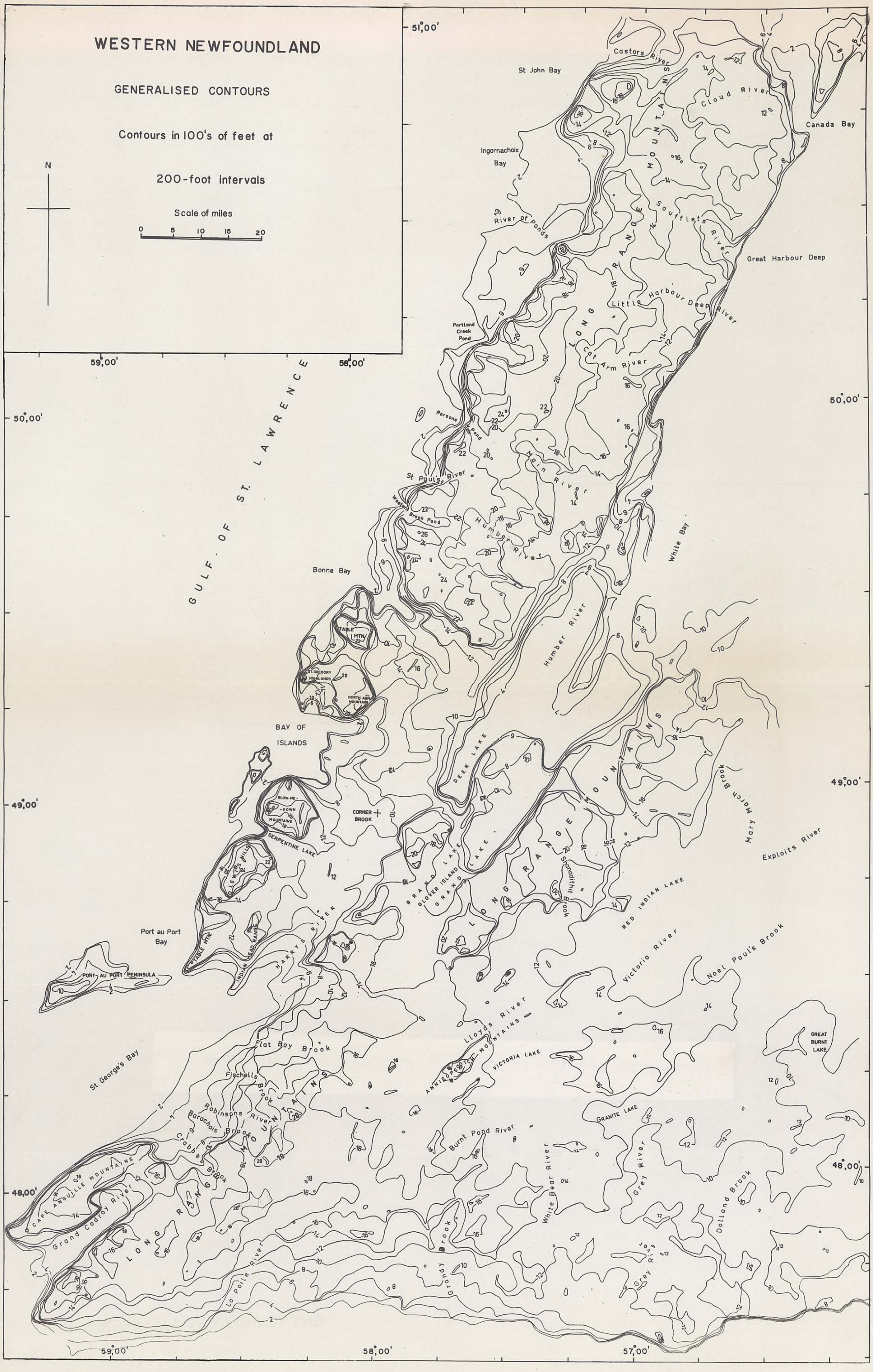


PLATE II
SIMPLIFIED GEOLOGICAL MAP OF
WESTERN NEWFOUNDLAND
(from various sources after 1955)
SCALE OF MILES
0 12 24 36
LEGEND
Precambrian granites, granite-gneisses & gabbros.
Early Palaeozoic igneous & meta-morphic rocks:
a. acid intrusive. b. metasediments.
c. dioritic. d. anorthosite. e. volcanic.
Early Palaeozoic ultramafic rocks, including Bay of Islands Igneous Complex: a. gabbro. b. serpentine
Cambrian sediments & low-grade metasediments.
Ordovician St.George & Table Head Groups.
" Humber Arm Group
a. volcanic.
Silurian sediments & volcanics.
Devonian sediments
Carboniferous a. Mississipian Anguille Group
b. " Codroy Group
c. Pennsylvanian Barachois Group
Pleistocene deposits
GULF OF ST. LAWRENCE
LONG RANGE MTNS.
WHITE BAY
Bonne Bay
Bay of Islands
Humber Rvr.
GRAND LAKE
GLOVER ISLAND
Victoria River
Port au Port Peninsula
ST. GEORGE'S BAY
LONG RANGE MOUNTAINS
La Poile Bay
58°
57°
56°
50°
49°
48°
60°
59°
I.A. BROOKES, 1964

END PLATE THREE: GENERALISED CONTOURS OF WESTERN NEWFOUNDALND


WESTERN NEWFOUNDLAND
GENERALISED CONTOURS
Contours in 100's of feet at
200-foot intervals
N
Scale of miles
0 5 10 15 20
51°00'
50°00'
49°00'
48°00'
59°00'
58°00'
57°00'
GULF OF ST. LAWRENCE
St John Bay
Castors River
Cloud River
Canada Bay
Ingornachoix Bay
River of Ponds
Soufflets River
Great Harbour Deep
Little Harbour Deep River
Portland Creek Pond
Cat Arm River
LONG RANGE MOUNTAINS
Parsons Pond
St. Pauls River
Main River
Western Brook Pond
Humber River
White Bay
Bonne Bay
TABLE MTN
ST GREGORY HIGHLANDS
NORTH ARM MOUNTAIN
BAY OF ISLANDS
BLOW-ME-DOWN MOUNTAINS
CORNER BROOK
SERPENTINE LAKE
LEWIS HILLS
DEER LAKE
GRAND LAKE
GLOVER ISLAND
Sandy Lake
Mary March Brook
Exploits River
Shoudithit Brook
RED INDIAN LAKE
Port au Port Bay
TABLE MTN
INDIAN HEAD RANGE
PORT AU PORT PENINSULA
HARRY'S RIVER
Victoria River
Noel Paul's Brook
Lloyds River
ANNIEOPSQUOTCH MOUNTAINS
VICTORIA LAKE
GREAT BURNT LAKE
St. George's Bay
Flat Bay Brook
Fischells Brook
Robinsons River
Barachois Brook
Crabbes Brook
GRANITE LAKE
Burnt Pond River
White Bear River
Grey River
CAPE ANGUILLE MOUNTAINS
Grand Codroy River
Dolland Brook
Grandy Brook
La Poile River
Grey River

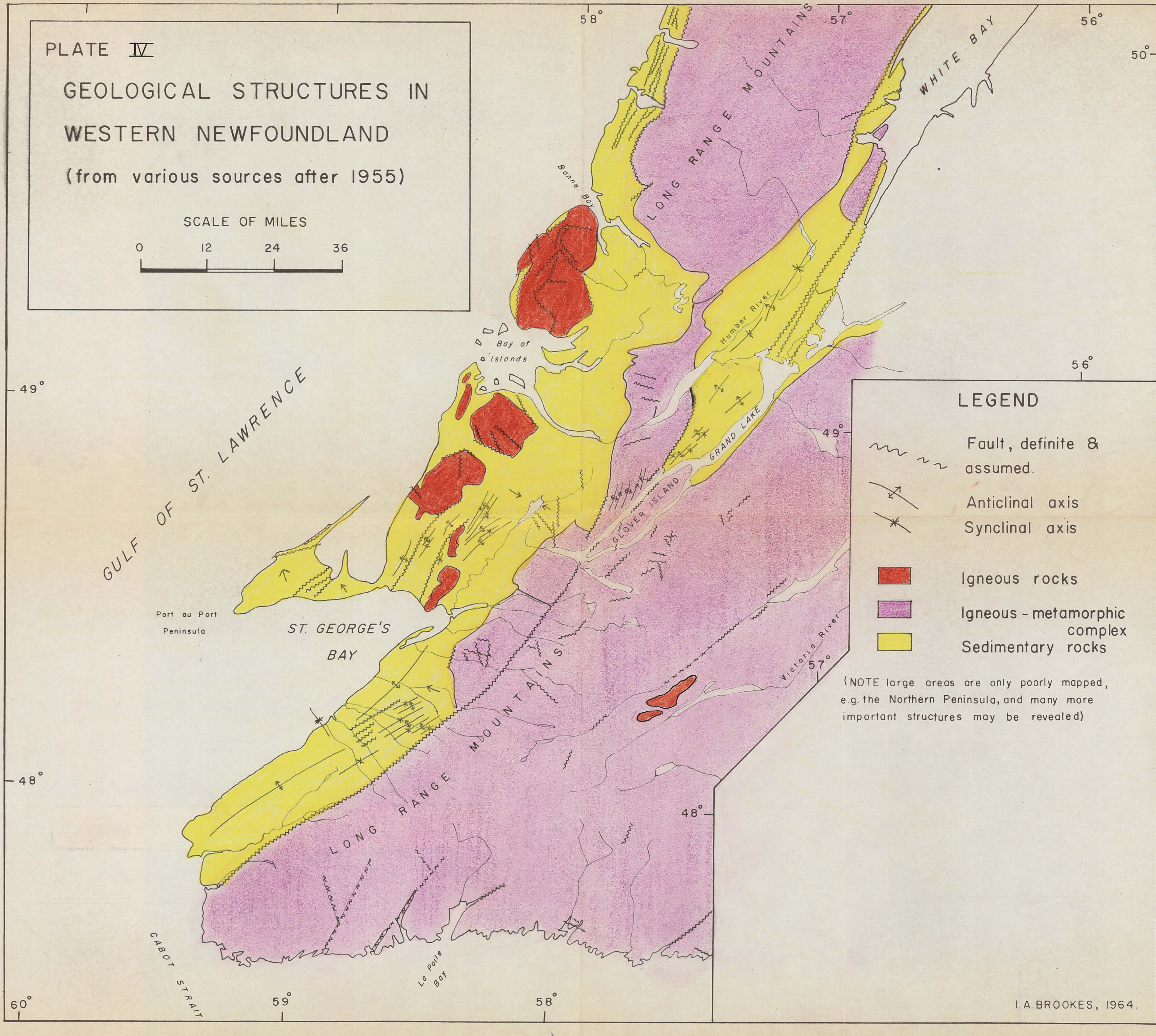


PLATE IV

GEOLOGICAL STRUCTURES IN WESTERN NEWFOUNDLAND

(from various sources after 1955)

END PLATE FIVE: SUPERIMPOSED COMPOSITE PROFILES ACROSS THE CAPE ANGUILLE AND LONG RANGE MOUNTAINS, NORTHWEST TO SOUTHEAST


CAPE ANGUILLE MOUNTAINS
CODROY VALLEY
LONG RANGE MOUNTAINS
N.W.
S.E.
2000
1500
1000
500
0
feet above sea level
Gulf of
St.Lawrence
0
5
10
BASE LINE
15
20
25
30
35
40
MILES


CAPE ANGUILLE MOUNTAINS
CODROY VALLEY
LONG RANGE MOUNTAINS
N.W.
Gulf of
St.Lawrence
S.E.
Gulf of
St.Lawrence
2000
1500
1000
500
0
feet above sea level
0
5
10
BASE LINE
15
20
25
30
35
40
MILES

END PLATE SIX: SUPERIMPOSED COMPOSITE PROFILES ACROSS THE LONG
RANGE FORELAND AND MOUNTAINS, NORTHWEST TO SOUTHEAST


LONG RANGE FORELAND
LONG RANGE MOUNTAINS
HIGH CENTRAL PLATEAU
N.W.
S.E.
feet above sea level
2000
1500
1000
500
0
MILES
5
10
BASE LINE
15
20
25
30
35
40
45


N.W.
Gulf of
St. Lawrence
LONG RANGE FORELAND
LONG RANGE MOUNTAINS
LA POILE RIVER
S.E.
feet above sea level
2000
1500
1000
500
0
MILES
5
10
BASE LINE
15
20
25
30
35
40

END PLATE SEVEN: DRAINAGE PATTERN OF WESTERN NEWFOUNDLAND, MAJOR DIVIDES AND GEOLOGICAL STRUCTURES

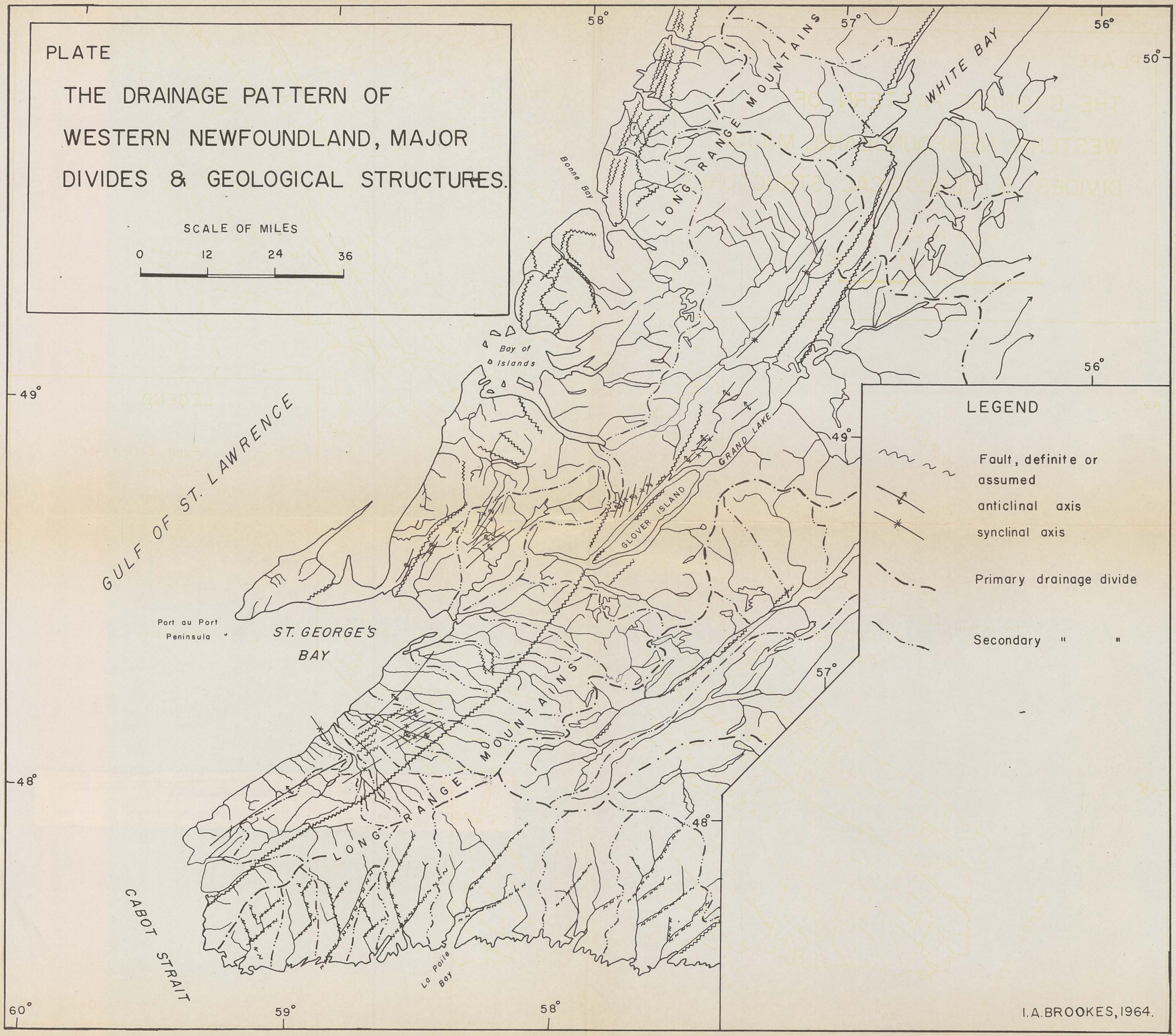


PLATE
THE DRAINAGE PATTERN OF
WESTERN NEWFOUNDLAND, MAJOR
DIVIDES & GEOLOGICAL STRUCTURES.
SCALE OF MILES
0
12
24
36
LEGEND
Fault, definite or assumed
anticlinal axis
synclinal axis
Primary drainage divide
Secondary " "
GULF OF ST. LAWRENCE
Port au Port Peninsula
ST. GEORGE'S BAY
Bay of Islands
Bonne Bay
WHITE BAY
LONG RANGE MOUNTAINS
LONG RANGE MOUNTAINS
GRAND LAKE
GLOVER ISLAND
CABOT STRAIT
La Poile Bay
58°
57°
56°
50°
49°
48°
60°
59°
I.A.BROOKES, 1964.