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Evolution of an early Proterozoic continental margin: the Coronation geosyncline and associated aulacogens of the northwestern Canadian shield

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The Coronation geosyncline developed in the early Proterozoic along the western margin of a continental platform (the Slave Province) of Archaean rocks older than 2300 Ma, and culminated between 1725 and 1855 Ma ago with the emplacement of a pair of batholiths (the Bear Province). The evolution of the geosyncline has a strong family resemblance to Phanerozoic geosynclines believed to delineate ancient continental margins and have been controlled by global plate interactions. Such geosynclines are unknown in Archaean orogenic belts, from which it is inferred that creation of the first large rigid continental platforms marked the end of the Archaean and the beginnings of actualistic plate tectonics.

The geosyncline began with deposition of a westward-facing continental shelf, consisting of a lower formation dominated by orthoquartzite, derived from the platform, and an upper cyclic stromatolitic

dolomite formation. West of the shelf edge, the dolomite passes abruptly into a much thinner mudstone sequence with dolomite debris-flows, and the orthoquartzite into a thick laminated silt and mudstone sequence with quartzite turbidites. The oldest rocks west of the shelf edge, an area interpreted to have been a continental rise, are pillow basalts and volcanic breccias, extruded above a basement of unknown character.

The principal turning point in the evolution of the geosyncline came with the foundering of the continental shelf. It is draped by a thin laminated pyritic black mudstone sequence, overlain by a westward-thickening clastic wedge resulting from intrusion and erosion of the batholiths to the west. The clastic wedge begins with a thick sequence of coarse greywacke turbidites that passes eastward into concretionary mudstone on the platform. The mudstone grades upward into laminated shaly limestone with minor greywacke turbidites, overlain in turn by cross-bedded red lithic sandstone.

The supracrustal rocks of the geosyncline have been compressed and tectonically transported toward the platform. Adjacent to the batholithic belt, the continental rise and clastic wedge sequences are penetratively deformed and recrystallized by regional low-pressure metamorphism. To the east, the unmetamorphosed continental shelf and clastic wedge sequences have been flexurally folded and overthrust above a basal detachment surface. East of the thrust zone, relatively thin rocks on the platform are nearly flat-lying except around large anticlinal basement uplifts.

Unusual features of the platform are its two aulacogens – long-lived deeply subsiding fault troughs that extend at high angles from the geosyncline far into the interior of the platform. During every phase in the evolution of the geosyncline, the aulacogens received much thicker sedimentary sequences, commonly with the addition of basaltic volcanics, than adjacent parts of the platform. Although equal in thickness to the geosyncline, the aulacogens were never subjected to the batholithic intrusions, regional metamorphism or low-angle overthrusting characteristic of the geosyncline. The Athapuscow aulacogen, in the region of Great Slave Lake, is interpreted as having been an incipient rift, located over a crustal arch, during the continental shelf stage of the geosyncline, but sagged to become a crustal downwarp during the clastic wedge stage, ultimately with sufficient transverse compression to produce broad folds. Finally, the aulacogen became part of a regional transcurrent fault system, along which thick fanglomerates accumulated in local troughs.

The batholithic belt consists of two batholiths, eroded to different depths, separated by the north-trending 350 km long Wopmay River fault. The Hepburn batholith, east of the fault, is a composite intrusion of mesozonal granodiorite plutons. The foliated and migmatitic borders of the plutons are normally concordant with wall rock sheaths of sillimanitic paragneiss. Along the eastern margin of the batholith, metamorphosed rocks of the continental rise sequence dip gently to the west beneath the batholithic rocks. Belts of intensely deformed and metamorphosed supracrustal rocks within the batholithic terrain include sequences of pillow basalt, pelites and granite-pebble conglomerate, perhaps the lower part of the continental rise deposited during the initial rifting of the continental margin. The Great Bear batholith, west of the fault, consists of discordant epizonal plutons, mostly adamellite, that intrude broadly folded but regionally unmetamorphosed sequences of welded rhyodacitic ash-flow tuff, trachybasalt and derived sedimentary rocks. The volcanic rocks, intruded by dense dyke swarms radiating from the plutons and by felsite plugs, are interpreted to be comagmatic with the plutons.

Mapping is as yet insufficient to establish, speculations aside, the possible relations of the two batholiths to arc-trench systems. Furthermore, the western margin of the batholithic belt, a region of critical importance, is covered by a veneer of younger Proterozoic and Paleozoic sedimentary rocks. Until fossil arc-trench systems are outlined, the contention that the Coronation Geosyncline involved global plate interactions is based on indirect evidence – the analogous evolution of the geosyncline east of the batholithic belt with Phanerozoic geosynclines in which fossil arc-trench systems have been found.

Introduction

Plate tectonics has given geology the confidence, no doubt overconfidence but confidence none-the-less, that comes from having a set of rules to aid in reconstructing crustal history. The realization that geosynclines are features of continental margins whose evolutionary development is controlled by global plate interactions (Mitchell & Reading 1969; Dewey & Bird 1970; Dickinson 1971) promises that the rules of plate tectonics may be applied as far back in Earth history as geosynclines are found. The question is, 'How far back is that?'.

It seems that geosynclines do not occur in the Archaean. The Archaean 'greenstone belts' (Anhaeusser, Mason, Viljoen & Viljoen 1969; Salop & Scheinmann 1969; Pettijohn 1970) lack the fundamental asymmetry of geosynclines. Absent are the paired miogeosynclinal nad eugeosynclinal facies belts, the grand scale of foreland-directed overthrusting, and the paired high-pressure and low-pressure metamorphic belts of arc-trench systems. Instead of developing at the margins of stable platforms, the greenstone belts are surrounded by granitic terrains of such pervasive mobility as to suggest a pre-cratonic stage in crustal evolution during which time rigid crustal blocks, the *sine qua non* of plate tectonics, did not yet exist.

In the early Proterozoic, however, the Archaean rocks of the northwestern Canadian shield were stabilized to form an extensive continental platform, along the western margin of which developed the Coronation geosyncline. The depositional and structural evolution of the geosyncline has a strong family resemblance to Phanerozoic geosynclines. As with all geosynclines, the Coronation has its idiosyncrasies – specifically, long-lived deeply subsiding fault troughs, called 'aulacogens' (Salop & Schienmann 1969), that extend at high angles from the geosyncline far into the interior of the platform. But such idiosyncrasies do not detract from the overwhelming impression that, in the northwestern Canadian Shield, early Proterozoic tectonics are essentially like those of the Phanerozoic. Here, and perhaps elsewhere, the Archaean–Proterozoic transition is the greatest tectonic discontinuity in the 3800 Ma of earth history for which we have rocks.

Regional geologic setting

On the tectonic map of Canada (Stockwell 1968), the northwestern part of the Canadian shield is subdivided into the Slave, Churchill and Bear Provinces (see figure 1). Early Proterozoic supracrustal rocks doubtless once covered the entire region but are now preserved mainly in three structural basins – the Epworth basin southwest of Coronation Gulf, the Goulburn basin around Bathurst Inlet, and the Great Slave basin around the east arm of Great Slave Lake.

The Coronation geosyncline (see figure 2) developed as a belt of thick supracrustal rocks along the western margin of the Slave Province and is now best exposed in the Epworth basin. The geosyncline culminated with the emplacement of a belt of granitic batholiths in the Bear Province. Thick supracrustal rocks correlative with those in the geosyncline accumulated far to the east of the Bear Province in two aulacogens, now exposed in the Goulburn and Great Slave basins.

Archaean basement rocks

The Slave Province consists mainly of Archaean granitic and metamorphosed supracrustal rocks with K–Ar radiometric ages that cluster between 2300 and 2600 Ma. The supracrustal rocks (McGlynn & Henderson 1970) occur in tightly compressed synclinal belts (see figure 1), in which metamorphism of a low-pressure facies series increases in grade symmetrically toward the belt margins. The normal stratigraphic succession is submarine basalt, thousands of metres thick, overlain by perhaps equally thick pelitic rocks with greywacke turbidites. At the tops of the basalt piles are subordinate silicic volcanics, cross-bedded quartzites, cherty carbonates and conglomerates with cobbles of basalt, felsite and granite. In the few places where the base of the supracrustal succession is exposed, the underlying rocks are highly chloritized granites.

The supracrustal belts are separated by mesozonal batholiths (see figure 1) of homophonous to foliated granodiorite. Although the granodiorite intrudes the supracrustal rocks, the tendency of the basalt piles to be penecondordant with, and face away from the contacts suggests that the batholiths are mantled domes of remobilized basement from beneath the supracrustal rocks. Within the supracrustal belts are discordant epizonal plutons.

The Churchill Province is also underlain mainly by Archaean rocks, generally metamorphosed to a higher grade than is typical of the Slave Province. In addition, they have suffered extreme subsequent cataclastic and retrogressive metamorphism, producing a northeast structural grain and a broad range of K–Ar ages – from 1605 to 2460 Ma. Although 1700 to 1800 Ma

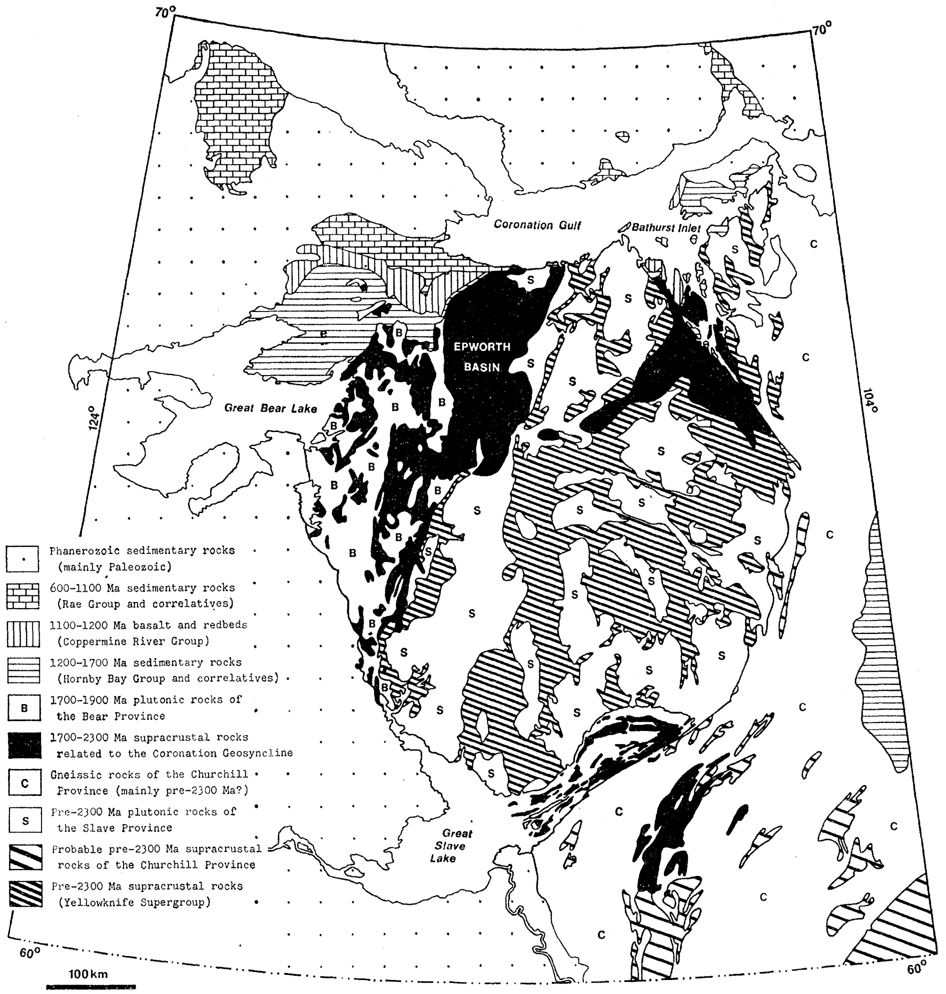

FIGURE 1. Geologic map of the northwestern corner of the Canadian Shield. The Epworth basin is located south of Coronation Gulf, the Goulburn basin south of Bathurst Inlet, the Great Slave basin in the east arm of Great Slave Lake, and the Nonacho basin southeast of Great Slave Lake (modified from Stockwell 1968).

plutonism characterizes other parts of the Churchill Province, the granitic rocks south of the Great Slave basin and east of the Goulburn basin are all believed to have been intruded in the Archaean.

Early Proterozoic supracrustal rocks

During the evolution of the Coronation geosyncline, most of the Slave Province was a continental platform on which the Proterozoic sedimentary rocks were thin and relatively little deformed. Thick successions accumulated only in the geosyncline along the western margin of the Slave Province and in the two aulacogens (see figure 2).

The Epworth basin (Fraser 1960; Fraser & Tremblay 1969; Hoffman, Fraser & McGlynn 1970; Hoffman, Geiser & Gerahian 1971) is a northerly-trending fold and thrust belt in which tectonic compression and transport, related to plutonism in the Bear Province, is directed eastward toward the Slave Province platform.

The Great Slave basin (Stockwell 1933, 1936; Stockwell *et al.* 1967, 1968; Hoffman 1968, 1969) and the Goulburn basin (Fraser 1964; Fraser & Tremblay 1969; Tremblay 1971) are far from the batholithic belt and lack the regional metamorphism and low-angle overthrusting

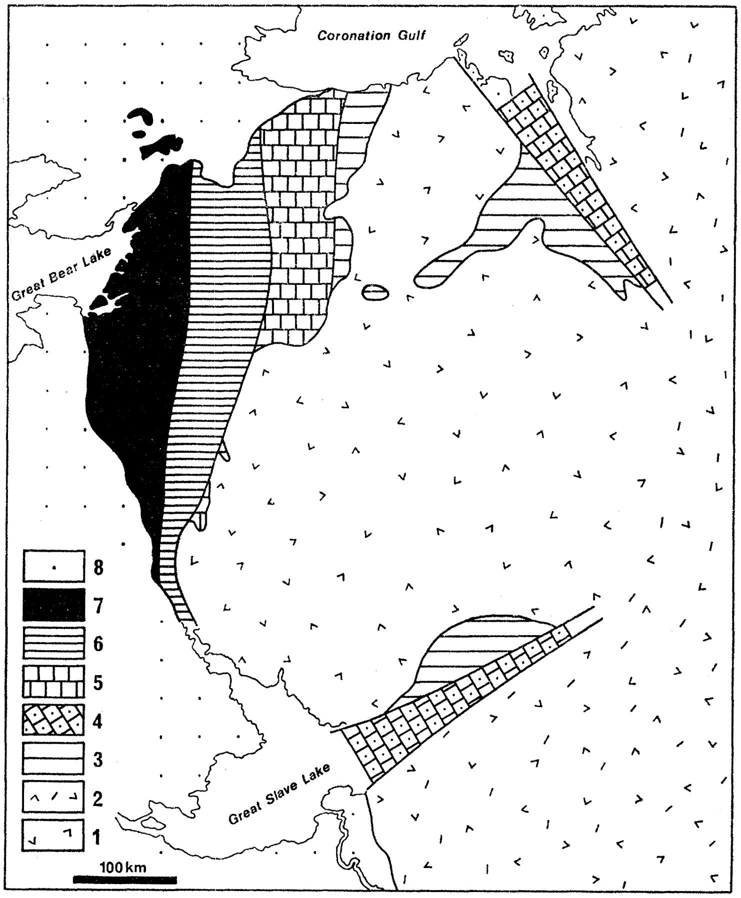

FIGURE 2. Tectonic subdivision of early Proterozoic rocks in the northwestern Canadian Shield.
Unit 1. Exhumed Archaean basement rocks.
Unit 2. Mainly exhumed Archaean basement rocks subjected to early Proterozoic catacliastic and retrogressive metamorphism.
Unit 3. Thin early Proterozoic supracrustal rocks of the continental platform.
Unit 4. Thick early Proterozoic supracrustal rocks deposited in aulacogens.
Unit 5. Thick early Proterozoic supracrustal rocks of the continental shelf and clastic wedge of the Coronation geosyncline.
Unit 6. Metamorphosed early Proterozoic supracrustal rocks of the continental rise and clastic wedge of the Coronation geosyncline intruded by the mesozonal Hepburn batholith.
Unit 7. Comagmatic granitic and volcanic rocks of the epizonal Great Bear batholith.
Unit 8. Middle Proterozoic and younger cover rocks.

of the Epworth basin. Both basins are associated with regional high-angle fault systems – the northeast trending McDonald fault system of Great Slave Lake and the southeast trending Bathurst fault system of Bathurst Inlet.

Stratigraphic correlations have been established between the three basins (see table 1). The Great Slave basin is unique in having remnants of an early Proterozoic succession (Wilson Island group), dominated by cross-bedded quartzite thousands of metres thick, that was intensely deformed before even the start of the Coronation geosyncline. The structural basins in the Churchill Province along the Taltson River fault system southeast of Great Slave Lake (see figure 1) contain supracrustal rocks probably correlative with the Et-then Group, the youngest in the Great Slave basin.

TABLE 1. EARLY PROTEROZOIC CORRELATION CHART

Epworth basin	Great Slave basin	Goulburn basin
unnamed	Et-then group unconformity	unnamed
Takiyuak Fm.	Christie Bay group	Brown Sound Fm.
Cowles Lake Fm.	Pethei group	Kuuvik Fm.
Recluse Fm.	Kahochella group upper Sosan group	Peacock Hills Fm. Burnside River Fm.
Rocknest Fm.	Duhamel Fm.	absent?
Odjick Fm.	Hornby Channel Fm. Union Island group unconformity	Western River Fm.
absent	Wilson Island group	absent?

Early Proterozoic batholithic rocks

The Bear Province (Fraser, Hoffman, Irvine & Mursky 1972) is the batholithic belt of the Coronation geosyncline. It is divided into the Hepburn batholith to the east and the Great Bear batholith to the west (see figure 2). The mesozonal Hepburn batholith intrudes regionally metamorphosed supracrustal rocks of the Epworth basin. The epizonal Great Bear batholith intrudes mainly silicic volcanics, comagmatic with the batholith, and sediments derived from them. The batholithic rocks range in K–Ar age from 1725 to 1855 Ma, with the older ages occurring only in the Hepburn batholith. Archaean rocks are known only along the eastern margin of the Bear Province.

In the subsurface, plutonic rocks of the Great Bear batholith can be traced southward beneath Palaeozoic cover west of Great Slave Lake (see figure 1), but the western limits of the batholithic rocks are not known.

CORONATION GEOSYNCLINE

The type area of the Coronation geosyncline is at the latitude of the Epworth basin, within 200 km of Coronation Gulf (see figure 3). Here, the supracrustal rocks east of the Hepburn batholith are subdivided into three subparallel zones on the basis of contrasting styles of deformation (see figure 4) – a western zone of metamorphic tectonites, a central zone of imbricate thrust sheets, and an eastern autochthonous zone with anticlinal basement uplifts. The eastern limit of penetrative cleavage, separating the western and central zones, is termed the tectonite front. The eastern limit of major low-angle thrust faults, separating the central and eastern zones, is called the thrust front. The threefold structural subdivision is especially convenient because the most important sedimentary facies changes coincide with the tectonite and thrust fronts.

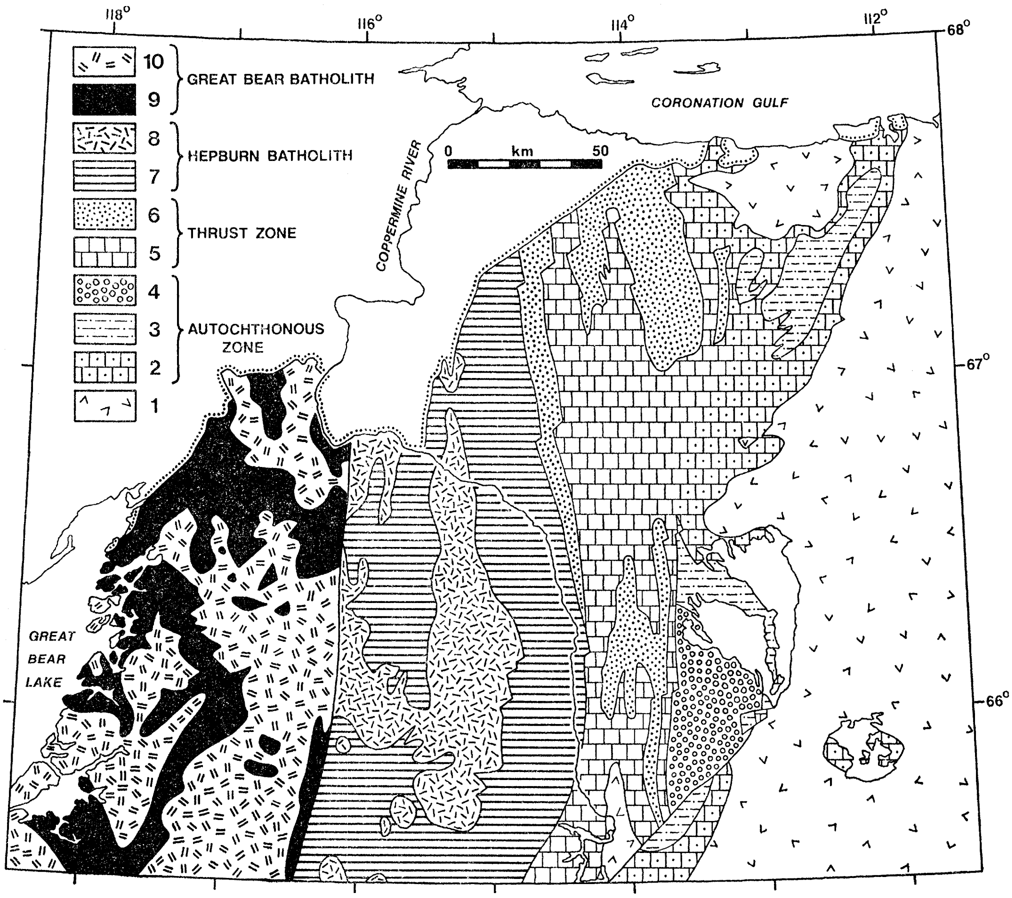

FIGURE 3. Geologic map of the Coronation geosyncline in the type area south of Coronation Gulf (modified from Fraser 1960).

Unit 1. Exhumed Archaean basement rocks.

Unit 2. Platform facies of the pe-quartzite, quartzite and dolomite phases of deposition in the autochthonous zone.

Unit 3. Platform facies of the pre-flysch and flysch phases of deposition in the autochthonous zone.

Unit 4. Calc-flysch, molasse and fanglomerate phases of deposition in the autochthonous zone.

Unit 5. Continental shelf facies of the pre-quartzite, quartzite and dolomite phases of deposition in the thrust zone.

Unit 6. Clastic wedge facies of the pre-flysch and flysch phases of deposition in the thrust zone.

Unit 7. Metamorphosed continental rise and clastic wedge facies in the tectonite zone.

Unit 8. Mesozonal granodiorite plutons of the Hepburn batholith.

Unit 9. Silicic welded tuff, trachybasalt and derived sedimentary rocks related to the Great Bear batholith.

Unit 10. Epizonal plutons of the Great Bear batholith.

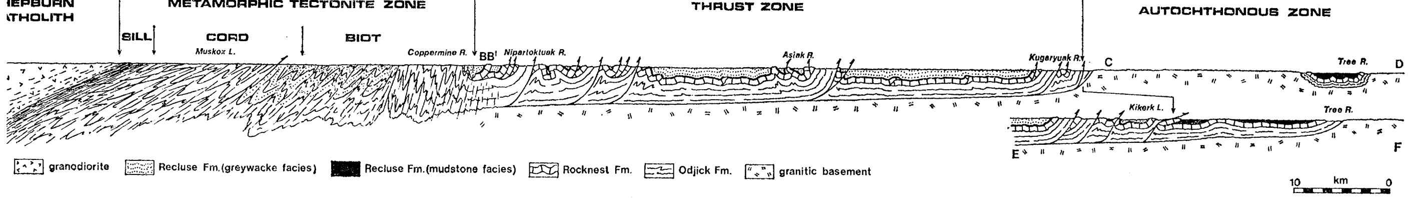

FIGURE 4. Representative structural cross-section of the north end of the Epworth basin showing the relative positions of the sillimanite (SILL), cordierite (CORD) and biotite (BIOT) isograds, tectonite front and the thrust front. Note that the Hepburn batholith overrides the metamorphosed supracrustal rocks at the western edge of the basin. The cross-section is drawn with no vertical exaggeration.

Depositional history

Eight phases of deposition are distinguished in the Epworth basin (see figure 5):
(8) fanglomerate phase (unnamed),
(7) molasse phase (Takiyuak Formation),
(6) calc-flysch phase (Cowles Lake Formation),

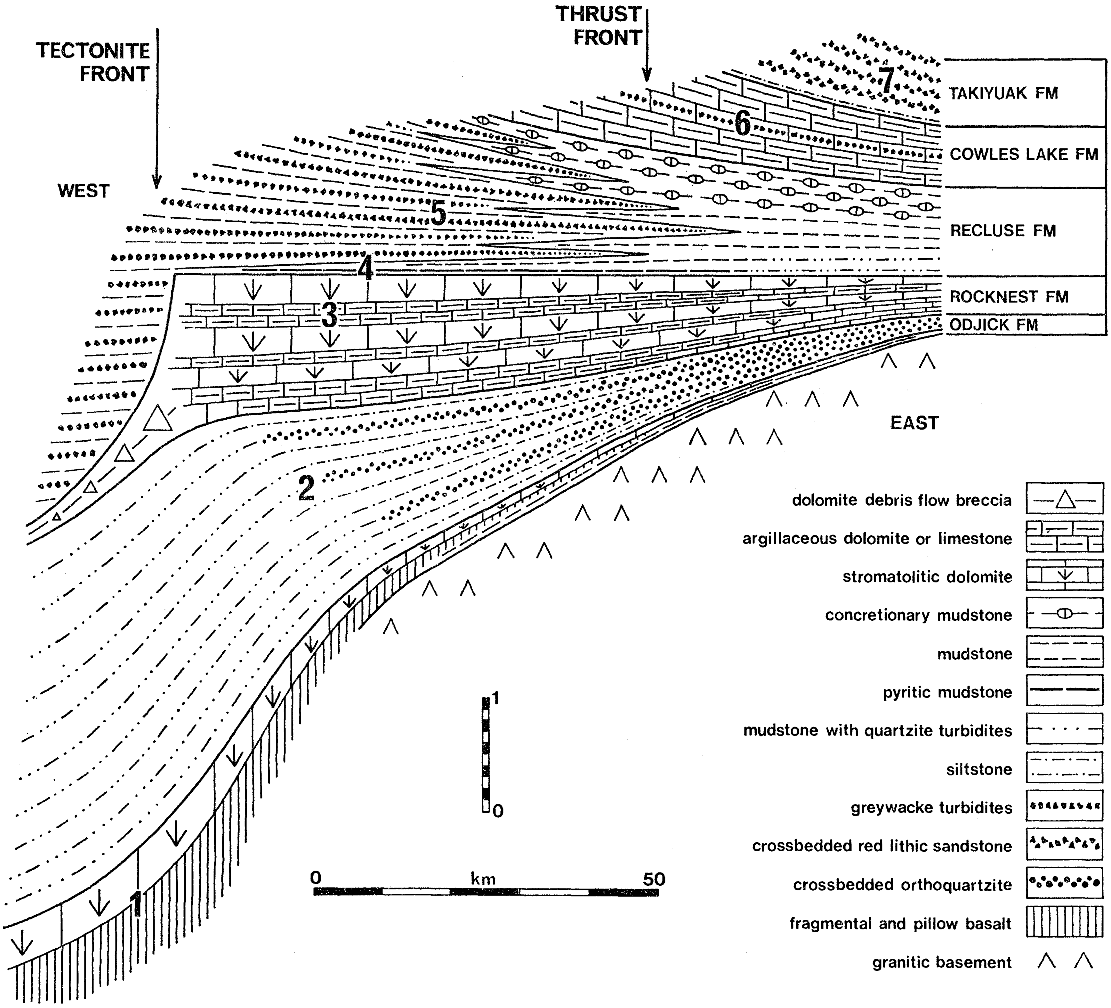

FIGURE 5. Stratigraphic cross-section of the Epworth basin drawn with no correction for tectonic shortening. The numerals indicate the phases of deposition described in the text.

(5) flysch phase (Recluse Formation),
(4) pre-flysch phase (basal Recluse Formation),
(3) dolomite phase (Rocknest Formation),
(2) quartzite phase (Odjick Formation),
(1) pre-quartzite phase (basal Odjick Formation).

The dolomite phase is by far the best exposed, being more resistant and less lichen-covered than the terrigenous rocks. The calc-flysch and younger phases are preserved only in the autochthonous zone at the south end of the basin.

Pre-quartzite phase

Although relatively thin east of the tectonite front, the pre-quartzite phase is significant in having the only volcanic rocks east of the Hepburn batholith.

In the autochthonous zone, the Archaean basement is overlain by dark grey mudstone, which thickens westward from 20 to 80 m, with graded siltstone laminations, dolomite nodules and a discontinuous veneer of arkose at the base. The mudstone is capped by stromatolitic dolomite, which thickens westward to 30 m at the thrust front, with beds containing basic volcanic ash.

In the thrust zone, the basal mudstone contains the principal surface of tectonic detachment. The overlying stromatolitic dolomite, with units of basic ash-fall tuff up to 5 m thick, occurs on the hanging-wall of several major thrust faults. Only at the south end of the basin is Archaean basement, there overlain by basic lava flows without intervening mudstone, exposed within the thrust zone.

Granitic basement has not been found in the tectonite zone. The oldest rocks there are hundreds of metres of fragmental and pillow basalt, overlain intergradationally by 250 m of dolomite containing cherty stromatolites and coarse quartz sand.

Quartzite phase

Sharply overlying the stromatolitic dolomite of the pre-quartzite phase is a westerly thickening wedge of compositionally mature terrigenous sediments derived from the platform to the east.

In the autochthonous zone, the quartzite phase consists mainly of varicoloured crossbedded orthoquartzite (see figure 6, plate 9), as little as 140 m thick, with interbeds of dark grey ferruginous mudstone, quartz pebblestone and sandy dolomite. The ratio of mud to sand increases westward, as does the total thickness – from 750 m in the easternmost thrust sheet to perhaps as much as 3000 m at the tectonite front. The alluvial and coastal marine sediments of the east pass westward into submarine fan deposits in the tectonite zone, where there are great thicknesses of laminated silty mudstone with thin quartzite turbidites, channels filled by pebbly quartzite and sedimentary slump folds and breccias.

Dolomite phase

The westward deepening quartzite provided the foundation for a westward-facing shallow-water dolomite shelf, 500 m thick in the autochthonous zone and up to 1200 m in the thrust zone.

The shelf sequence is made up of stacked shoaling-upward cycles (see figure 7, plate 10), each 2 to 20 m thick, in which laminated argillaceous dolomite, invariably with a sharp base, grades upward into oolitic (see figure 6) or stromatolitic dolomite with chert nodules. The cycles result from progradation of carbonate tidal flats across a shallow shelf lagoon and many can be traced for tens of kilometres across the basin. The ratio of argillaceous to cherty dolomite oscillates upward through the sequence but is greatest at the base and least at the top.

Approaching the tectonite front from the east, the cyclic pattern becomes less evident and the sequence contains mound-shaped masses of stromatolites, tens of metres in diameter, instead of persistent beds. The stromatolite mound facies is typical of the outer margin of the shelf. West of the tectonite front, off the edge of the shelf, the sequence thins dramatically to less than

110 m of mudstone with beds of dolomite breccia, containing blocks up to 50 m in length, deposited by submarine debris flows.

Pre-flysch phase

The dolomite shelf is sharply overlain by a thin succession of black laminated pyritic mudstone, only 80 m thick at the tectonite front. The mudstone thickens eastward and, in the autochthonous zone, contains thin quartzite turbidites derived from the east. The black mudstone accumulated during foundering of the dolomite shelf into deep water before the flysch phase of deposition.

Flysch phase

Over the pre-flysch was deposited a wedge of coarse greywacke turbidites (see figure 8, plate 10) that thickens westward to more than 1400 m above the dolomite shelf-edge at the tectonite front. The greywacke occurs in composite graded beds, commonly as much as 3 m thick and persistent laterally for at least 2 km. Sole markings are poorly developed and ripple-drift laminations occur only at the tops of the graded flow units. The greywacke is derived from the batholithic belt to the west and contains unsorted angular clasts, up to 5 mm in diameter, mostly of plutonic and metamorphosed sedimentary rock types (see figure 6).

The greywacke turbidite tongues pinch out near the thrust front and, in the autochthonous zone, the flysch phase is represented by about 800 m of dark green and grey mudstone with thin dolomite and granular hematite ironstone beds near the base, and abundant calcareous concretions near the top (see figure 8).

Calc-flysch phase

Gradationally overlying the concretionary mudstone of the flysch phase in the autochthonous zone are 500 m of grey laminated argillaceous limestone with widely spaced thin greywacke turbidite beds. The limestone laminations, 1 to 3 mm thick, typically pinch and swell, and are separated by equally thin partings of dark grey fissile mudstone. Beds of mudstone crowded with flakey limestone clasts are common but current-produced sedimentary structures such as ripple marks are absent except in the greywacke turbidites, which are derived from the west. Beds of brecciated red argillaceous limestone with stromatolites and mudcracks occur at the top of the succession.

Molasse phase

Compositionally immature terrigenous sediments of the molasse are derived, as is the greywacke below, from the batholithic belt to the west but were deposited in mainly alluvial, rather than submarine environments. At the base are 125 m of red mudstone with thin beds of ripple-laminated and mudcracked buff siltstone. These rocks are overlain by a minimum of 500 m of red mudstone with fining-upward units of friable red lithic sandstone many metres thick. The sandstone, commonly with sets of tabular crossbeds up to 2 m thick, is typically laminated, well-sorted, calcite-cemented and contains abundant angular to subrounded clasts of terrigenous and carbonate sedimentary rocks, intermediate to silicic volcanic rocks, and subsidiary plutonic and metamorphic rocks.

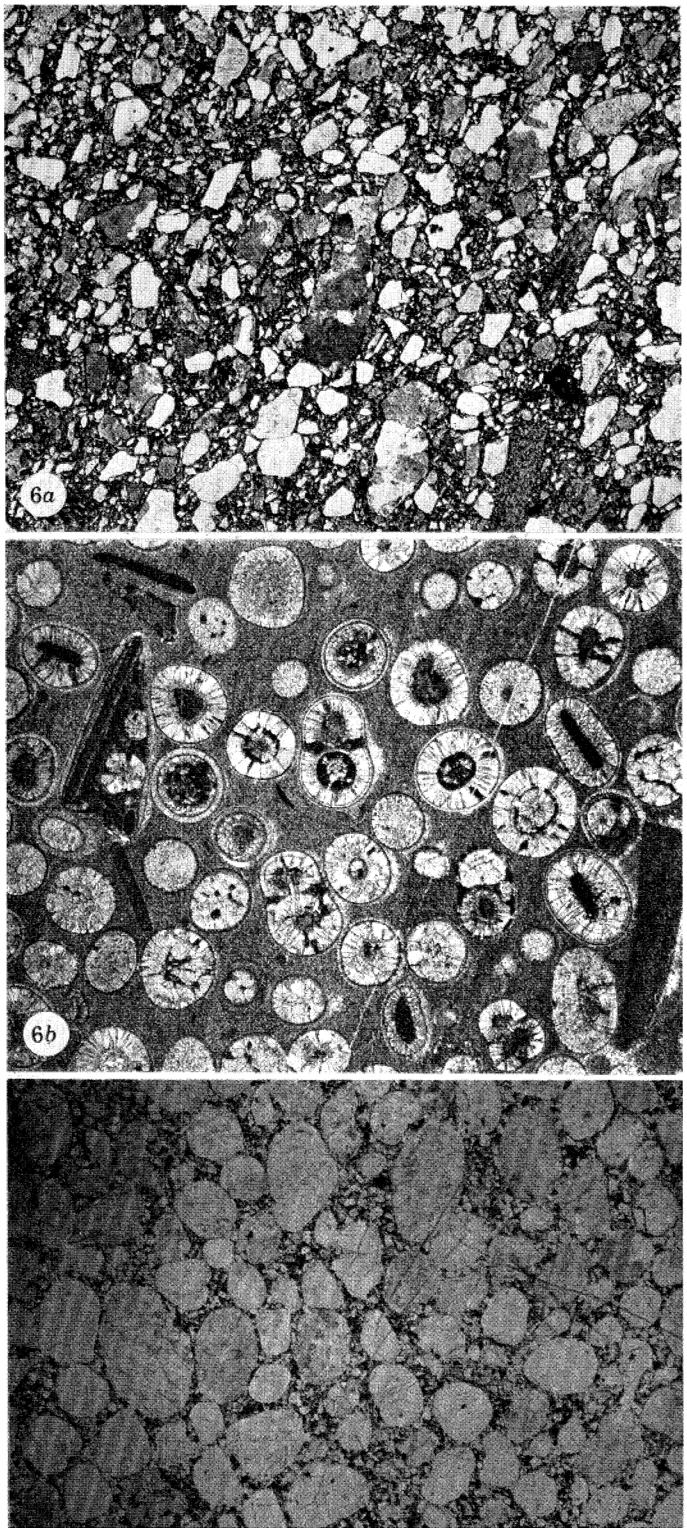

FIGURE 6. (*a*) Microphotograph of coarse greywacke from the Recluse Formation composed of unsorted angular fragments of quartz, plutonic and metamorphic rocks fragments set in a dark fine-grained matrix. (*b*) Dolomite from the Rocknest Formation composed of ooids, compound ooids and intraclasts set in a mainly fine-grained dolomite matrix. The high degree of textural preservation is typical of the formation despite complete dolomitization. (*c*) Orthoquartzite from the Odjick Formation composed of rounded coarse sand set in a matrix of coarse silt. The dark areas are void spaces.

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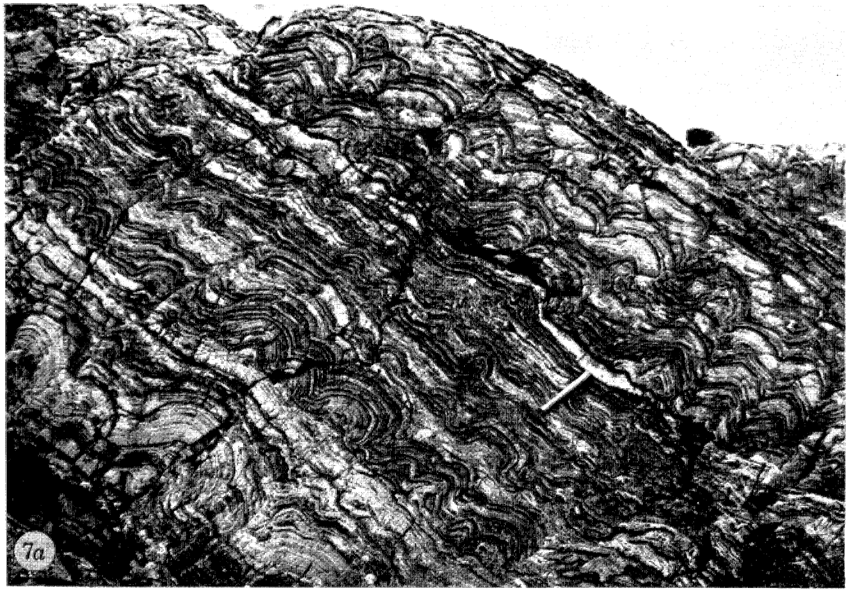


FIGURE 7. (*a*) Laterally linked domal stromatolites in cherty dolomite in the upper part of a shoaling-upward cycle in the Rocknest Formation, east of Eokuk Lake. Shaly dolomite of the lower part of the cycle can be seen in the lower left corner of the photograph. A geological pick is in the right-central part of the photograph for scale. (*b*) Cyclic alternation of dark shaly dolomite and light cherty stromatolitic dolomite in the Rocknest Formation north of Kikerk Lake. The photograph includes about 50 m of section in the foreground.

FIGURE 8. (*a*) Dark green mudstone with calcareous concretions in the Recluse Formation east of the feather edges of the greywacke turbidite tongues, Kikerk Lake. The scale is divided into 3 cm intervals (tenths of feet). (*b*) Inclined air photograph of greywacke turbidites in the Recluse Formation, near the east side of the Coppermine River at latitude 66° 30′. About 1000 m of steeply dipping beds are shown in the farground of the photograph, taken from an elevation of 600 m above the ground. Note the great lateral continuity of the beds, most of which are composed of several amalgamated turbidites.

Fanglomerate phase

This phase is represented only by a few erratic blocks, up to 3 m in diameter, of red and buff, calcite-cemented conglomerate containing cobbles of molasse sandstone, older sedimentary rocks and granitic gneiss, probably from the Archaean basement.

Synthesis of the depositional history

East of the tectonite front, the pre-quartzite, quartzite and dolomite phases constitute a westward-facing shelf sequence constructed at the margin of a continental platform of Archaean rocks. The outer edge of the shelf becomes more sharply delineated upward in the sequence, concomitant with decreased influx of terrigenous sediment from the platform to the east.

West of the tectonite front, pillow basalt capped by shallow-water dolomite was deposited over basement of unknown character during the pre-quartzite phase. The basaltic volcanism may have accompanied rifting that created the platform margin but I can provide no direct evidence of contemporaneous faulting to support the hypothesis of initial rifting. Subsequently, subsidence was sufficient to accommodate the thick sequence of relatively deep-water shelf-derived sediments that accumulated during the quartzite phase. The off-shelf basin became starved of terrigenous sediment during construction of the dolomite shelf. These allochthonous rocks may originally have been parts of a continental rise west of the shelf.

The pivotal event in the history of the supracrustal belt of the geosyncline was the foundering of the shelf. As the shelf subsided into deeper water, it was mantled by the starved black mudstone of the pre-flysch phase. Meanwhile, greywacke turbidites of the flysch phase were filling the submarine depression west of the drowned shelf-edge and eventually spilled eastward over the pre-flysch. The turbidite tongues pinch out near the thrust front, probably on a sea bottom shoaling onto the platform. The succession shoals upward as well as eastward, with flysch succeeded by argillaceous limestone of the calc-flysh phase and ultimately by non-marine red beds of the molasse phase. The terrigenous sediments of the flysh, calc-flysch and molasse phases constitute a clastic wedge derived from the batholithic belt to the west, in contrast to the platform-derived quartzite of the underlying shelf and rise.

From fragmentary remains, it is possible to say only that the fanglomerate phase was deposited unconformably over the molasse and older rocks, and that, judging from the cobbles of Archaean granite, deposition may have accompanied the basement uplifts peculiar to the autochthonous zone. But there are limits to how far even I am willing to push the interpretation of a few erratics.

Metamorphism and structural zonation

Tectonite zone

There is a progressive increase in the grade of regional metamorphism from the tectonite front westward to the Hepburn batholith (see figure 4). Weakly recrystallized phyllite is converted, as much as 13 km from the edge of the batholith, to fine-grained schist with macroscopic aggregates of biotite. Cordierite, in the form of highly poikilitic ovoid porphyroblasts, makes its appearance about 2 km to the west. Coarse-grained augen paragneiss with rusty bands containing sillimanite, in both fibrous and coarsely crystalline habits, occurs within 1 to 5 km of the batholith. Small quantities of garnet and andalusite occur in some of the cordierite schists and paragneisses, but staurolite and kyanite are absent. The mineral assemblages are typical of low-pressure metamorphic facies series (Miyashiro 1961). The growth of cordierite and andalusite porphyroblasts was contemporaneous with the development of cleavage.

Folds west of the tectonite front approach similar geometry and are transected by penetrative cleavage fans. At the tectonite front, the folds are upright, the cleavage near-vertical, and the fold hinges are north-trending and generally near-horizontal. To the west, the folds become progressively more overturned, with west-dipping cleavage fans. Strain indicators, mainly stretched chlorite blebs in the basalt flows of the pre-quartzite phase, are elongate in the cleavage planes perpendicular to the fold hinges. Post-metamorphic kink folding of the cleavage is ubiquitous. In the paragneiss, foliation dips gently to the west beneath the concordant margin of the Hepburn batholith.

At least one major west-dipping thrust fault, which places a great thickness of quartzite-phase schist over flysch-phase meta-greywacke, occurs in the central part of the zone. The cordierite isograd passes obliquely across the trace of the thrust fault without dislocation. Thrusting predated the metamorphism and development of cleavage.

Also predating metamorphism are gabbro sills, up to 200 m thick, that intrude the flysch and older rocks, particularly near the tectonite front. These are unrelated to the plethora of post-tectonic dolerite dykes of middle Proterozoic age.

Thrust zone

East of the tectonite front, cleavage is developed only locally and metamorphism is limited to the contacts of the post-tectonic dolerite dykes.

Folds are predominantly flexural, with broad gently folded synclinoria, up to 22 km across, in which the flysch is exposed, separated by more tightly compressed anticlinoria in which rocks of the quartzite and dolomite phases are exposed (see figure 4). Box-shaped folds with sharp hinges are common and, although most are upright, their east limbs are generally steeper or overturned. In the cores of such folds, tectonic mega-breccia is typical of the dolomite phase and cleavage is developed in mudstone of the quartzite phase.

Thrust faults, most of which dip 40 to 50°, come to the surface on the west limbs of the anticlinoria. Most of the major thrusts have the stromatolitic dolomite of the pre-quartzite phase on the hanging-wall, suggesting that the thrusts are generated from a detachment surface in the basal mudstone overlying the Archaean basement. Many of the thrusts place pre-quartzite over flysch and carry well over 1000 m of section. Subsidiary thrusts are detached from a specific argillaceous dolomite unit two-thirds of the way from the top of the dolomite phase. Mapping of the thrust traces reveals that most are concordant with rocks on the hanging-wall but regionally discordant with those of the foot-wall. In one of the easternmost anticlinoria, a major thrust that dips the typical 40° on the west limb is folded to the point of being slightly overturned where it reappears on the east limb. Along the strike of this anticlinorium, the stratigraphic separation between the hanging-wall and foot-wall diminishes and ultimately becomes a fully recumbent but coherent fold. Small-scale thrust wedges that repeat 10 to 100 m of section are very common in the dolomite phase, particularly in the west half of the thrust zone.

Basement uplifts

The autochthonous supracrustal rocks east of the thrust front are nearly flat-lying except around the margins of anticlinal basement uplifts (see figure 4). The best known uplift, at the northeast corner of the Epworth basin (see figure 3), is 55 km in diameter and, on its steep southeast flank, has a minimum structural relief of 2000 m over a distance of 3 km. Only 13 km to the southeast is the erosional edge of a second uplift, oversteepened to the northwest. Between the opposing uplifts is the box-shaped northeast-trending Tree River synclinorium.

On the oversteepened flanks of the uplifts, the basement-cover surface is recumbently folded. The folds, about 800 m in wavelength and 200 m in amplitude, have arched anticlines and tight synclines. The overturned limbs dip as little as 20°, the basement-cover surface having been rotated through 160° (see figure 9, plate 11). In the tight synclines, the basal mudstone of the pre-quartzite phase is cleaved and strain indicators, flattened in the plane of cleavage, are elongate perpendicular to the fold hinges.

The most remarkable aspect of the uplifts is that the basement-cover contact is an intact erosional unconformity (see figure 9). There is little evidence of detachment or slip on the unconformity and the contact is non-depositional only where cut by transcurrent faults that post-date the uplifts. A distinctive argillaceous dolomite bed less than 1 m thick occurs at the base of the pre-quartzite phase and can be traced for tens of kilometres around the uplifts.

The apparent ease with which the basement granites and gneisses were deformed along with their thin unmetamorphosed cover requires explanation. The basement rocks have no macroscopic fabric related to the uplifts but they have been broken into discrete variously oriented blocks, many metres in diameter. At the unconformity, crevasses between such blocks have, in places, been tectonically infilled with the overlying mudstone (see figure 9). Internal deformation of the basement may have been accomplished mainly by movement between but not within the individual blocks, giving the basement rocks the rigidity of a sack of potatoes.

In the supracrustal rocks around the oversteepened margins of the uplifts are cascading folds, about 1200 m in wavelength, with alternating vertical and horizontal limbs, and *en échelon* hinges (see figure 7). The folds, generally northeast- to east-trending, intersect and warp the north-trending folds and thrust sheets at the eastern margin of the thrust zone. This suggests that the thrusting predated at least the final stages of emplacement of the uplifts. However, the thrust front is regionally refracted by the uplifts (see figure 3), from which it is concluded that the initial stages of uplifting occurred before or during thrusting.

Transcurrent faults

Northeast-trending faults with dextral near-horizontal displacement of up to 5 km occur throughout the supracrustal belt, particularly in the thrust zone. They transect all other structures but pre-date the middle Proterozoic dolerite dykes. Complimentary sinistral northwest-trending faults are much less common and occur mainly in the tectonite zone.

Synthesis of the structural zonation

Deformation in the tectonite and thrust zones involves compression and transport of supracrustal rocks toward the platform. This may be viewed as part of a continuum that began with eastward transport of sediment during the flysch phase of deposition. Both sedimentation and tectonism may be genetically related to the rise and eastward spreading of the Hepburn batholith over the paragneisses at the western margin of the Epworth basin (see figure 4).

In the tectonite zone, flysch sedimentation preceded thrusting, which in turn preceded penetrative deformation and metamorphism. Over-extending Walther's law, we may speculate that flysch sedimentation, thrusting and penetrative deformation were occurring simultaneously in adjacent zones, zones that migrated eastward with time, producing the observed temporal sequence of events, until frozen with the tectonite and thrust fronts 30 km and 70 to 85 km respectively from the present erosional edge of the batholith. The absolute amount of eastward tectonic transport has not been estimated but shortening of 33 % in the thrust zone would place the palinspastic outer edge of the dolomite shelf as far west as the edge of the batholith.

Compression of the supracrustal rocks in the thrust zone occurs above a basal detachment surface. Nowhere is there evidence of basement involvement in the thrusting (see figure 4). The situation could hardly be more different east of the thrust front, where asymmetric anticlinal basement uplifts are responsible for the deformation of autochthonous supracrustal rocks. The basement uplifts are believed to be contemporaneous with thrusting because the thrust front is both refracted by the uplifts and folded by them.

The age of molasse deposition relative to thrusting is unknown but the molasse could not have been derived from either the leading edges of the thrust sheets or from the basement uplifts – the high proportion of silicic volcanic rock fragments is incompatible with the source rocks involved.

Summary of the Coronation geosyncline

The geosyncline developed initially as adjacent shelf and rise sequences along the western margin of a continental platform of Archaean rocks. Foundering of the shelf was followed by the deposition of a clastic wedge of immature terrigenous sediments shed eastward toward the platform from the batholithic belt to the west.

Compression and eastward transport of the supracrustal rocks toward the platform, probably the result of eastward spreading of the Hepburn batholith, produced a threefold structural zonation in the Epworth basin. In the western zone, thrust faulting was followed by low-pressure metamorphism contemporaneous with the development of penetrative cleavage and similar folds. In the central zone, the supracrustal rocks are unmetamorphosed but were folded flexurally and overthrust eastward above a basal detachment surface. In the eastern zone, the supracrustal rocks are autochthonous and deformed mainly around asymmetric anticlinal basement uplifts.

Athapuscow aulacogen

The Goulburn and Great Slave basins contain supracrustal rocks correlative and doubtless once continuous with the Epworth basin, but their tectonic setting is different. Instead of being part of the Coronation geosyncline marginal to the continental platform, they were deeply subsiding aulacogens extending far into the interior of the platform (see figure 2). The Great Slave basin is described here in detail and the name 'Athapuscow aulacogen' is introduced, using the aboriginal name for Great Slave Lake.

Supracrustal rocks deposited in the Athapuscow aulacogen occur along the southeast side of the Great Slave basin (see figure 10). Correlative rocks on the northwest side of the basin were deposited on the platform adjacent to the aulacogen. The long history of subsidence in the aulacogen is manifested by abrupt changes in sedimentary facies and thickness (see figure 11) from the platform into the aulacogen. In addition, the aulacogen is marked by volcanic rocks at several stratigraphic levels lacking volcanics elsewhere. The supracrustal rocks in the aulacogen thicken toward the southwest and the aulacogen presumably opens into the southern extension of the Coronation geosyncline beneath the Palaeozoic cover west of Great Slave Lake.

The rocks in the aulacogen are broadly folded about northeast-trending axes and cut by a complex system of high-angle faults that parallel the folds (see figure 12). On the platform

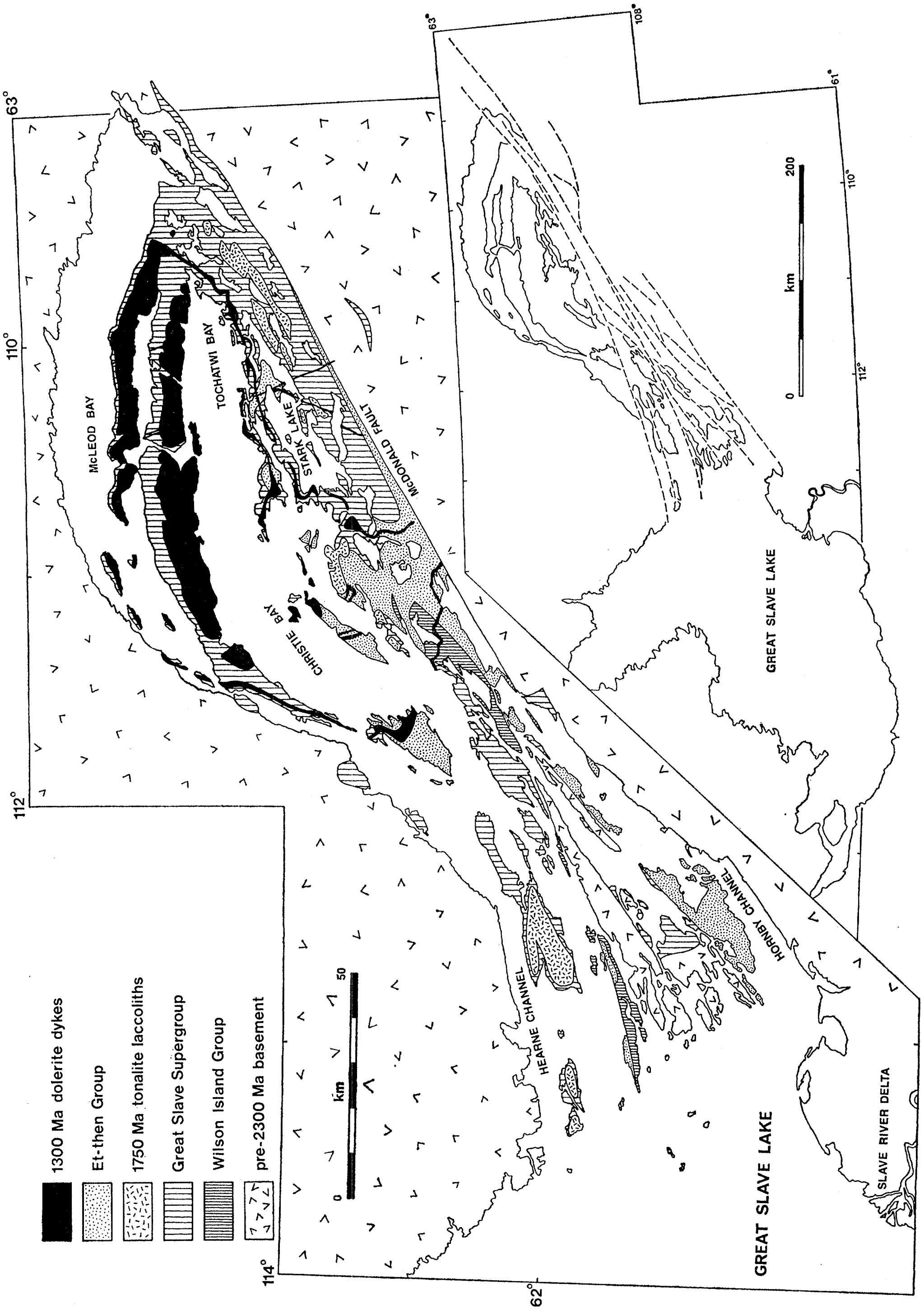


FIGURE 10. Geologic map of the Great Slave basin, showing some of the major faults in the inset map in the lower right (modified from Stockwell 1936).

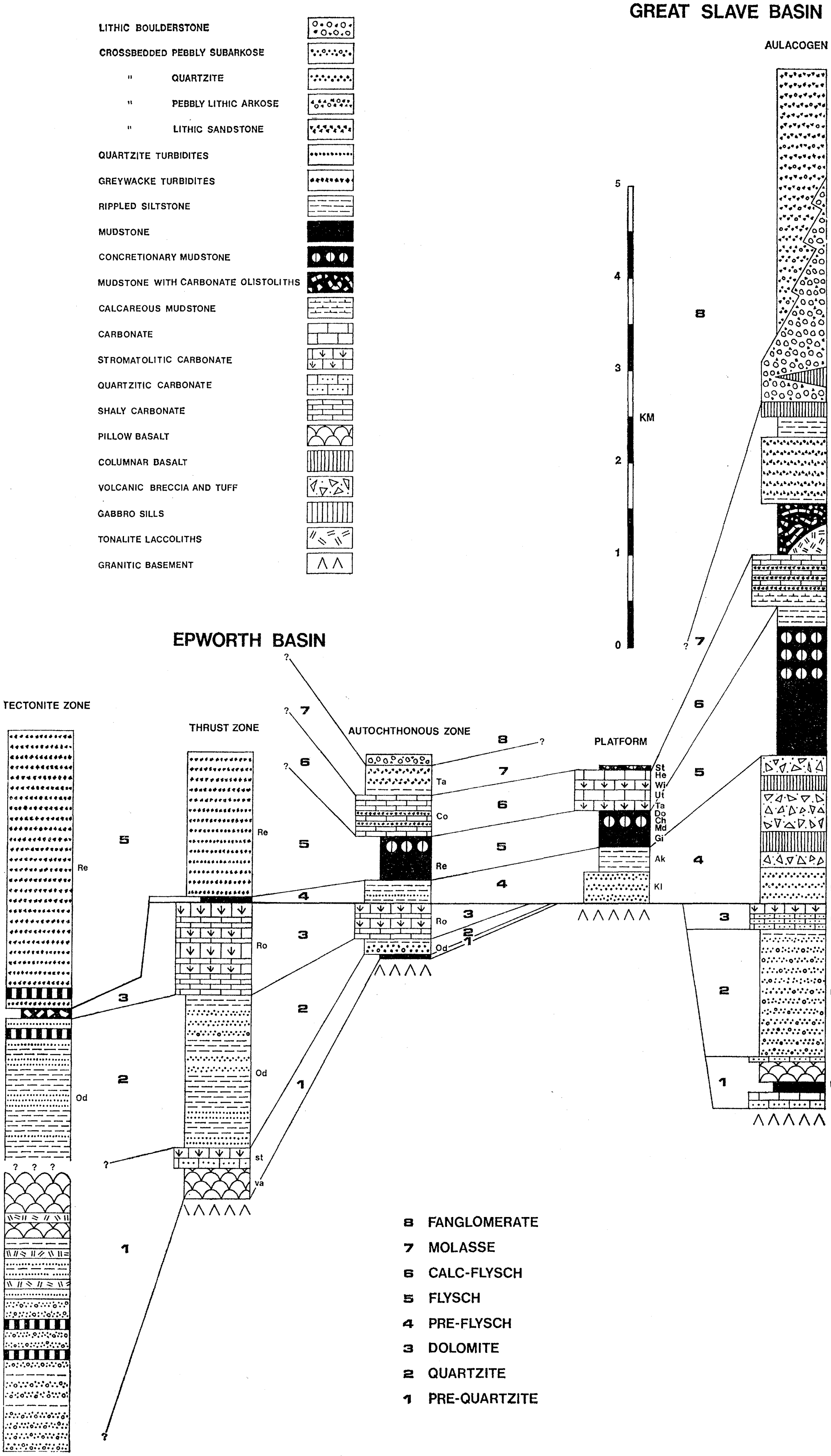


FIGURE 11. Representative columnar stratigraphic sections of the Epworth and Great Slave basins. Note the great thickening from the north side of the Great Slave basin to the south side. The lower part of the section in the tectonite zone of the Epworth basin is from a metamorphic belt within the Hepburn batholith, and both its internal stratigraphy and correlation with the pre-quartzite phase in the thrust zone are highly speculative. The abreviated formation names, on the right sides of the columns, are keyed to figures 13, 15, 16, 20 and 22.

the north, the supracrustal rocks dip gently toward the aulacogen. The Archaean basement south of the aulacogen was stripped of its cover during the fanglomerate phase and the depositional history of this area must be interpreted from the boulders shed northward into the aulacogen at that time. Archaean granitic rocks and isoclinally folded supracrustal rocks of the Wilson Island group are exposed on horsts within the aulacogen. Unlike the geosyncline adjacent to the plutonic belt, the aulacogen lacks regional metamorphism and there is only local development of cleavage and low-angle thrust faults.

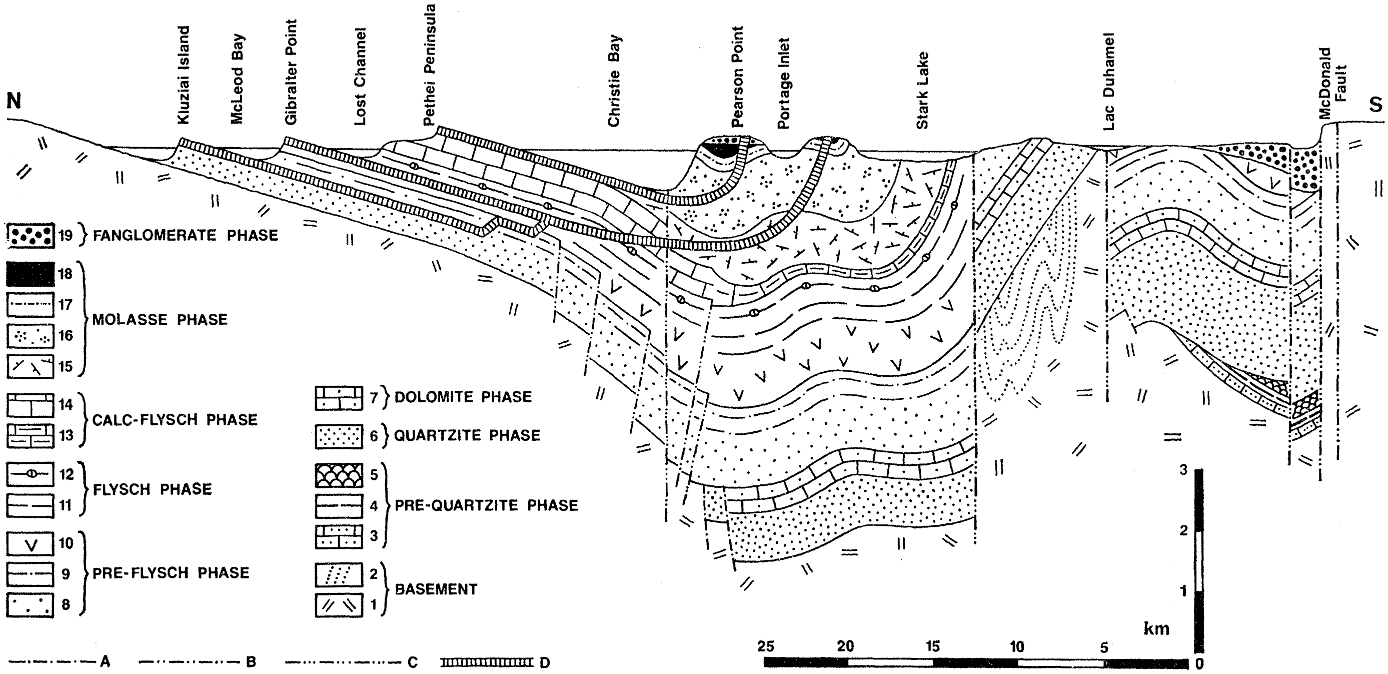


FIGURE 12. Representative structural cross-section of the Great Slave basin showing the postulated relations between faults formed during the rifting stage (A), sagging stage (B) and transcurrent stage (C) in the evolution of the aulacogen, as explained in the text, and the post-tectonic dolerite dykes and sills (C).

Unit 1. Archaean granitic and metamorphic rocks.
Unit 2. Wilson Island group.
Unit 3. Quartzitic dolomite (Union Island group).
Unit 4. Black mudstone (Union Island group).
Unit 5. Pillow basalt (Union Island group).
Unit 6. Coarse subarkose (Hornby Channel Formation).
Unit 7. Quartzitic and stromatolitic dolomite (Duhamel Formation).
Unit 8. Fine quartzite (Kluziai Formation).
Unit 9. Red siltstone (Akaitcho River Formation).
Unit 10. Volcanic rocks (Seton Formation).
Unit 11. Red mudstone (Gibralter Formation).
Unit 12. Red concretionary mudstone (McLeod Bay Formation).
Unit 13. Shaly limestone (trough facies of Pethei group).
Unit 14. Stromatolitic limestone and dolomite (platform facies of Pethei group).
Unit 15. Red mudstone with carbonate olistostromes (Stark Formation).
Unit 16. Red lithic sandstone (Tochatwi Formation).
Unit 17. Red siltstone (Portage Inlet Formation).
Unit 18. Columnar basalt (Pearson Formation).
Unit 19. Fanglomerate (Et-then group).

Depositional history

Pre-quartzite phase

Rocks of the quartzite phase lie directly on basement in all but the southwestern part of the aulacogen. There, they paraconformably overlie a pre-quartzite succession that begins with several hundred metres of non-stromatolitic dolomite, with a discontinuous veneer of arkose and granite rubble at the base (see figure 13). The dolomite is overlain by black mudstone with thin beds of dolomite at the base and graded siltstone at the top. Above the mudstone is a complex of pillow basalt, pillow breccia and gabbro sills, capped by cherty dolomite with discontinuous beds of quartz pebblestone.

Quartzite phase

The quartzite phase is coarser grained and more feldspathic in the aulacogen than on the shelf of the geosyncline. Pebbly trough-cross-bedded subarkose (see figure 14, plate 11) thickens from less than 200 m at the northeast end of the aulacogen to more than 1600 m at the southwest end, but is absent on the platform (see figure 13). Granite pebblestone beds occur near the base and glauconitic quartzite, with ripple-laminated silty mudstone and basic ash-fall tuff with accretionary lapilli, occurs at the top. A single bed of stromatolitic dolomite drapes the basement granite at the southwest end of the aulacogen. Sediment transport, mainly by braided rivers, was to the southwest, down the axis of the aulacogen.

Dolomite phase

The dolomite phase is absent at the northeast end of the aulacogen and on the platform. It occurs elsewhere in the aulacogen but is well exposed only in one area midway along its length. There, it is 295 m thick and consists of the same cyclic alternation of laminated argillaceous dolomite and cherty oolitic and stromatolitic dolomite as in the geosyncline, but with the addition of white cross-bedded orthoquartzite, probably derived from the uplifted margins of the aulacogen, capping many of the cycles.

Pre-flysch phase

Deposits of the pre-flysch phase are markedly different in the aulacogen than the starved black mudstone of the geosyncline. Paraconformably overlying the dolomite phase in the aulacogen are 470 m of pink and grey cross-bedded fine-grained even textured subarkose with abundant heavy-mineral laminations and quartz granules. Similar, but thinner, rocks overlie the Archaean basement on the platform (see figure 13). The subarkose was shed from the platform into the aulacogen and southwestward along its axis.

Overlying the subarkose on the platform are about 200 m of red mudstone with thin beds of ripple-laminated and mudcracked siltstone, white glauconitic subarkose with opposed cross-bedding, and granular hematitic ironstone, partly replaced by spherulitic carbonate, with calcite-filled gypsum casts.

In the southwestern part of the aulacogen, the red mudstone and, locally, the underlying subarkose, are complexly intercalated with volcanic rocks as much as 1400 m thick (see figure 15). The volcanics consist mainly of coalesced basic tuff-rings developed around breccia-filled pipes up to 1000 m in diameter. The tuff-rings are built up of vitric and lithic ash-fall tuff, base-surge deposits, laharic breccias and columnar basalt flows. Silicic flows and tuffs occur locally. Primary dips up to 25° occur close to the pipes and on the distal flanks of the tuff-rings are subaqueously reworked volcanic-pebble conglomerate and tuffaceous mudstone with beds of granular hematitic ironstone.

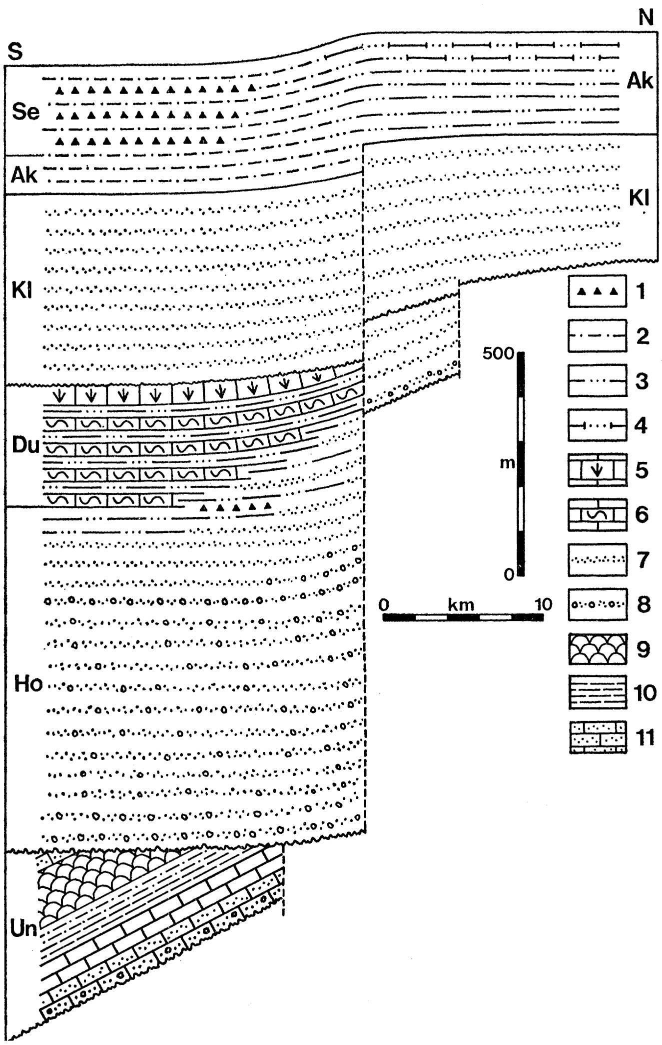


FIGURE 13. Diagrammatic stratigraphic cross-section of the Union Island and Sosan groups. The pre-quartzite phase is the Union Island group (Un), the quartzite phase is the Hornby Channel Formation (Ho), the dolomite phase is the Duhamel Formation (Du), and the pre-flysch phase includes the Kluziai Formation (Kl), the Akaitcho River Formation (Ak) and the Seton Formation (Se).

Unit 1. Fragmental basalt.

Unit 2. Red graded siltstone.

Unit 3. Red rippled and mudcracked siltstone.

Unit 4. Like Unit 3 but with thin beds containing granular hematitic ironstone, spherulitic limestone and gypsum casts.

Unit 5. Dolomite with large columnar stromatolites.

Unit 6. Dolomite with stromatolites and beds of crossbedded orthoquartzite.

Unit 7. Pink crossbedded fine quartzite with heavy-mineral laminations.

Unit 8. Crossbedded pebbly coarse subarkose (see figure 14).

Unit 9. Pillow basalt and pillow breccia.

Unit 10. Black mudstone with graded siltstone and thin dolomite beds.

Unit 11. Quartzitic dolomite.

Flysch phase

Whether flysch sedimentation in the geosyncline began before the end of volcanism in the aulacogen is not known. At the southwest end of the aulacogen, the volcanic rocks are intergradationally overlain by perhaps as much as 1450 m of red and green fissile mudstone (see figure 15). The mudstone thins to 350 m at the northeast end of the aulacogen and a similar thickness on the platform, where it is entirely red. The lower part of the mudstone succession contains thin beds of granular hematitic ironstone, intimately associated with spherulitic limestone, flat-pebble intraformational conglomerate and calcite-filled gypsum casts. In the upper part of the mudstone succession are abundant calcareous concretions with lags of concretion-pebble conglomerate.

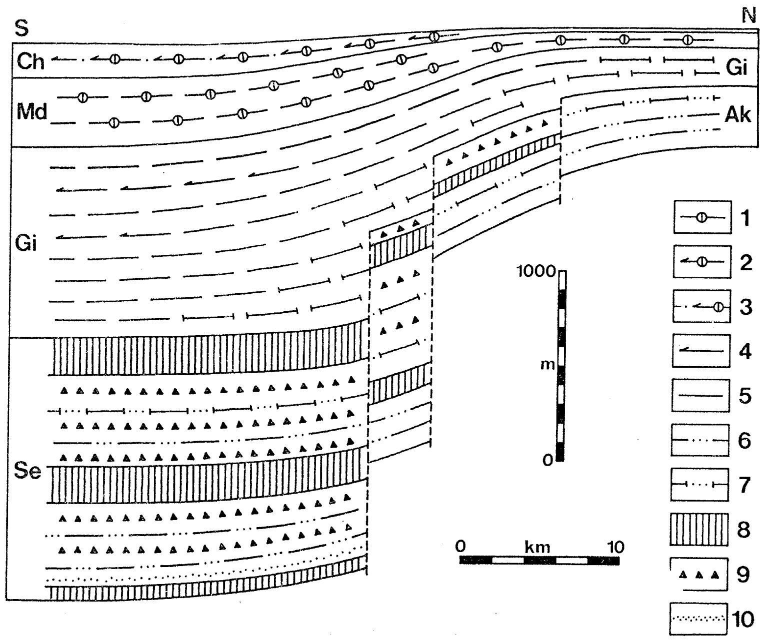
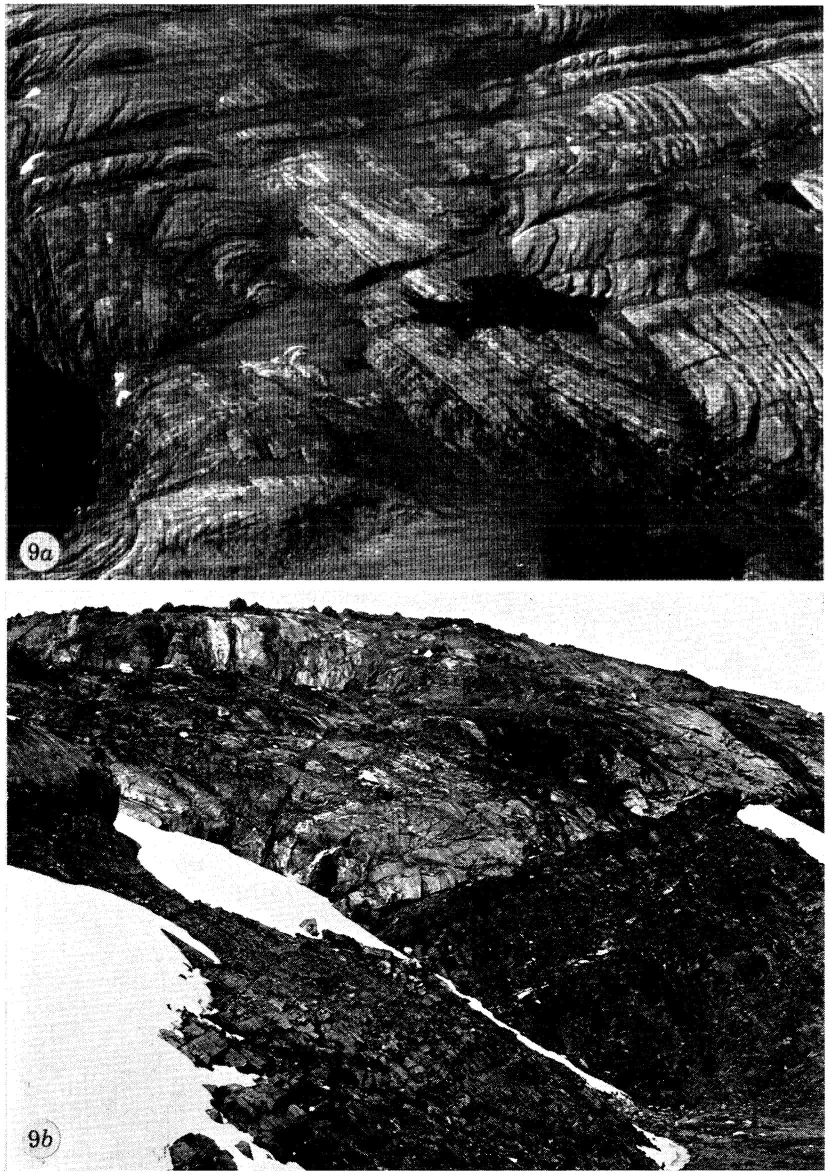


FIGURE 15. Diagrammatic stratigraphic cross-section of the Kahochella Group. The pre-flysch phase includes the Seton Formation (Se) and the Akaitcho River Formation (Ak); the flysch phase includes the Gibralter Formation (Gi), the McLeod Bay Formation (Md) and the Charlton Bay formation (Ch).

Unit 1. Red concretionary mudstone.
Unit 2. Dark green concretionary mudstone with bentonite beds.
Unit 3. Dark concretionary mudstone with graded siltstone beds.
Unit 4. Dark green mudstone.
Unit 5. Red mudstone.
Unit 6. Red rippled and mudcracked siltstone.
Unit 7. Like Unit 6 but with thin beds containing granular hematitic ironstone, spherulitic limestone and gypsum casts.
Unit 8. Columnar basalt flows.
Unit 9. Basaltic ash-fall tuff, base surge tuff, laharic breccia.
Unit 10. Crossbedded fine quartzite.

The red mudstone is capped, on the platform, by 10 m of dark green pyritic mudstone with thin bentonite beds and laterally coalesced concretions, individually up to 50 cm in diameter. The green mudstone thickens to 35 m at the northeast end of the aulacogen and more than 150 m at the southwest end, where it has thin graded siltstone beds.



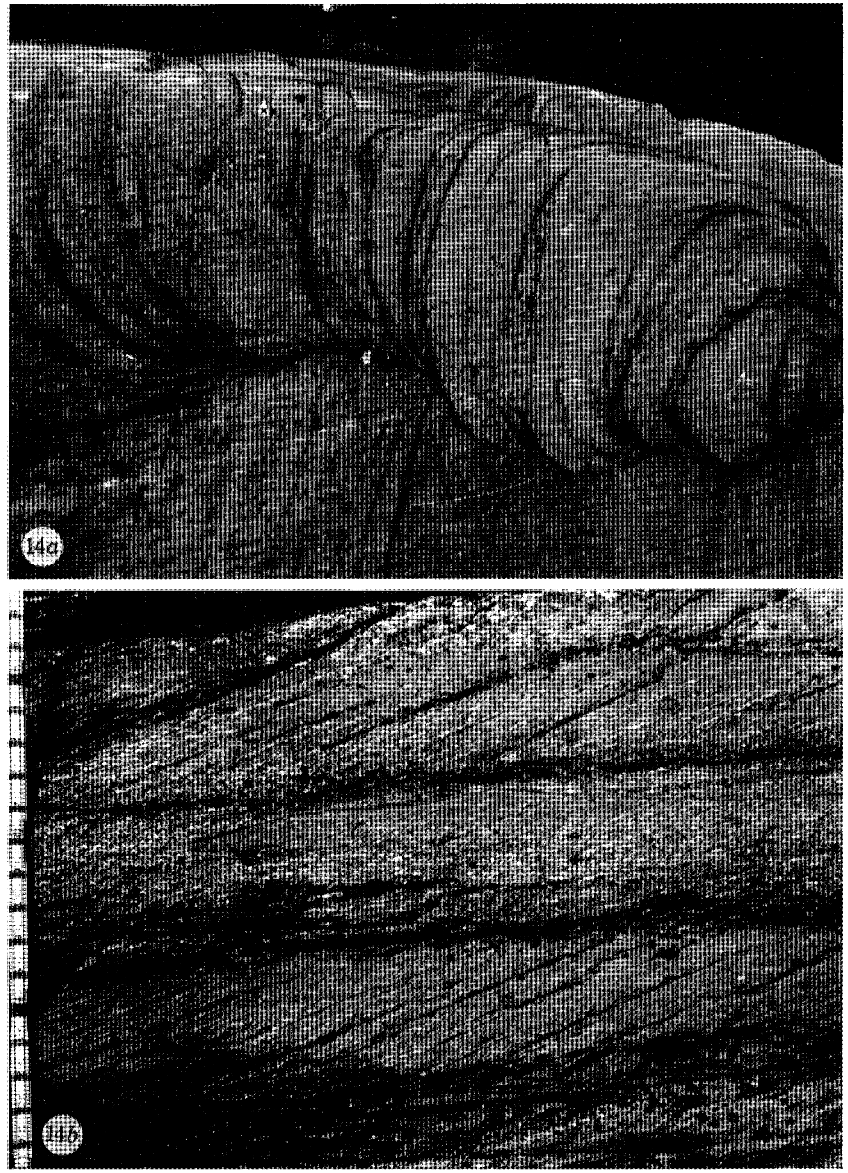


FIGURE 9. (*a*) Inclined air photograph of folds in dolomite of the Rocknest Formation, north of Kikerk Lake. The fold hinges trend easterly and the folds have steep south-dipping limbs and gentle north-dipping limbs. The middleground of the photograph shows about 1 km of the fold belt, which is related to a large basement uplift to the right (i.e. north) of the photograph. These folds are perpendicular to the folds in the thrust zone, about 2 km to the west. (*b*) Overturned unconformity on the southern margin of the large basement uplift north of Eokuk Lake. The upper part of the hill is Archaean granite and gneiss, and the lower slopes have about 40 m of overturned basal Odjick Formation. The unconformity dips about 40° but there has been no faulting or loss of section.

FIGURE 14. (*a*) Bedding surface of trough-type cross-bedding in subarkose of the Hornby Channel Formation, north Preble Island. Inferred palaeocurrent direction is from the left of the photograph, which shows about 1 m of the outcrop. (*b*) Cross-section of trough-type cross-bedding in the Hornby Channel Formation, south Simpson Island. Inferred palaeocurrent direction is from the right of the photograph. The scale is divided into 3 cm intervals.

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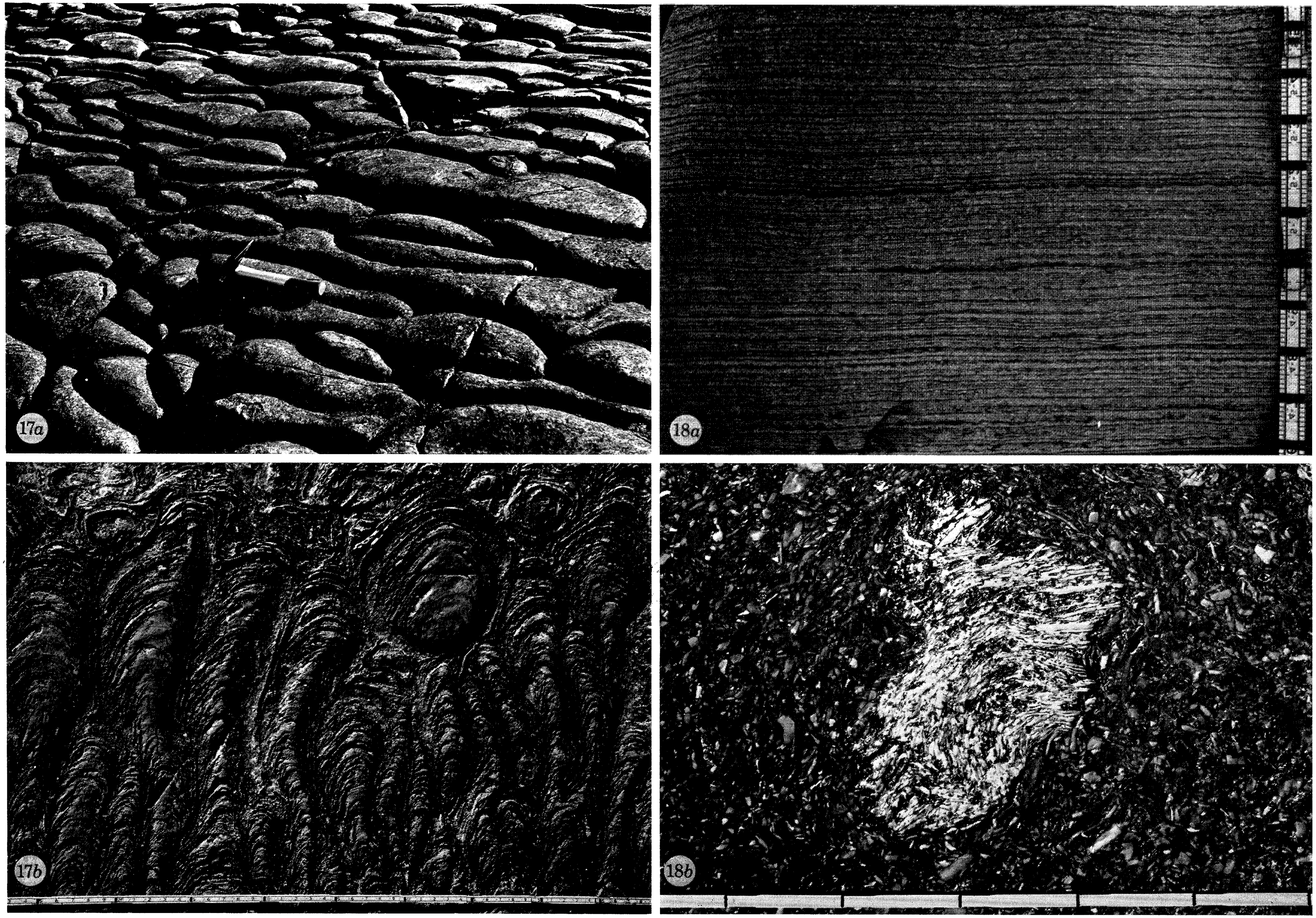


FIGURE 17. (*a*) Bedding surface of columnar stromatolites from the upper part of a 20 m thick bed at the top of the Taltheilei Formation, Utsingi Point. The elongation of the stromatolites, a very common feature, is everywhere perpendicular to the edge of the platform and parallel to paleocurrents determined from ripple marks in associated beds. (*b*) Relatively small elongate columnar stromatolites in the Taltheilei Formation, northeast Blanchet Island. The scale is divided into 3 cm intervals.

FIGURE 18. (*a*) Grey laminated shaly limestone in the Pekanatui Point Formation, south Blanchet Island. Closer to the edge of the platform, the dark shaly laminations become more widely spaced. The scale is divided into 3 cm intervals. (*b*) Bedding surface of debris-flow breccia bed in the Pekanatui Point Formation, Fairchild Point. The breccia consists of twisted blocks of thin-bedded limestone and is interstratified concordantly with non-brecciated beds of the same facies. The scale is divided into 15 cm intervals.

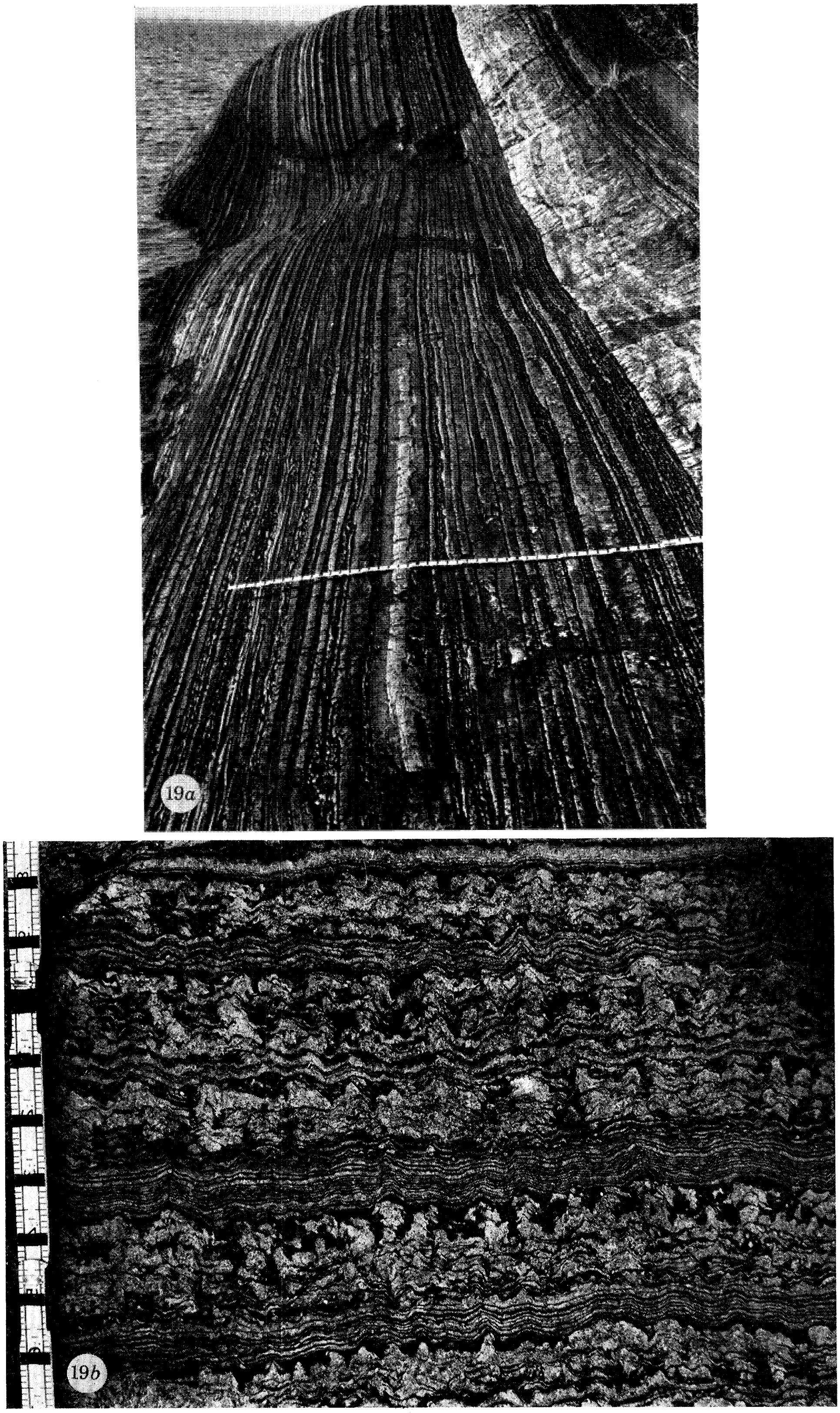
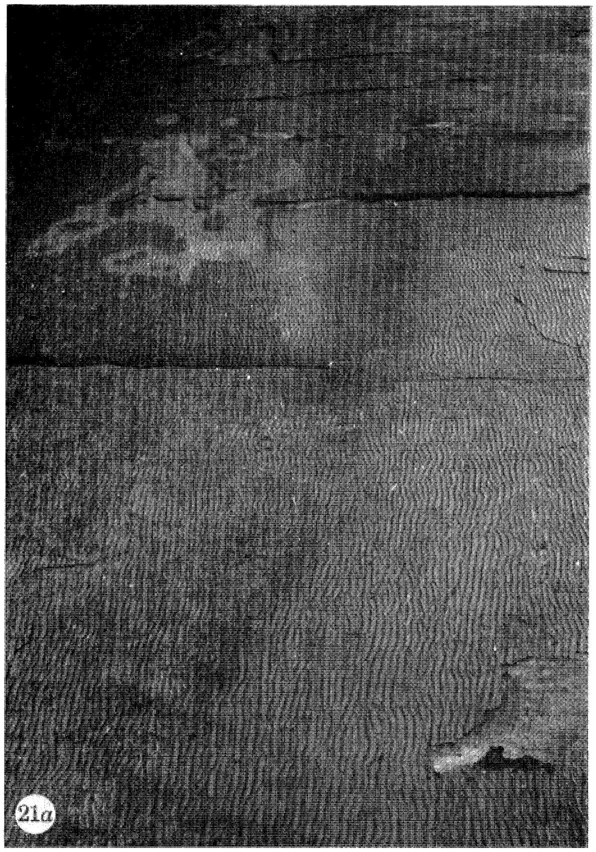
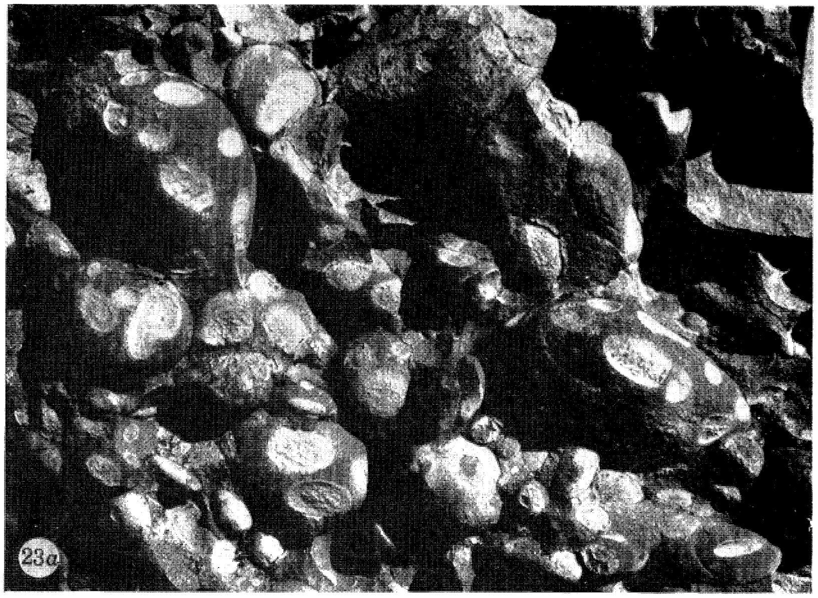
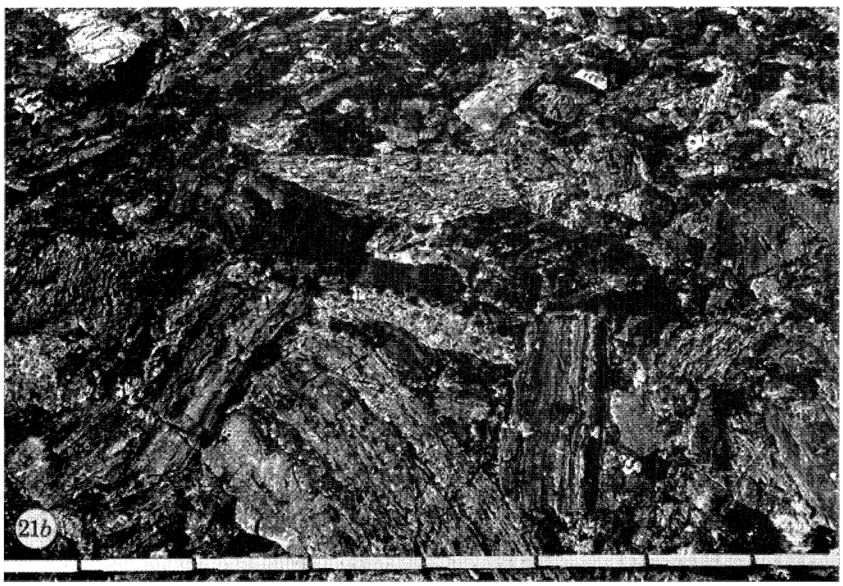


FIGURE 19. (*a*) Thin-bedded greywacke turbidites interstratified with laminated shaly limestone in the Blanchet Formation, south Blanchet Island. The scale is divided into 3 cm intervals. (*b*) Dark green mudstone with digitate calcareous growth structures in the McLean Formation, south Blanchet Island. This facies is repeatedly interstratified with greywacke turbidites and grades northward into digitate loferites of the Utsingi Formation on the platform. The scale is divided into 3 cm intervals.







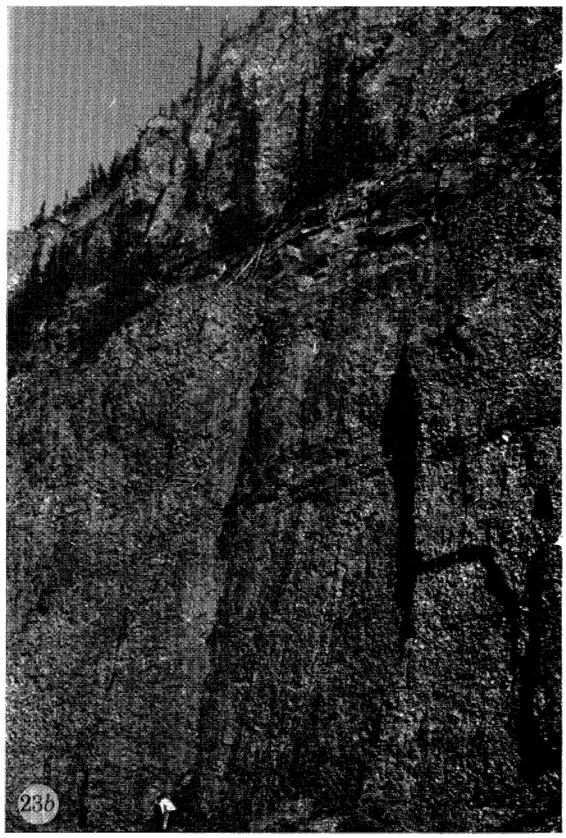


FIGURE 21. (*a*) Bedding surface of red rippled lithic sandstone in the Tochatwi Formation, east Stark Lake. (*b*) Breccia containing closely spaced angular blocks of rippled and stromatolitic carbonate in a matrix of red mudstone. Commonly, the blocks are more dispersed and of much more variable size than in this outcrop. The scale is divided into 15 cm intervals.

FIGURE 23. (*a*) Fanglomerate in the Murky Formation composed of quartzite boulders from the Kluziai Formation. The largest boulders in the photograph have diameters of 50 cm and their superficial red pigment has been leached around their points of contact, resulting in their spotted appearance. (*b*) Two thick fanglomerate beds separated by a thinner sandstone at the base of the trees. Note the lack of stratification in the fanglomerate. A geologist is at the base of the cliff for scale.

Calc-flysh phase

The contrast in facies between the platform and the aulacogen is most dramatic in the calc-flysch phase (see figure 16). On the platform are 400 m of stromatolitic, loferitic (Fischer 1964) and oolitic limestone and dolomite (see figure 17, plate 12) deposited in shallow water. At the northeast end of the aulacogen, correlative rocks consist of 250 m of non-stromatolitic thin-bedded argillaceous limestone and mudstone with digitate calcareous growth structures (see figures 18 and 19, plates 12 and 13). The trough facies thicken to 500 m at the southwest end of the aulacogen with the addition of westerly-derived tongues of graded greywacke beds, deposited

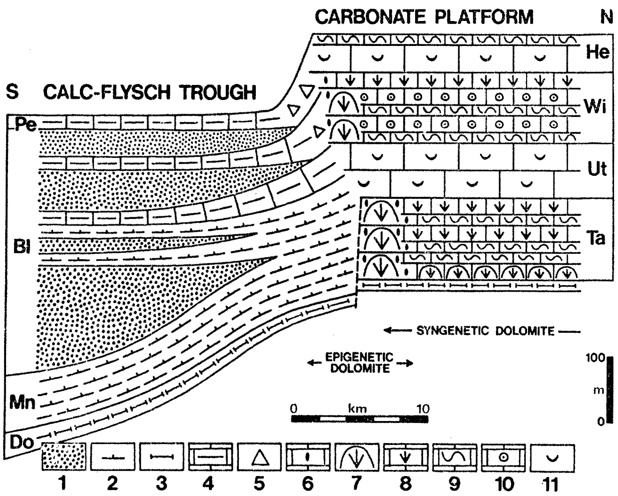


FIGURE 16. Diagrammatic stratigraphic cross-section of the Pethei group. The platform sequence consists of the Taltheilei Formation (Ta), the Utsingi Formation (Ut), the Wildbread Formation (Wi) and the Hearne Formation (He). The trough sequence consists of the Douglas Peninsula Formation (Do), the McLean Formation (Mn), the Blanchet Formation (Bl) and the Pekanatui Point Formation (Pe).

Unit 1. Greywacke turbidites (see figure 19).

Unit 2. Dark green and brown mudstone with digitate calcareous growth structures (see figure 19).

Unit 3. Red laminated shaly limestone.

Unit 4. Grey rhythmic thin-bedded limestone with shaly partings and laminated shaly limestone (see figure 18).

Unit 5. Grey rhythmic thin-bedded limestone with debris flows composed of blocks of thin-bedded limestone (see figure 18).

Unit 6. Crossbedded intraclast grainstone filling tidal channels between stromatolite mounds.

Unit 7. Large mounds of columnar stromatolites.

Unit 8. Thick beds of columnar stromatolites (see figure 17).

Unit 9. Undulatory stromatolites and discoidal oncolites.

Unit 10. Rippled ooid grainstone.

Unit 11. Digitate loferite, or birdseye limestone.

Dolomitization is extensive only in the Taltheilei Formation.

by turbidity currents that flowed up the axis of the aulacogen parallel to the edge of the shallow-water carbonate platform. The platform edge is marked by a belt of stromatolite mounds, up to 80 m in diameter and 20 m thick, separated by anastomosing tidal channels, oriented perpendicular to the edge of the platform, filled by crossbedded intraclast grainstone. Beds of limestone debris-flow breccia and sedimentary slump folds and thrusts occur locally in the trough facies close to the edge of the platform. The facies transition from the platform to the trough is commonly obscured by growth faults and vuggy coarsely crystalline dolomite.

Molasse phase

The molasse is better exposed in the aulacogen than in the geosyncline but has been almost completely eroded from the platform (see figure 20). Sharply overlying the calc-flysch are 650 m of red mudstone with halite casts and extensive olistostromes, in which angular blocks of stromatolitic dolomite and limestone are chaotically dispersed in a red mudstone matrix (see figure 21, plate 14). The largest blocks, commonly emplaced upside-down or recumbently folded, reach 45 m in thickness and 1000 m in length. At the top of the red mudstone succession are many thin beds of carbonate-pebble conglomerate.

Above the red mudstone and its olistostromes are 850 m of red laminated lithic sandstone in tabular crossbedded units up to 5 m thick with abundant mudcracks and ripple marks (see figure 21). Carbonate-pebble conglomerate occurs in the lower part of the succession and the sandstone contains a high proportion of sedimentary and silicic volcanic rock fragments. The sandstone is lithologically indistinguishable from the molasse of the geosyncline and, like it, is derived from the west, having been transported northeastward up the axis of the aulacogen.

The sandstone is overlain by 230 m of red mudcracked mudstone with buff siltstone beds and abundant halite and gypsum casts. The youngest rocks preserved beneath the fanglomerates are 180 m of columnar basalt flows.

Fanglomerate phase

Overlying the upturned edges of the molasse and older rocks in the aulacogen are up to 4000 m of red and buff alluvial fanglomerate and sandstone (see figure 22), nearly flat-lying except close to major faults. The fanglomerate occurs in unbedded cliff-forming units up to 30 m thick (see figure 23, plate 14) and consists of rounded boulders, locally up to 1 m in diameter, derived from older sedimentary rocks shed from the uplifted margins of the aulacogen or horsts within the aulacogen. The boulders have a matrix of calcite-cemented lithic sandstone, or rarely mudstone, and there are thin sandstone lenses within the fanglomerate units. The units are separated by intervals of red mudcracked mudstone and siltstone, less than 5 m thick, locally enclosing pedestal-shaped calcareous stromatolites, up to 2 m high, with a digitate internal structure. Locally, along the southeast margin of the aulacogen, from which most of the sediment came, are fossil talus breccias of monolithologic angular blocks. Near the southwest end of the aulacogen are basalt flows with zeolite-filled amygdules.

The fanglomerates are intergradationally overlain by cross-bedded sandstone with quartz and granite pebbles. The sandstone contains both sedimentary lithic fragments and clasts derived from unroofed Archaean basement. Where intercalated with mudstone, the sandstone occurs in fining-upward cycles with erosional bottoms. Sediment transport is locally variable but there is a prevailing trend southwestward down the axis of the aulacogen.

Synthesis of the depositional history

The aulacogen was a remarkably persistent trough that extended from the geosyncline far into the platform. It subsided deeply, relative to the adjacent platform, during every phase in the depositional history of the geosyncline. Sediment transport in the aulacogen was generally longitudinal, rather than transverse as in the geosyncline.

A three-stage evolution of the aulacogen can be ascertained by considering whether the adjacent parts of the platform were elevated or depressed. This seemingly innocent distinction is critical in interpreting the structural history and origin of the aulacogen.

In the first stage, deposits of the pre-quartzite through dolomite phases accumulated in the aulacogen but are missing from the adjacent platform (see figure 11). This suggests that the aulacogen was fault-bounded and had Archaean basement rocks exposed on either side (see figure 24). The proximity of basement rocks exposed along marginal fault scarps perhaps

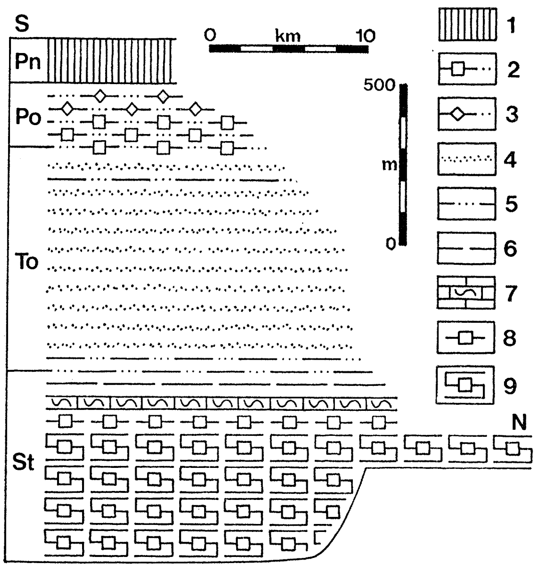


FIGURE 20

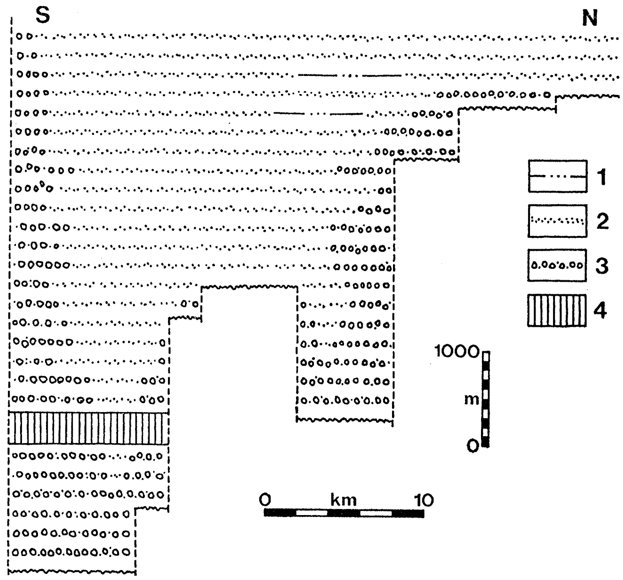


FIGURE 22

FIGURE 20. Diagrammatic stratigraphic cross-section of the Christie Bay group. The molasse phase consists of the Stark Formation (St), the Tochatwi Formation (To), the Portage Inlet Formation (Po) and the Pearson Formation (Pn).
Unit 1. Columnar basalt flows.
Unit 2. Red rippled and mudcracked siltstone with halite casts.
Unit 3. Red rippled and mudcracked siltstone with gypsum casts.
Unit 4. Red crossbedded lithic sandstone.
Unit 5. Red rippled and mudcracked siltstone with carbonate pebbles.
Unit 6. Red mudstone.
Unit 7. Stromatolitic and rippled limestone and dolomite.
Unit 8. Red mudstone with halite casts.
Unit 9. Red mudstone with olistostromes containing carbonate blocks up to 1 km in length.

FIGURE 22. Diagrammatic stratigraphic cross-section of the Et-then Group.
Unit 1. Red rippled and mudcracked siltstone.
Unit 2. Red and buff pebbly lithic and arkosic sandstone (Preble Formation).
Unit 3. Red lithic fanglomerate in units up to 30 m thick (Murky Formation).
Unit 4. Columnar and amygdaloidal basalt flows.

explains why the quartzite is coarser grained and more feldspathic in the aulacogen than on the shelf of the geosyncline, and why the dolomite phase contains quartzite beds only in the aulacogen.

The pre-flysch marks an important turning point in both the geosyncline and the aulacogen. In the geosyncline, foundering of the shelf was followed by the development of a westerly derived flysch wedge (see figure 24). Adjacent to the aulacogen, sediments for the first time accumulated on the platform (see figure 11), indicating that the Archaean basement there was being depressed, rather than standing high as it had earlier.

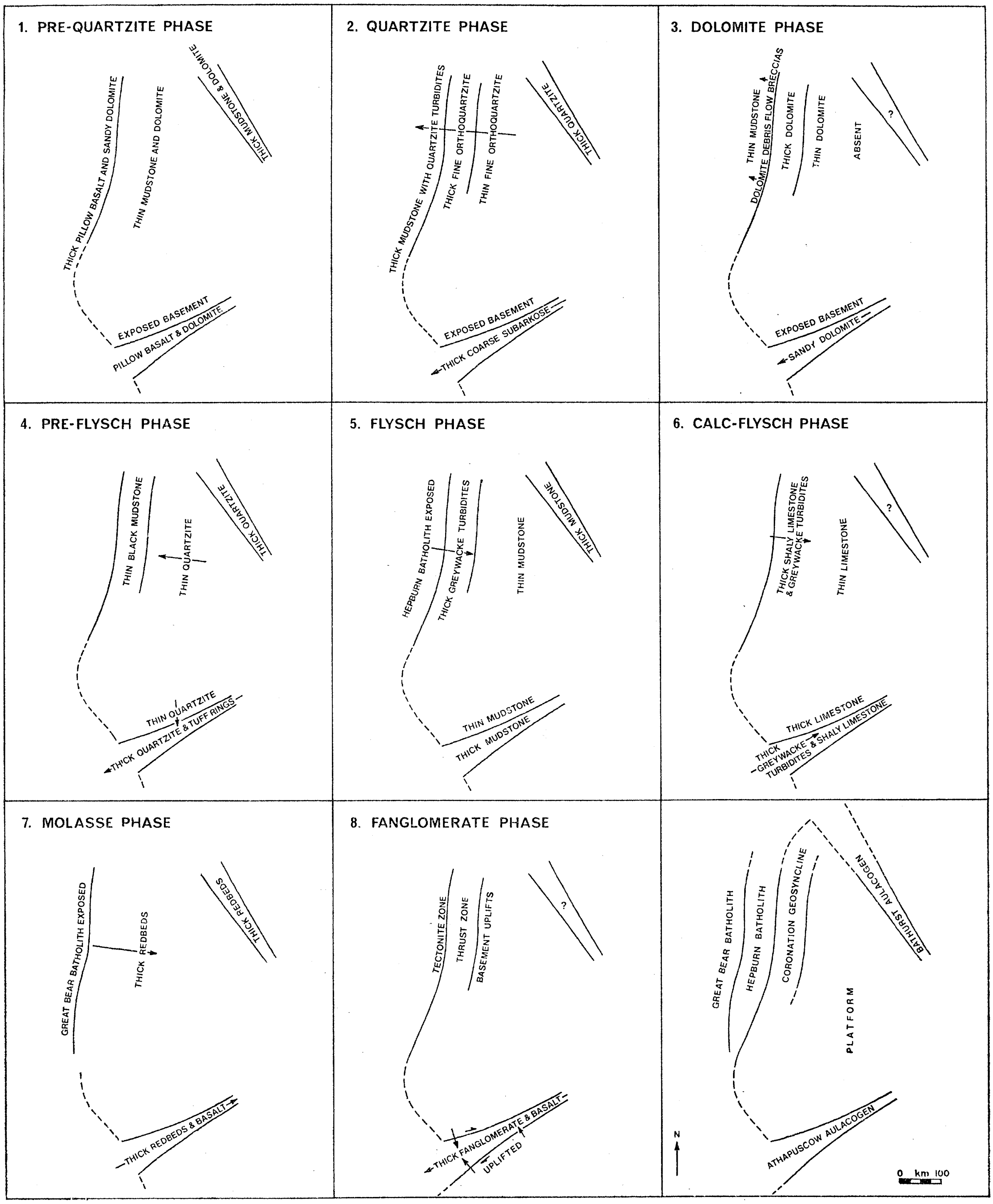


FIGURE 24. Summary of the depositional histories of the Coronation geosyncline and the Athapuscow aulacogen. The arrows indicate the regional palaeocurrent directions, generalized from 12000 measurements. The area south of the Athapuscow aulacogen is difficult to interpret because the early Proterozoic supacrustal rocks were eroded away during the fanglomerate phase.

The pre-flysch phase within the aulacogen was marked by extensive volcanism. The volcanic rocks filled the southwest end of the aulacogen and spilled out onto the adjacent platform where, in some places, they lie directly on Archaean rocks. The accumulation of volcanics prevented the establishment of deep water in the aulacogen, thus producing a temporary barrier to the influx of greywacke turbidites during the flysch phase (see figure 24).

The platform marginal to the aulacogen continued to subside, although not so much as the aulacogen itself, until the molasse phase (see figure 11). Growth faults bounding the aulacogen continued to influence sedimentation – notably in the lateral platform-to-trough facies zonation of the calc-flysch phase (see figure 16) and possibly in the emplacement of the olistostromes in the lower part of the molasse phase. Basalts were extruded, again only in the aulacogen, at the end of the molasse phase.

The third stage in the evolution of the aulacogen corresponds to the fanglomerate phase of deposition, which followed a period of compressional deformation at the end of the molasse phase. Sedimentation was once again fault-controlled and limited to the aulacogen. For the first time since the dolomite phase, the adjacent platform was the site of erosion (see figure 24).

Magmatism and structural evolution

Experimental analysis of fracture patterns (Cloos 1955) indicates that there are at least three different ways of producing a fault-bounded trough (see figure 25). In the first way, the classic graben is produced by extension of the crust, commonly over a broad upwarp. Characteristically, the graben is bounded by normal faults, the lips of the graben stand high, the peripheral supracrustal rocks dip gently away from the graben, the supracrustal rocks within the graben are undeformed, and the base of the crust rises beneath the graben. A second type of fault-bounded trough, of different geometry, is produced by dropping a slice of crust downward into the mantle. In this case, the lips of the trough sag inward such that the trough becomes bounded by high-angle reverse faults, the peripheral supracrustal rocks dip gently toward the trough creating a broad downwarp, the supracrustal rocks within the trough are folded by transverse compression, and the base of the crust is depressed beneath the trough. The third type of trough, of which perhaps the best example is the Ridge basin (Crowell 1954) on the San Andreas–San Gabriel fault system of California, is produced by transcurrent faults. This type of trough can be recognized by the braided pattern of splay faults in plan view and by the tendency for the supracrustal rocks within the trough to be homoclinally tilted with dips down the axis of the trough.

The Athapuscow aulacogen is interpreted as having been a trough of each type during its three-stage evolution (see figure 26).

Rifting stage

During the pre-quartzite through dolomite phases of deposition, the aulacogen resembled a graben produced by crustal extension or incipient rifting. The lips of the aulacogen stood high and shed sediment into the trough. The sediments within the trough were undeformed. Basaltic volcanism occurred during both the pre-quartzite and quartzite phases and its chemistry, based on limited data (very limited – two rapid analyses), is subalkalic, as might be expected in an active rift with thinned crust.

Sagging stage

With the beginning of the pre-flysch phase, the aulacogen changed character. The lips began to sag and sediments accumulated in a broad synclinal depression centred over the aulacogen. At the end of the molasse phase, the sediments within the aulacogen were compressed into broad folds and those on the adjacent platform tilted toward the aulacogen (see figure 12). Laccoliths, up to 25 km in length (see figure 10), of granodiorite and tonalite, were intruded

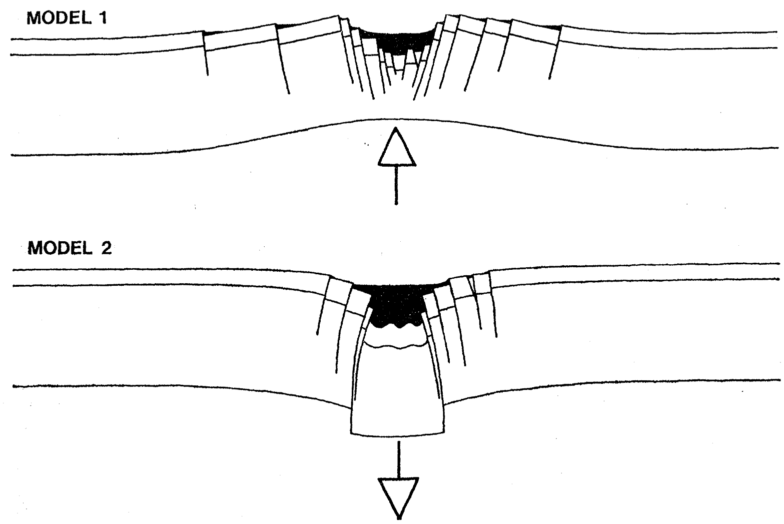


FIGURE 25. Contrasting geometric properties of fault-bounded troughs produced by extension over an uplifted arch (model 1) and by the down-dropping of an unsupported central block (model 2).

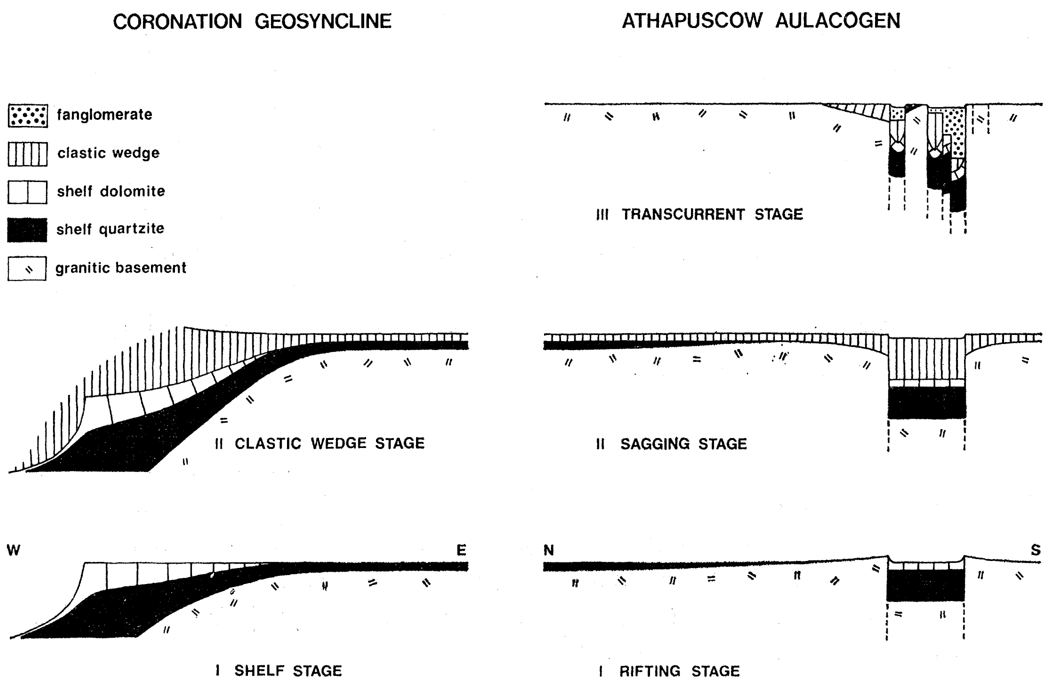


FIGURE 26. Proposed three-stage evolution of the Athapuscow aulacogen and its relation to the evolution of the Coronation geosyncline.

into the lower part of the molasse phase and volcanism, predominantly basaltic, occurred during the pre-flysch and molasse phases. The three available analyses of the basalts are strongly alkalic, as might be expected in an area of thickened crust undergoing compression.

The sagging model may also explain the results of a crustal refraction traverse across the aulacogen, from which the base of the crust is calculated to be about 4 km deeper beneath the aulacogen than beneath the platforms on either side (Barr 1971).

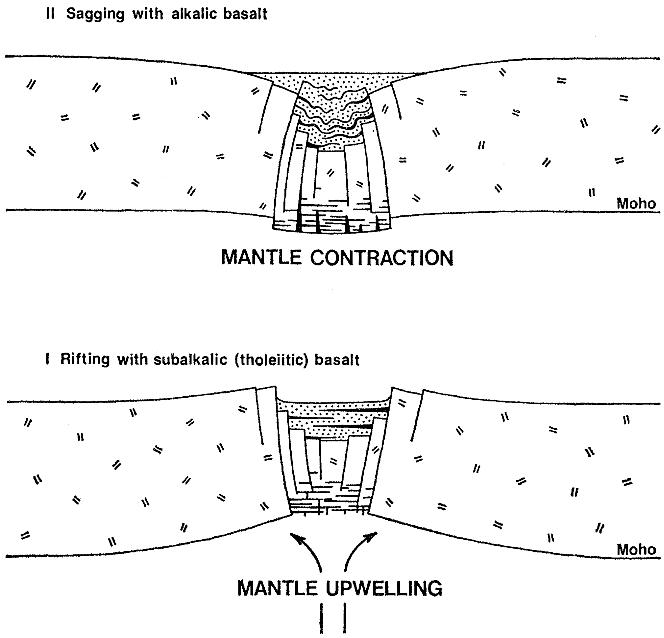


FIGURE 27. Model for the transition from incipient rifting to crustal sagging in aulacogens in response to abandonment of a zone of mantle upwelling.

Transcurrent stage

Sedimentation during the fanglomerate phase was controlled by a complex pattern of uplifted and downdropped blocks, separated by a braided network of faults, many of which are 100 km or more in length (see figure 10). Close to the faults, the sediments are homoclinally folded with dip-directions subparallel to the fault traces. I earlier interpreted the faults as having vertical displacements only (Hoffman 1969), but 7 months in California has convinced me of the error of my ways. The vertical displacements I now interpret as being subsidiary to large-scale, probably dextral, transcurrent movement. I offer my apologies to Reinhardt (1969), whose late-stage-vertical interpretation of the McDonald fault system was at least in part a result of my arm-twisting.

Synthesis of the structural evolution

Three stages are recognized in the evolution of the aulacogen – an incipient rifting stage during which the lips of the fault-trough stood high, followed by a sagging stage during which the lips were depressed and the aulacogen was located in a broad downwarp, and a final stage during which time the aulacogen was part of a regional transcurrent fault system (see figure 26).

The unusual sagging stage can perhaps be understood by presuming that the initial rifting was the result of mantle upwelling (see figure 27). During the rifting stage, the crust would have been thinned beneath the aulacogen by distension and beneath its lips by erosion. Crustal density beneath the aulacogen might have been increased by the extrusion and intrusion of basaltic magma. Cessation of mantle upwelling would have caused the relatively thin and dense crust in the vicinity of the aulacogen to sag because of mantle contraction resulting from heat dissipation. According to this model, broad downwarps with central fault-troughs should be the rule, not the exception over abandoned rifts.

The model contains no provision, so far as I can see, for the aulacogen to become a transcurrent system in its final stage.

Summary of the Athapuscow aulacogen

The aulacogen developed as a deeply subsiding trough that extended from the geosyncline far into the platform and in which sedimentary rocks much thicker than on the adjacent platform accumulated during every phase of the geosyncline. The succession in the aulacogen is marked by at least five periods of basaltic volcanism, only the oldest of which occurs elsewhere, and by the intrusion of tonalite to granodiorite laccoliths.

During the construction of the shelf along the western margin of the platform, the aulacogen was an incipient rift merging at a high angle with the shelf. As the shelf foundered and was buried by the clastic wedge derived from the batholithic belt to the west, rifting of the aulacogen ceased. The trough continued to subside, at times bounded by growth faults, in the centre of a broad downwarp and was ultimately compressed, producing broad folds with axes parallel to the trend of the aulacogen. Finally, the aulacogen became part of a regional transcurrent fault system in which thick alluvial sediments were deposited in scattered down-dropped basins between active splay faults.

Bathurst aulacogen

The Goulburn basin (see figure 1), which I have not seen and therefore cannot describe in detail, is similar in many respects to the Athapuscow aulacogen. A thick succession of supracrustal rocks correlative with those in the Coronation geosyncline (see table 1), accumulated along what was to become the Bathurst fault system (see figure 2). To the southwest, on the adjacent platform, correlative rocks are much thinner and less deformed. Like the Athapuscow aulacogen, the Bathurst aulacogen is bounded on both sides by Archaean basement rocks, not flanked by intrusive batholiths as is the geosyncline itself.

Batholithic belt

The batholithic belt is split by the north-trending 350 km long Wopmay River fault (see figure 2). East of the fault, metamorphosed supracrustal rocks co-extensive with the tectonite zone of the Epworth basin are intruded by the mesozonal Hepburn batholith and various smaller igneous bodies. West of the fault, the epizonal Great Bear batholith intrudes its own extrusive equivalents and sedimentary rocks derived from them. The descriptions below are based on minimal reconnaissance mapping west of the Epworth basin (see figure 3).

Hepburn batholith

Batholithic rocks

The Hepburn batholith is a composite intrusion of massive to weakly foliated, commonly porpyroblastic, biotite granodiorite. Migmatite, with melanosomes of sillimanitic paragneiss and garnetiferous amphibolite, occurs along the margins of the individual plutons and around the periphery of the batholith. Foliation is concordant with that in the surrounding metamorphic rocks and dips steeply in the central part of the batholith but progressively more gently inward toward the eastern margin, where the batholithic rocks overlie metamorphic rocks of the Epworth basin (see figure 4).

Discordant plutons, up to **3** km in diameter, of leucocratic equigranular adamellite occur along the margins of the batholith or within metamorphic rocks close to the batholith. Discordant masses of anorthositic gabbro also occur along the margins of the batholith.

Pre-batholithic rocks

Metamorphosed and intensely deformed supracrustal rocks are exposed in a belt, up to 32 km wide, within the Hepburn batholith (see figure 3). The grade of metamorphism is highly variable and northwest-trending sinistral transcurrent faults commonly juxtapose rocks of greenschist and upper amphibolite facies. Migmatite, with melanosomes of porphyroblastic garnet-sillimanite paragneiss and garnetiferous amphibolite, occur in broad zones peripheral to the batholith. In the interior of the metamorphic belt, pelitic schists contain prophyroblasts of cordierite, andalusite, garnet and rarely staurolite. Kyanite is absent and the metamorphism is of a low-pressure intermediate facies series (Miyashiro 1961).

The original character of the rocks can be ascertained in general terms but, because of the structural complexity, their internal stratigraphy cannot be known with certainty without much more detailed mapping. There are three lithologically distinct successions of which the first, structurally the lowest, contains great thicknesses of arkose and granite-pebble conglomerate with subsidiary pelitic and volcanic rocks, intruded by thick gabbro sills. The second assemblage, intergradational with the first, is composed mainly of mudstone with thin graded beds of feldspathic quartzite and is intruded by sills or laccoliths of granodiorite or tonalite, in most places converted into banded and lineated blastomylonite. The third succession, structurally the highest, consists of many hundreds of metres of pillow basalt and gabbro sills with minor intercalations of mudstone and bedded chert.

All of the metamorphic rocks are penetratively deformed and most have a well-developed, north-trending, gently plunging lineation. Deformed pebbles and other strain indicators are spindle-shaped, with axial ratios of more than 10:1, and oriented parallel to the lineation. This is in contrast to the west-plunging strain elongation east of the batholith. Curved inclusion trains in poikilitic garnet, cordierite and andalusite indicate rotation during growth about axes parallel to the lineation.

Most of the rocks have a strong foliation that is compressed into upright chevron-shaped folds, with wavelengths of several kilometres and hinges that parallel the regional lineation. Bedding, discernible only in the least deformed rocks, is isoclinally folded in the plane of the foliation. The bedding fold hinges do not everywhere parallel the foliation fold hinges.

Great Bear batholith

Batholithic rocks

The epizonal plutons of the Great Bear batholith are composed of massive to porphyritic, nowhere foliated, granodiorite and adamellite, with subsidiary granite, tonalite and diorite. They have clearly magmatic textures and their mafic minerals, in decreasing order of abundance, are hornblende, biotite and uralitized pyroxene, mostly occurring together in clotted aggregates.

The margins of the plutons are either coarse grained and equigranular, or porphyritic with phenocrysts of plagioclase and quartz in a granophyric groundmass. Where porphyritic, the plutons are commonly separated from the wall rocks by a non-porphyritic fine-grained melanocratic diorite. Dense swarms of dykes of the same composition and texture as the prophyritic border phase of the plutons extend outward as far as 2 km into the wall rocks. The dykes, which may constitute 90 % of the total rock volume close to the plutons, are separated from the plutons by the border diorites.

Pre-batholithic rocks

The plutons discordantly intrude volcanic rocks, probably comagmatic with the plutons, and sedimentary rocks derived from them. The supracrustal rocks are broadly folded and metamorphism is limited to contact hornfelses not extending more than a few hundred metres from the plutons.

The volcanic rocks consist of great thicknesses of rhyodacitic crystal-rich welded ash-flow tuff, in single cooling units hundreds of metres thick. The welded tuff units are capped by thinner but complex sequences of trachybasalt flows, laharic breccias, laminated tuffaceous mudstone, cross-bedded lithic sandstone and volcanic boulderstone. The sedimentary rocks are derived exclusively from the associated volcanic and hypabyssal intrusive rocks.

The supracrustal rocks are, in places, intruded by plugs of vertically flow-banded felsite, commonly associated with radial dyke swarms. Many of the volcanic rocks have undergone severe hydrothermal alteration, particularly near plutons with sulphide mineralization.

Age relations and unroofing of the batholiths

The outstanding problems of the batholithic belt are the age relations of the two batholiths and the nature of the Wopmay River fault. The main difference between the two batholiths is the depth of subsequent erosion. The level at which the Great Bear batholith is exposed is epizonal and the plutons are still partly mantled by their extrusive equivalents. In contrast, the Hepburn batholith has been unroofed to mesozonal depths. It intrudes regionally metamorphosed supracrustal rocks of the Coronation geosyncline. Extrusive and epizonal equivalents of the Hepburn batholith, presuming such rocks existed, have long since been eroded away. Thus, the Wopmay River fault juxtaposes batholithic terrains eroded to different depths.

Petrology of the clastic wedge in the Coronation geosyncline, derived from the plutonic belt, indicates that the Wopmay River fault does not merely juxtapose different levels of the same batholith. The plutonic and metasedimentary rock fragments in the flysch phase came from a more deeply eroded source area than the volcanic and hypabyssal igneous rock fragments in the molasse phase. Because the flysch phase was deposited first, its plutonic source must have been relatively deeply eroded before the magmatism that supplied sediment to the molasse phase. Only the Hepburn batholith has been eroded deeply enough to have supplied sediment to the flysch phase and must, therefore, be older than the Great Bear batholith, the probable source for the molasse phase. The time interval between the two batholiths is perhaps represented in the clastic wedge by the calc-flysch phase, relatively deficient in terrigenous detritus.

The hypothesis that the flysch phase was deposited during unroofing of the Hepburn batholith is complicated by the fact that the flysch phase is metamorphosed near the eastern margin of the batholith (see figure 4). This requires that the beginning of flysch deposition, resulting from unroofing of the westernmost plutons, be succeeded by eastward migration of plutonism, ultimately with the flysch phase being intruded by the easternmost plutons. Thus, flysch deposition would be in part contemporaneous with the emplacement, as well as unroofing of the Hepburn batholith.

Summary of the batholithic belt

The batholithic belt of the Coronation geosyncline is composed of two batholiths separated by the north-trending Wopmay River fault – the mesozonal Hepburn batholith east of the fault and the epizonal Great Bear batholith west of the fault.

The Hepburn batholith intrudes rocks of the Coronation geosyncline and was probably emplaced and unroofed contemporaneously with deposition of the flysch phase in the geosyncline. Within the batholith are belts of intensely deformed and metamorphosed supracrustal rocks, including pillow basalt and granite-pebble conglomerate, perhaps correlative with the pre-quartzite phase of deposition in the tectonite zone of the Epworth basin.

Plutons of the Great Bear batholith are still partly mantled by comagmatic volcanic rocks and sediments derived from them. The volcanics include great thicknesses of welded rhyodacitic ash-flow tuff, complexly intercalated with thin trachybasalt flows, laharic breccias and hypabyssal intrusives. The Great Bear batholith is the probable souce of the silicic volcanic rock fragments in the molasse phase of the geosyncline and would, in that case, be younger than the Hepburn batholith.

Regional tectonic evolution

The Coronation geosyncline (see figure 28) developed along the western margin of a continental platform, the basement rocks of which consist of stabilized Archaean plutonics and metamorphics older than 2300 Ma. The geosyncline began with the deposition of shallow-water platform-derived quartzite and dolomite on a continental shelf, flanked to the west by a continental rise on which mudstone, with quartzite turbidites derived from the shelf edge, was deposited in relatively deep water above a basal sequence, now preserved only as intensely metamorphosed remnants in the Hepburn batholith, containing granite-pebble conglomerate and pillow basalt. Sediments deposited on the platform east of the geosyncline are much thinner than on the shelf or rise, except on two incipient rifts, the Athapuscow and Bathurst aulacogens, that extended at high angles to the shelf far into the interior of the platform.

The rise and eastward spreading of the Hepburn batholith in the area of the continental rise initiated a diachronous sequence of events – (1) the continental shelf foundered and was draped by a starved sequence of black pyritic mudstone, (2) coarse greywacke turbidites, derived from the batholithic terrain, spread eastward over the old rise and shelf sediments, (3) the greywacke turbidites and all older supracrustal rocks were thrust eastward above a basal detachment surface, and (4) the old rise and overlying greywackes were penetratively compressed, regionally metamorphosed and overridden by the advancing eastern margin of the batholith. East of the thrust front, the fine grained distal equivalents of the greywacke turbidites are overlain by relatively deep water shaly limestone that passes upward into non-marine red lithic sandstone containing silicic volcanic rock fragments, shed from the top of the Great Bear batholith, emplaced behind the older and more deeply eroded Hepburn batholith. Meanwhile, on the platform, large cold anticlinal basement uplifts were rising in front of the advancing thrust sheets and at least one of the old incipient rifts, the Athapuscow aulacogen, continued to subside deeply, by this time centred over a broad downwarp, and ultimately suffered mild transverse compression and was intruded by hypabyssal tonalite laccoliths.

Finally, the batholithic, supracrustal and basement rocks were all displaced by high-angle faults of regional extent, most of which trend northeast and had dominantly dextral transcurrent movement. The transcurrent faulting is most apparent in the northeast-trending Athapuscow aulacogen, where it was accompanied by the deposition of thick alluvial fanglomerates in localized fault-bounded troughs.

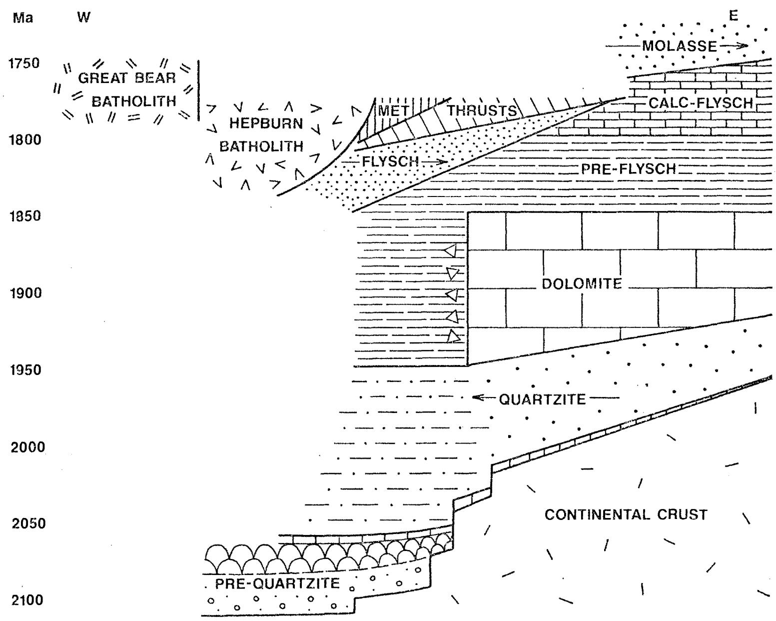


FIGURE 28. Diagrammatic time-distance cross-section of the Coronation geosyncline. The arrows indicate the directions of sediment transport. Metamorphism in the tectonite zone of the Epworth basin is represented by the area labelled MET. The time scale (not stratigraphic thickness) on the left is based on estimated rates of deposition and radiometric ages of the batholiths.

Comparison with Phanerozoic geosynclines

For all their differences in detail, Phanerozoic geosynclines have obvious features in common, most of which they share with the Coronation geosyncline. All develop marginal to continental platforms and are characterized, in their early stages, by paired miogeosynclines and eugeosynclinal facies belts. The miogeosynclines consist of platform-derived quartzite and shallow-water carbonate, whereas the eugeosynclines have mainly deep-water non-calcareous sediments, commonly with volcanics in the oldest parts.

At the beginning of the orogenic stage, the miogeosynclines founder and are draped by starved deep-water sediments – normally black mudstone in the pre-Mesozoic and shaly pelagic limestone and chert thereafter. These sediments are the precursors of thick clastic wedges that typically begin with greywacke turbidites (flysch) and grade upward into alluvial lithic sandstone (molasse).

A spectrum of orogenic events can be distinguished by the source and arrangement of the clastic wedges (Dewey & Bird 1970; Dickinson 1971). In Alpine-type geosynclines, the wedges are platform-directed and derived mainly from basement-cored nappes, in the root zones of which high-pressure metamorphism prevails and magmatism is relatively unimpressive. In Cordilleran-type geosynclines, wedges are shed from magmatic arcs with high-temperature metamorphism toward a platform on one side and toward an oceanic trench on the other. The clearest example of such a divergent pair of wedges is in the Cretaceous of western North America, where wedges were shed from the Sierra Nevada batholith eastward (the Rocky Mountain exogeosyncline) into the continental interior, and westward (the Great Valley sequence of California) toward a trench, now manifested by the underthrust Franciscan melanges of the Coast Mountains. Platform-directed overthrusting, without basement involvement at its extremities, follows in the wake of the platform-directed wedges. Anticlinal basement uplifts may rise on the platform in front of the advancing thrust sheets, as in the Rocky Mountains of Wyoming and adjacent states.

The post-orogenic stages of both Alpine-type and Cordilleran-type geosynclines is marked by continental sedimentation with minor volcanism controlled by high-angle faults.

The depositional and structural evolution of the Coronation geosyncline (see figure 28) mimics that of Phanerozoic Cordilleran-type geosynclines with remarkable fidelity. The only elements missing are the fossil trench and trench-directed clastic wedge, presumably buried beneath the Paleozoic cover west of the Great Bear batholith. All that is unusual about the Coronation Geosyncline is that, considering its antiquity, it is so well preserved.

Structures resembling the two aulacogens, on the other hand, are not so well known in the Phanerozoic and may, as certain Soviet geologists have suggested, be unique to Proterozoic platforms (Salop & Scheinmann 1969). Few Phanerozoic fault-troughs are so long-lived, although the Paleozoic Anadarko basin of Oklahoma (King 1959) may have a similar relationship to the Ouachita extension of the Appalachian geosyncline as the Athapuscow aulacogen has to the Coronation geosyncline. Moreover, subsurface exploration of modern continental margins has led to the discovery of buried deeply subsiding transverse fault-troughs, possibly incipient aulacogens, beneath several of the world's greatest rivers.

The beginnings of plate tectonics

The stimulus in the theory of plate tectonics is that it tries to see ancient geosynclines in terms of processes that are going on today. Modern continental margins are considered as geosynclines in various stages of evolution. In terms of plate tectonics, geosynclines are interpreted as paired shelf and rise sequences of aseismic Atlantic-type continental margins that undergo orogeny as a result of subsequent consumption of oceanic crust by subduction along trenches. Magmatism, particularly the emplacement of batholiths beneath volcanic arcs (Hamilton 1969), as in Cordilleran-type geosynclines, occurs principally in the region of high heat-flow above the inclined subduction zones. Ultimately, consumption of oceanic crust results in continental collision, as in Alpine-type geosynclines. Continental crust is not appreciably consumed because of its bouyancy, thus continental collision terminates subduction. The consumption of oceanic crust between collided continents accounts for its absence beneath ancient geosynclines, except as dismembered allochthonous slivers.

If Phanerozoic Cordilleran-type geosynclines are controlled by plate tectonics, surely the evolution of the Coronation geosyncline, so remarkably similar, must be as well. I cannot conceive of a totally different mechanism yielding an identical product. Reconstruction of the specific plate interactions involved in the Coronation geosyncline must await more detailed mapping of the batholithic belt and, probably, subsurface data from west of the Great Bear batholith. However, I believe it safe to conclude that, if the interpretation of Phanerozoic geosynclines in terms of plate tectonics is correct, plate tectonics was operative in the early Proterozoic.

This conclusion is particularly important in light of the contention, expressed in the introduction to this paper, that geosynclines do not occur in the Archaean and, by inference, that plate tectonics as we know it in the Phanerozoic did not exist. Thus, the early Proterozoic witnessed the first stabilization of large continental platforms, the first geosynclines – both Cordilleran types like the Coronation geosyncline and Alpine types like the Labrador geosyncline (Dimroth, Baragar, Bergeron & Jackson 1970) – and the beginnings of plate tectonics.

My field work in the Great Slave basin was the basis of a doctoral dissertation at The Johns Hopkins University under the supervision of F. J. Pettijohn and R. N. Ginsburg. This and subsequent field work in the Epworth basin was supported by the Geological Survey of Canada. I am indebted to J. C. McGlynn and J. A. Fraser of the Survey, and to my many associates in the field, particularly P. A. Geiser. The manuscript was prepared when I was a visiting lecturer at the Santa Barbara campus of the University of California, where I was much influenced by association with C. A. Hopson, J. C. Crowell and the Hardcore Study Group.

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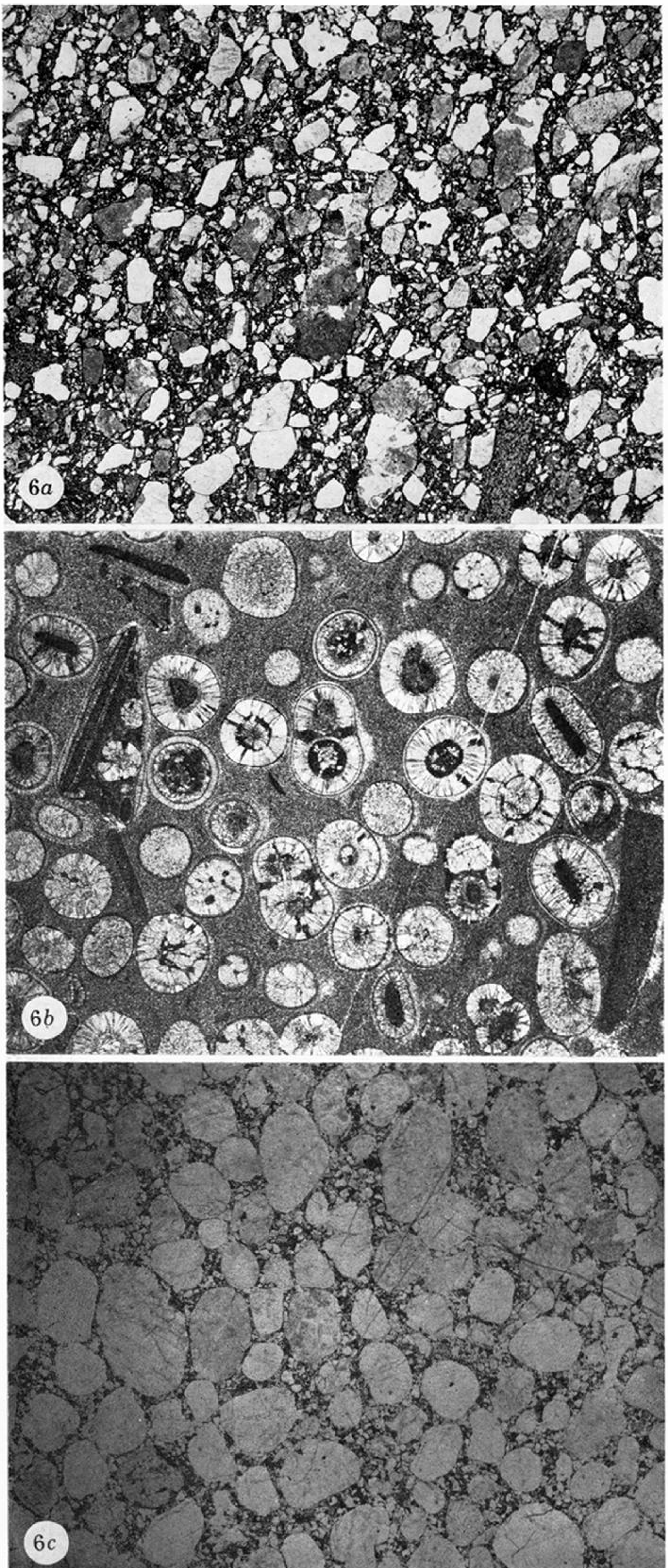


FIGURE 6. (*a*) Microphotograph of coarse greywacke from the Recluse Formation composed of unsorted angular fragments of quartz, plutonic and metamorphic rocks fragments set in a dark fine-grained matrix. (*b*) Dolomite from the Rocknest Formation composed of ooids, compound ooids and intraclasts set in a mainly fine-grained dolomite matrix. The high degree of textural preservation is typical of the formation despite complete dolomitization. (*c*) Orthoquartzite from the Odjick Formation composed of rounded coarse sand set in a matrix of coarse silt. The dark areas are void spaces.

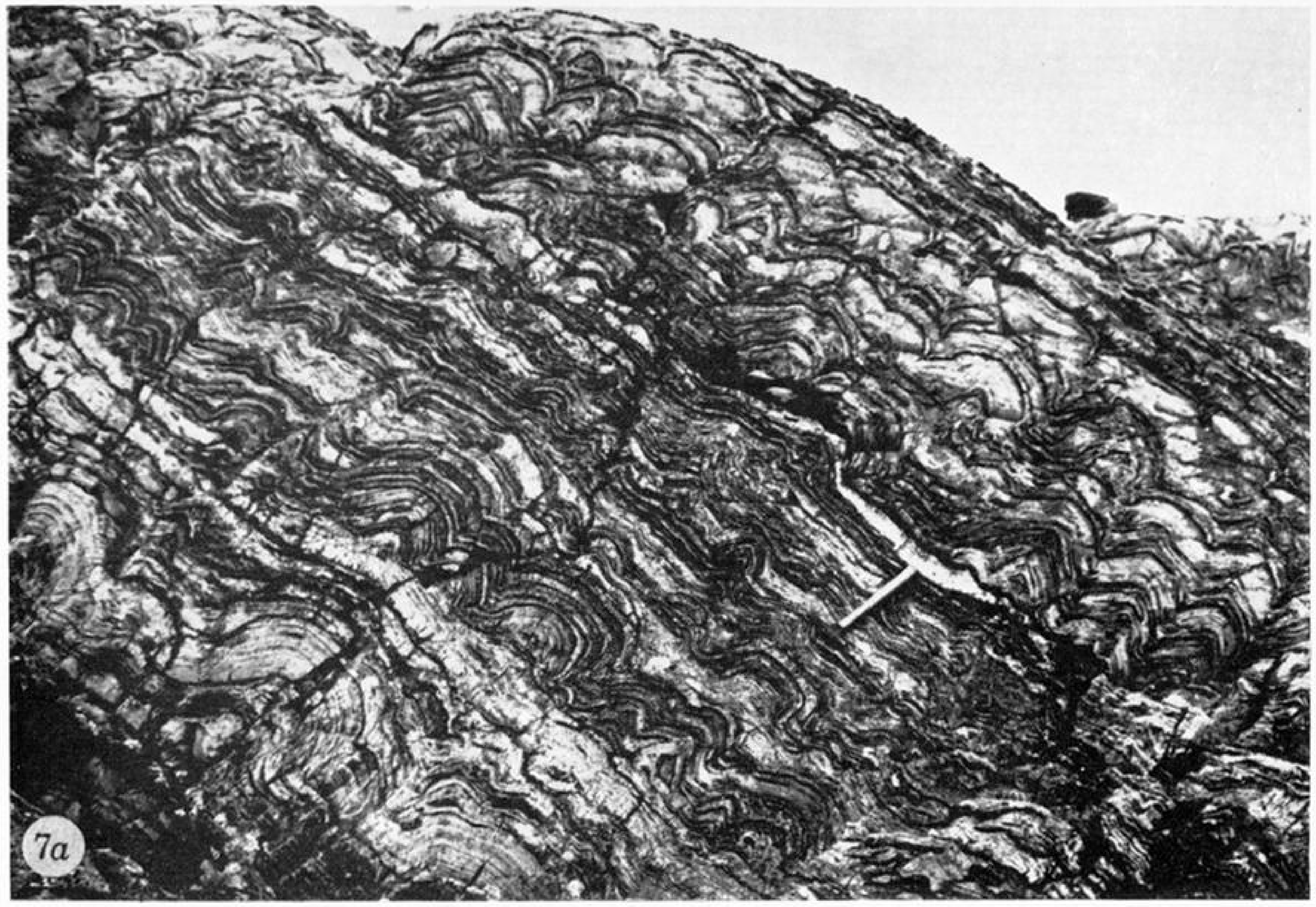
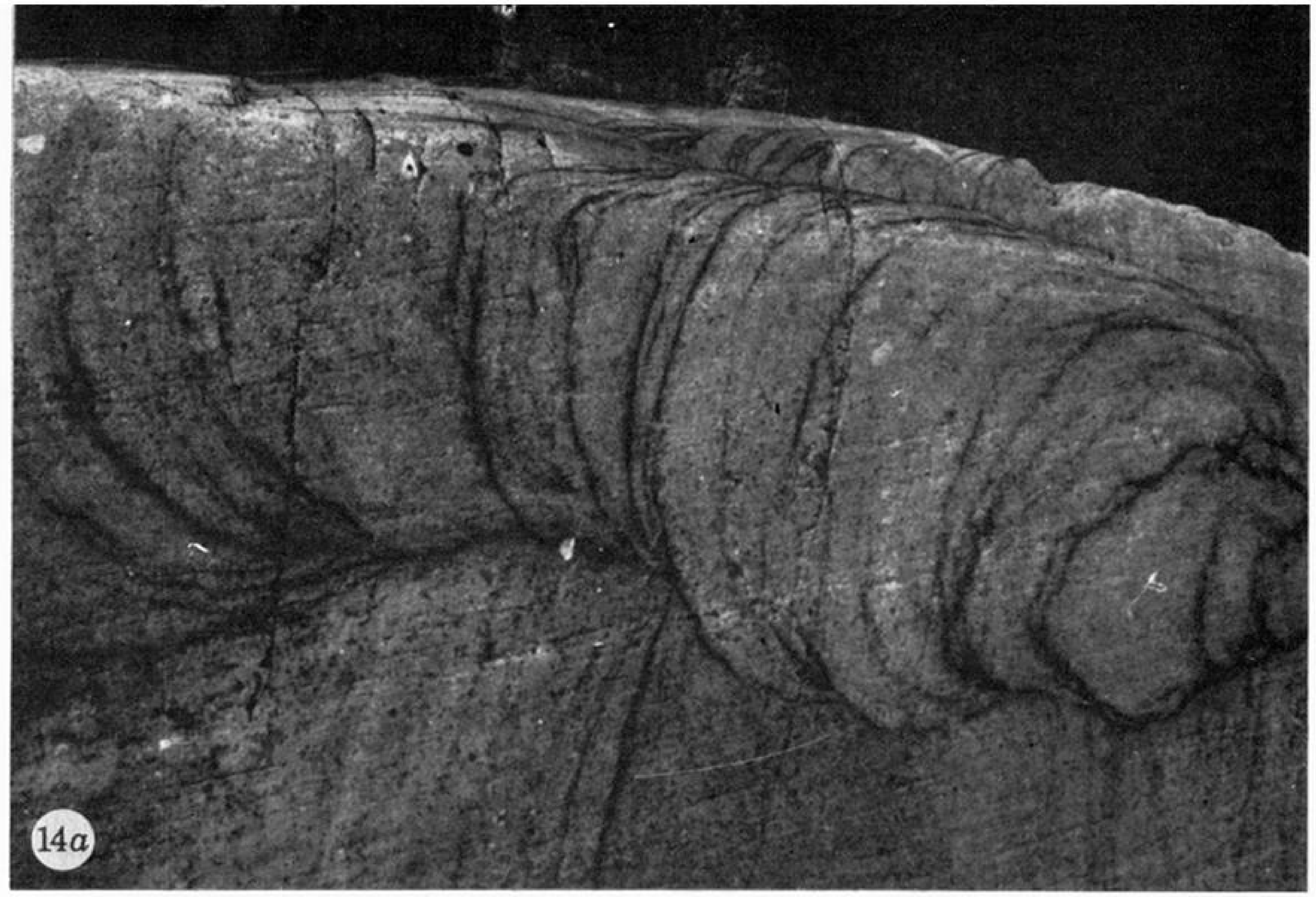


FIGURE 7. (*a*) Laterally linked domal stromatolites in cherty dolomite in the upper part of a shoaling-upward cycle in the Rocknest Formation, east of Eokuk Lake. Shaly dolomite of the lower part of the cycle can be seen in the lower left corner of the photograph. A geological pick is in the right-central part of the photograph for scale. (*b*) Cyclic alternation of dark shaly dolomite and light cherty stromatolitic dolomite in the Rocknest Formation north of Kikerk Lake. The photograph includes about 50 m of section in the foreground.

FIGURE 8. (*a*) Dark green mudstone with calcareous concretions in the Recluse Formation east of the feather edges of the greywacke turbidite tongues, Kikerk Lake. The scale is divided into 3 cm intervals (tenths of feet). (*b*) Inclined air photograph of greywacke turbidites in the Recluse Formation, near the east side of the Coppermine River at latitude 66° 30′. About 1000 m of steeply dipping beds are shown in the farground of the photograph, taken from an elevation of 600 m above the ground. Note the great lateral continuity of the beds, most of which are composed of several amalgamated turbidites.



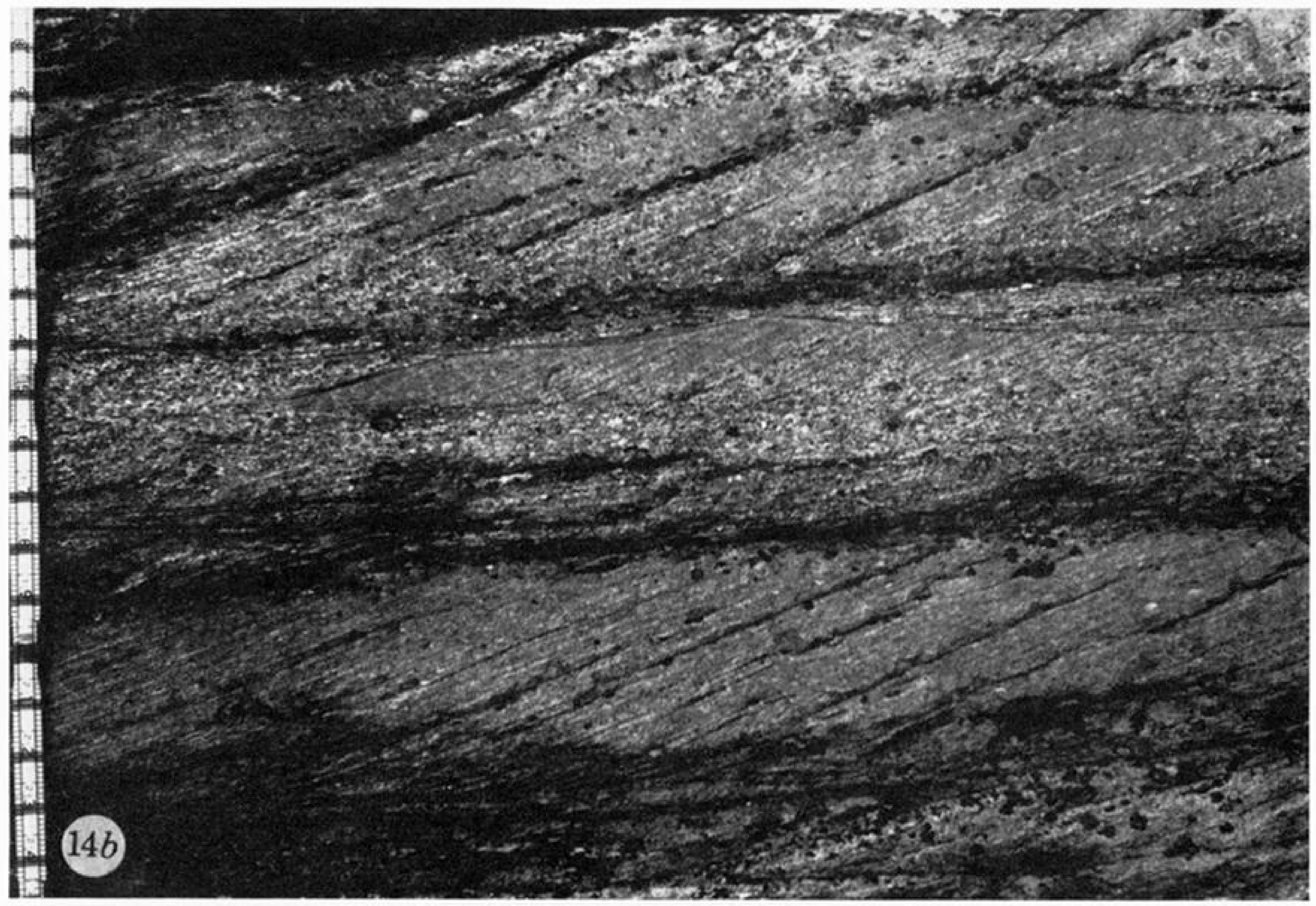
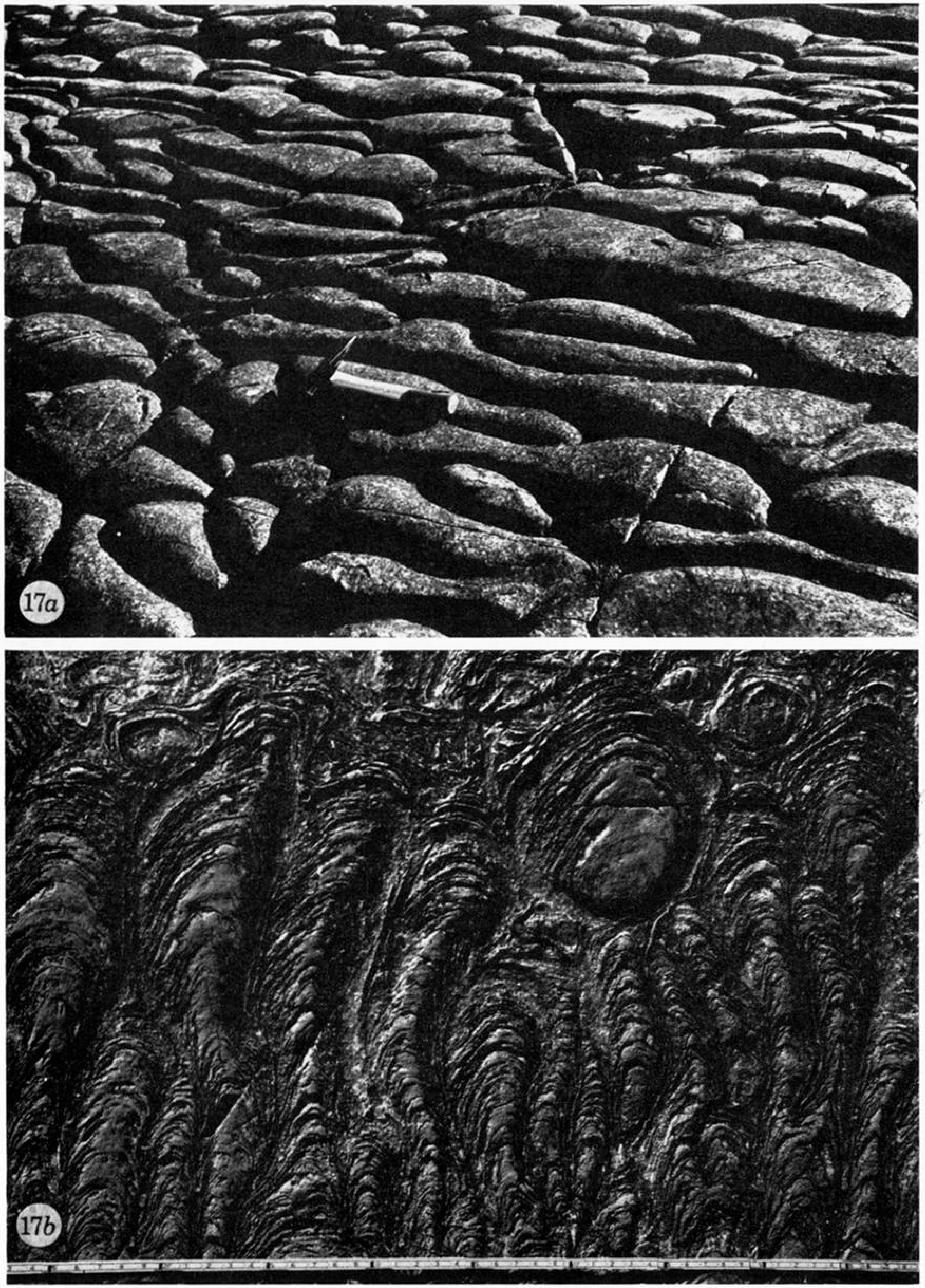


FIGURE 9. (*a*) Inclined air photograph of folds in dolomite of the Rocknest Formation, north of Kikerk Lake. The fold hinges trend easterly and the folds have steep south-dipping limbs and gentle north-dipping limbs. The middleground of the photograph shows about 1 km of the fold belt, which is related to a large basement uplift to the right (i.e. north) of the photograph. These folds are perpendicular to the folds in the thrust zone, about 2 km to the west. (*b*) Overturned unconformity on the southern margin of the large basement uplift north of Eokuk Lake. The upper part of the hill is Archaean granite and gneiss, and the lower slopes have about 40 m of overturned basal Odjick Formation. The unconformity dips about 40° but there has been no faulting or loss of section.

FIGURE 14. (*a*) Bedding surface of trough-type cross-bedding in subarkose of the Hornby Channel Formation, north Preble Island. Inferred palaeocurrent direction is from the left of the photograph, which shows about 1 m of the outcrop. (*b*) Cross-section of trough-type cross-bedding in the Hornby Channel Formation, south Simpson Island. Inferred palaeocurrent direction is from the right of the photograph. The scale is divided into 3 cm intervals.



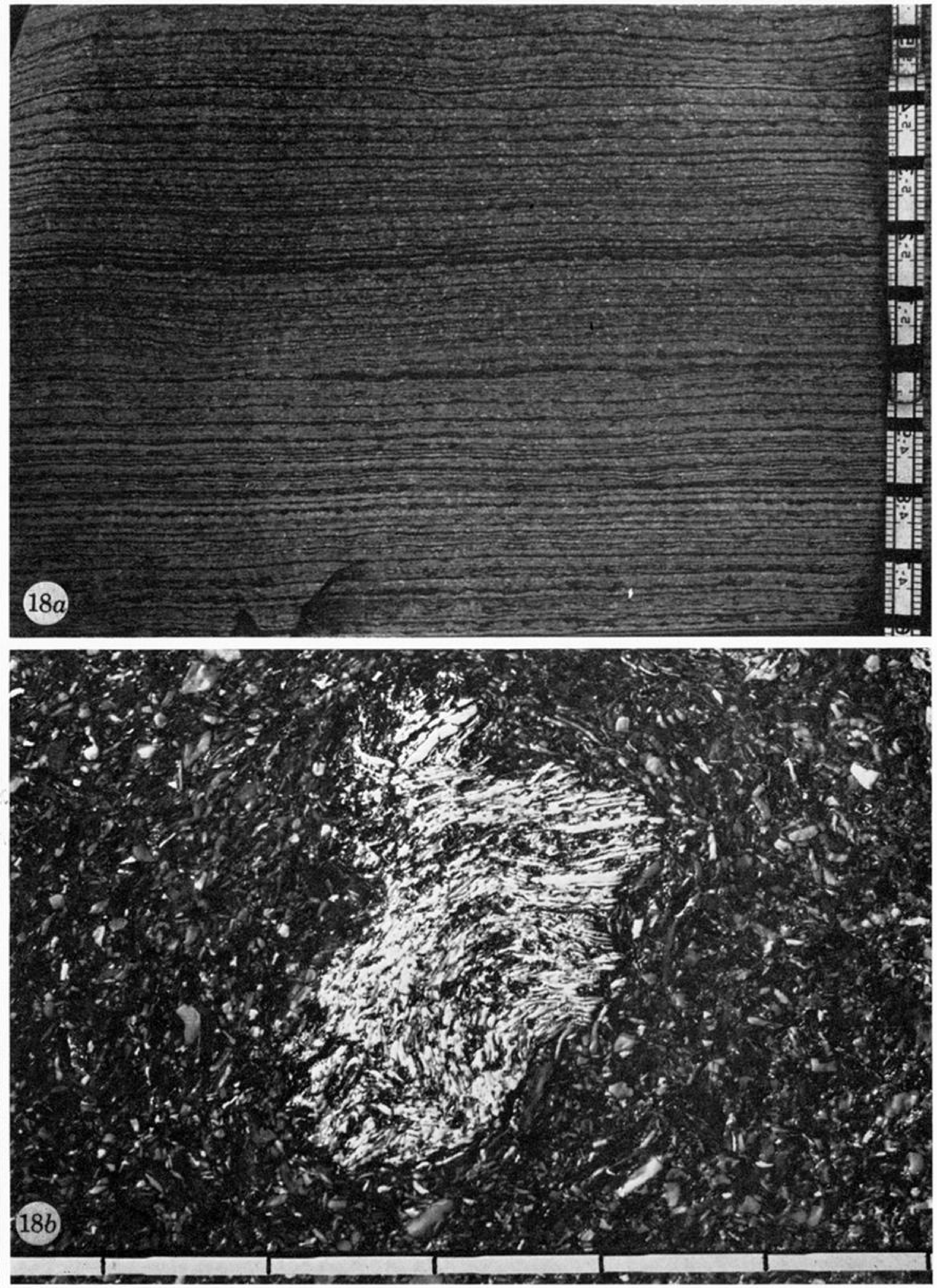
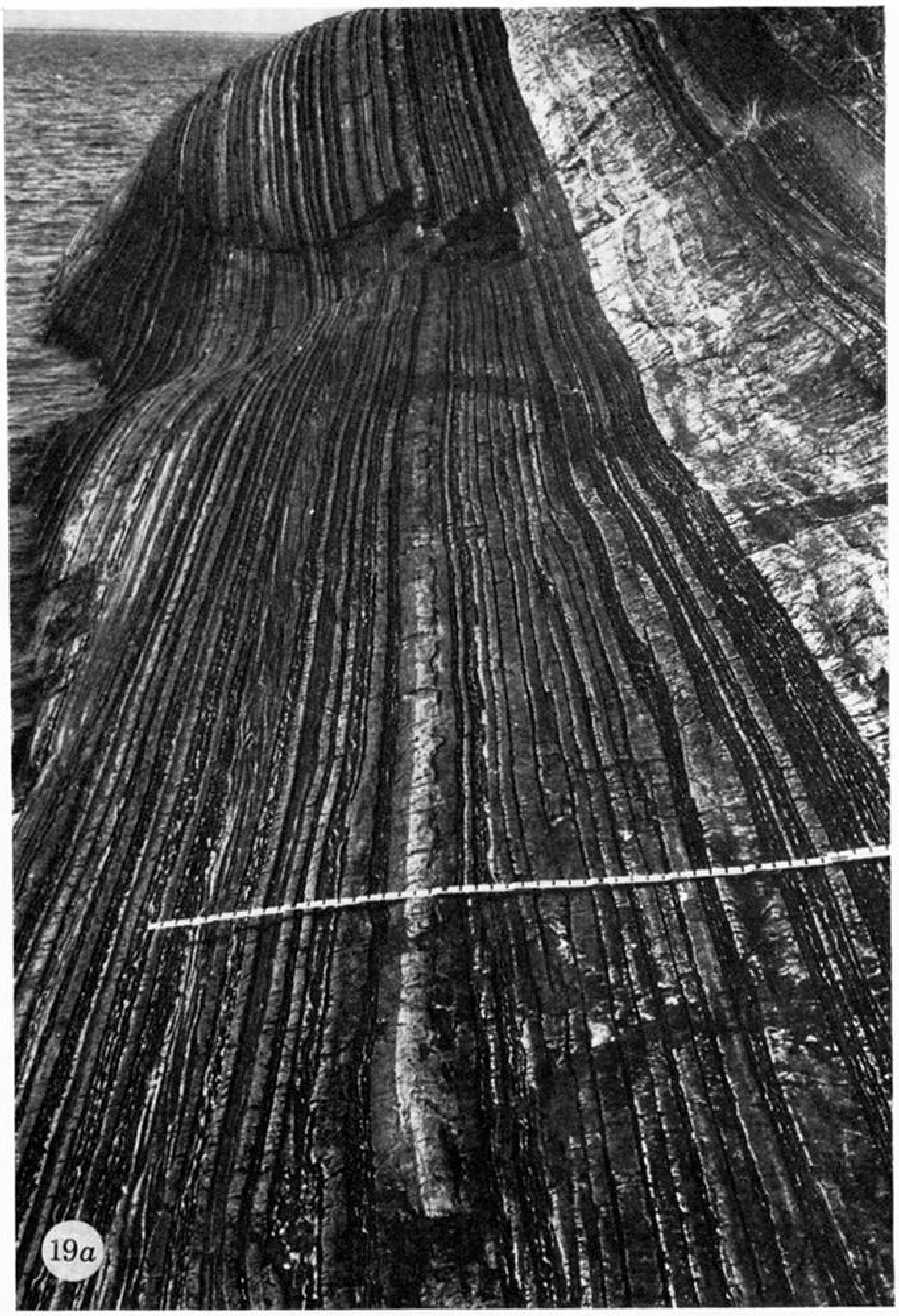


FIGURE 17. (*a*) Bedding surface of columnar stromatolites from the upper part of a 20 m thick bed at the top of the Taltheilei Formation, Utsingi Point. The elongation of the stromatolites, a very common feature, is everywhere perpendicular to the edge of the platform and parallel to paleocurrents determined from ripple marks in associated beds. (*b*) Relatively small elongate columnar stromatolites in the Taltheilei Formation, northeast Blanchet Island. The scale is divided into 3 cm intervals.

FIGURE 18. (*a*) Grey laminated shaly limestone in the Pekanatui Point Formation, south Blanchet Island. Closer to the edge of the platform, the dark shaly laminations become more widely spaced. The scale is divided into 3 cm intervals. (*b*) Bedding surface of debris-flow breccia bed in the Pekanatui Point Formation, Fairchild Point. The breccia consists of twisted blocks of thin-bedded limestone and is interstratified concordantly with non-brecciated beds of the same facies. The scale is divided into 15 cm intervals.



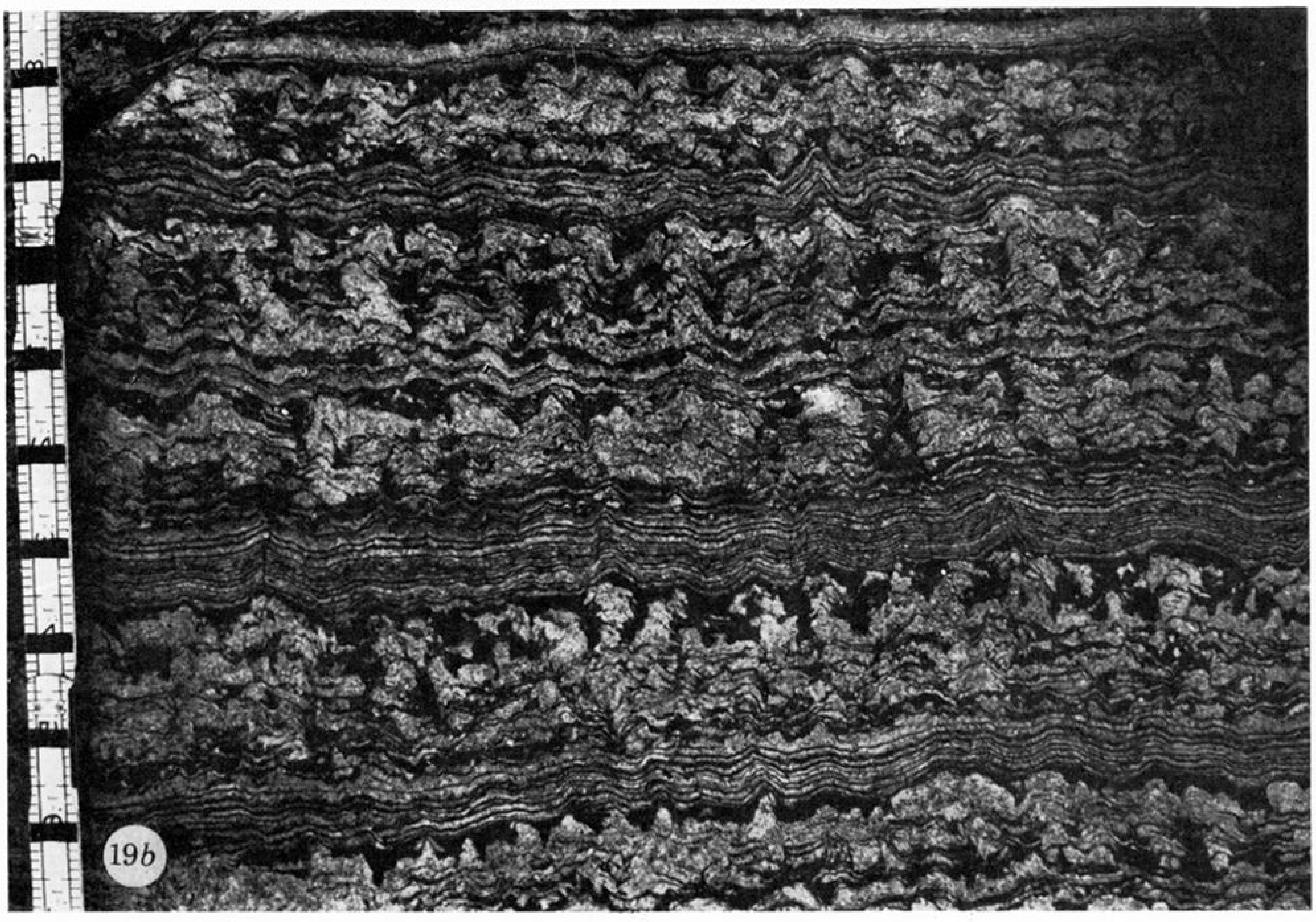
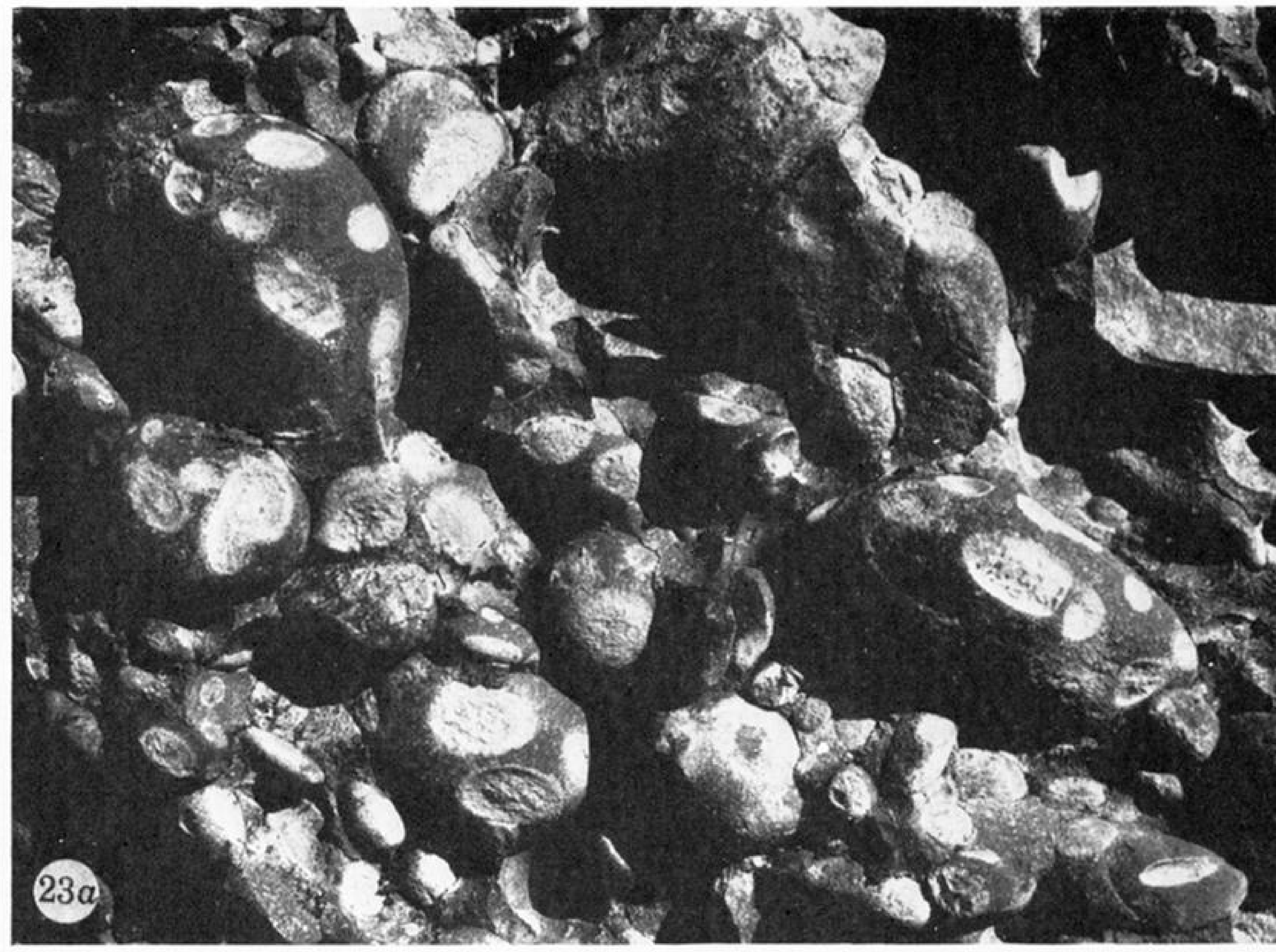
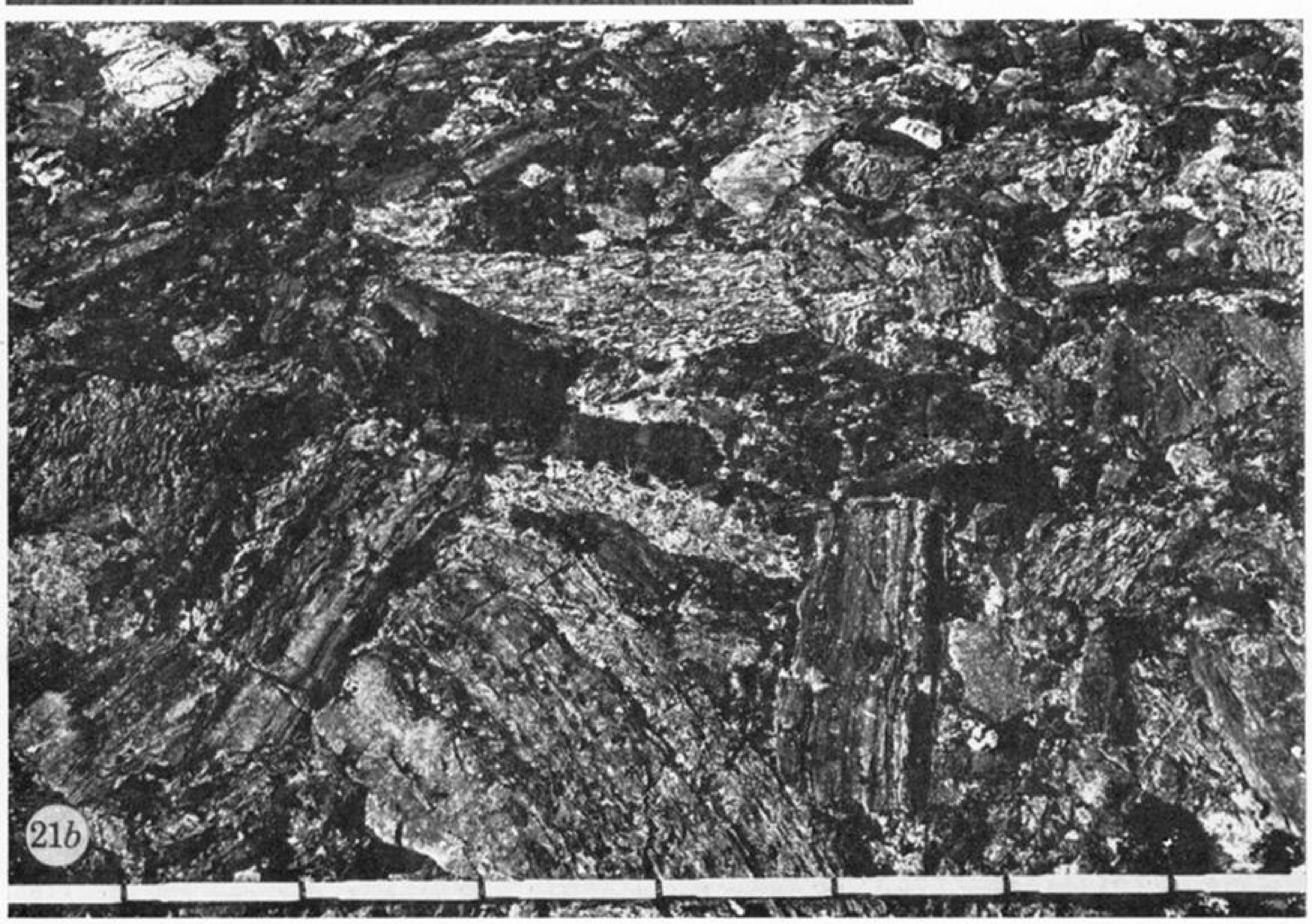


FIGURE 19. (*a*) Thin-bedded greywacke turbidites interstratified with laminated shaly limestone in the Blanchet Formation, south Blanchet Island. The scale is divided into 3 cm intervals. (*b*) Dark green mudstone with digitate calcareous growth structures in the McLean Formation, south Blanchet Island. This facies is repeatedly interstratified with greywacke turbidites and grades northward into digitate loferites of the Utsingi Formation on the platform. The scale is divided into 3 cm intervals.





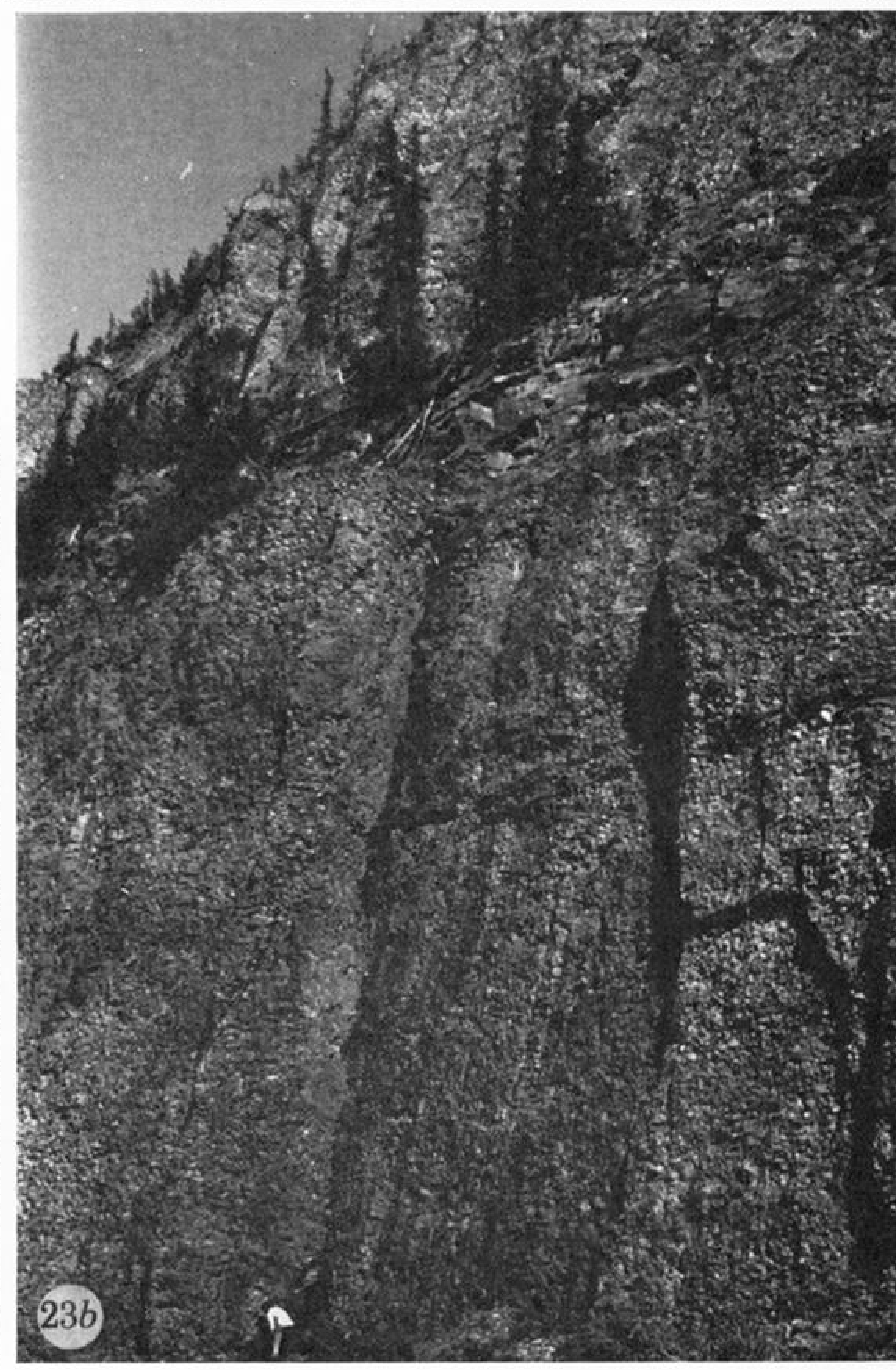


FIGURE 21. (*a*) Bedding surface of red rippled lithic sandstone in the Tochatwi Formation, east Stark Lake. (*b*) Breccia containing closely spaced angular blocks of rippled and stromatolitic carbonate in a matrix of red mudstone. Commonly, the blocks are more dispersed and of much more variable size than in this outcrop. The scale is divided into 15 cm intervals.

FIGURE 23. (*a*) Fanglomerate in the Murky Formation composed of quartzite boulders from the Kluziai Formation. The largest boulders in the photograph have diameters of 50 cm and their superficial red pigment has been leached around their points of contact, resulting in their spotted appearance. (*b*) Two thick fanglomerate beds separated by a thinner sandstone at the base of the trees. Note the lack of stratification in the fanglomerate. A geologist is at the base of the cliff for scale.