CTG'94 field trip, SW end of the Clare River Structure, near Tweed, Ontario

Pierre-Yves F. Robin, Department of Geology, University of Toronto

The schematic geological map on which the field trip stops have been (badly) marked, is from the M.Sc. thesis of W. A. Barclay (1989), ***Deformation fabrics along part of the southeastern limb of the Clare River synform, Grenville Province: evidence for a late episode of penetrative strain and synkinematic intrusion.*** The text on the Mazinaw terrane which follows the stop descriptions is part of the chapter on the Grenville Province, by R.M. Easton, in *Geology of Ontario* (edited by P.C. Thurston, H.R. Williams, R.H. Sutcliffe & G.M. Stott), Ontario Geological Survey Special Volume 4, 1992.

Part I: Development of tectonic layering in Bogart formation metasediments. Relationship of minor folds with the Clare River Structure.

1. Hill, in field south of the former post-office and general store of the Town of Bogart.

These micaceous and amphibolitic metasediments are metaturbidites/ metagreywackes, and define Chappell's **Bogart formation**. Graded bedding, possible flame structure,... Can stratigraphic tops be determined?

Excellent folds. Note that the orientations of foliations and axial planes of minor folds are not parallel to the axial plane of the Clare River Structure. Locally, on the east flank of the hill, the axial plane foliation is very intense and makes it difficult to discern the sedimentary bedding.

2. As the crow flies, only about 300 m to the SSE of Stop 1. But we have to take the cars, go south, park near a gate, and walk into the field on the west side of the road.

Very spectacular development of axial plane foliation in metasediments of the Bogart formation. Note how the development of this foliation is restricted to some laminae. Compare again the attitude of the axial plane foliation with that of the Clare River structure. Are these Z-folds or S-folds? Can we get a fold axis?

3. Continue south, turn left at the T-junction and stop to see the "Dead Cow" outcrop. There is no dead cow any more, and it is in fact not one but a series of low outcrops, some 200 or 300 m north of the road. Also, along the way, when you cross the fence from the road, look at the outcrop on which you step. Same Bogart formation, same kind of folds. But the small calcite-filled segregations, at an angle of 40 to 60° to the bedding, have acted as little faults; note that the actual displacements along these faults may be much smaller that the horizontal strike separation. What is the sense of separation? How do the orientations of bedding and of foliation relate to the regional structure?

4. Turn around, go back to Bogart and park at Crow corner. Go into field to the NE of Crow corner, to see the spectacular "Currie's fold", cleaned many years ago under the direction of Professor J. B. Currie.

Note the beautiful shape of fold, and the excellent axial plane foliation, marked by a good segregation. The rock looks like it still belongs to the graded Bogart formation, but note that some of the layers are calcite marble.

The segregated axial plane foliation is even more spectacular in 2 or 3 small, low outcrops some 100 m to the east in the same field.

5. South of, and across the road from the Laberge property: bedding and axial plane foliation. Start looking at the low oucrop on the side of the road (avoid parking on them). Which is bedding, which is foliation? Cross the fence, and look at many small outcrops in the field.

6. Continue on the road to the northeast, and try and visit the outcrops south of Trudeau's farm **(if Mr. Trudeau will let us, or if he is not in)**: recognise both bedding and foliation in quartz-biotite-staurolite-silimanite/andalusite-garnet metasediments. What metamorphic grade is indicated here? If we are not allowed in the field, examine outcrop just at the edge of the road, under the post.

Part II: granite dikes, including the Mitten dike, and the problem of intrusion of water-saturated eutectic granites during deformation.

7. Turn around, go back toward the SW, almost 1 km SW of Dike corner, and visit granite dikes in metasediment: 1-m wide (north of the road) to 20-m wide (south of the road) dikes intrusive into metasediments, with 2- to-5-cm wide pegmatite veins within them. Note angle with layering in metasediments. Discuss how the small pegmatite veins, or dikelets, within the 20-m wide dike may be related to the granite itself; second boiling? Imagine what might be going on within a crystallising water-saturated granite. Note the orientation of the pegmatite veins subparallel to the dike itself. Examine the strained contacts of the big dike with the host metasediment.

8. Move all the way to the road along the so-called 'Marble Septum', and park ~ 150 m east of the "Goat farm" (but the goats are apparently gone!), for a traverse across the Mitten dyke. Great place for lunch on top.

The **Mitten dyke** is probably a stem of the **Addington Granite** to the North (it certainly looks like it), and forms the ridge flanking the north side of the valley. We walk up the hill to cross it. Note the same small (i.e. 2 to 10 cm) to pegmatite 'dikelets-in-dike' observed in Stop 11. They are slightly oblique to the dike here. The similarity argues for the Mitten dike and the dikes of Stop 11 to be closely related. Note intrusive relation of the Mitten dike into the metasediments along its northen contact. Note the offshoots of smaller (20 cm to 1 m) pegmatitic dikes, somewhat pegmatitic, into the host; note their angular relationship with the layering in the host, and their deformation and boudinage.

Part III (only if we have time; do not count on it): the 'Marble Septum'; granite dikes, and marble tectonics at Dike Corner

9. Drive to the NE and park in the triangular zone at the road junction north of Rice Bridge. Walk across the bridge to examine the Mellon Lake granite to the south.

The **Mellon Lake Granite** is a biotite diorite, ascribed by Lumbers (1967) to an older granite generation than the Addington gneiss. No zircon dates are available. Note how different its mineralogy and its texture on the top, weathered surface, are from those of the Addington/Mitten granite. General orientation of the fabric is the same. Yet the difference becomes less obvious when one visits outcrops toward the SW. Could we be looking at different levels of exposure of the same granite complex, juxtaposed by a fault with a scissor motion?

10. Walk back across Rice bridge, and turn right to visit the treed little hill toward the NE in the **Marble Septum**. This hill exposes the mixture of deformed metasediments, deformed marble, and boudinaged and deformed dikes of pegmatitic granites which is typical of many outcrops within that valley. Note well-developed skarn reaction rims between granite and marble.

The "marble septum" was used to argue in favor of ***granitization***, in a classic argument by Ambrose and Burns (1956). That argument was not very strong even at the time, since it relied on these authors not having noticed the intrusive relationships we have observed in this and previous stops. Also, thin septa between intrusive igneous units are now considered quite mechanically possible: the magmas on either side of such a septum need simply intrude at somewhat different times. Describing it as a septum is in fact questionable: its northern contact with the Mitten dike certainly appears to be an intrusive (although deformed) contact; but its southern contact may be tectonic, rather than intrusive.

11. Drive to Dike Corner and park there. **NO HAMMERS**

Observations will be made on flat outcrops west of the road junction, on the ridge immediately SE of the junction, and the low ridge N of the E-W road.

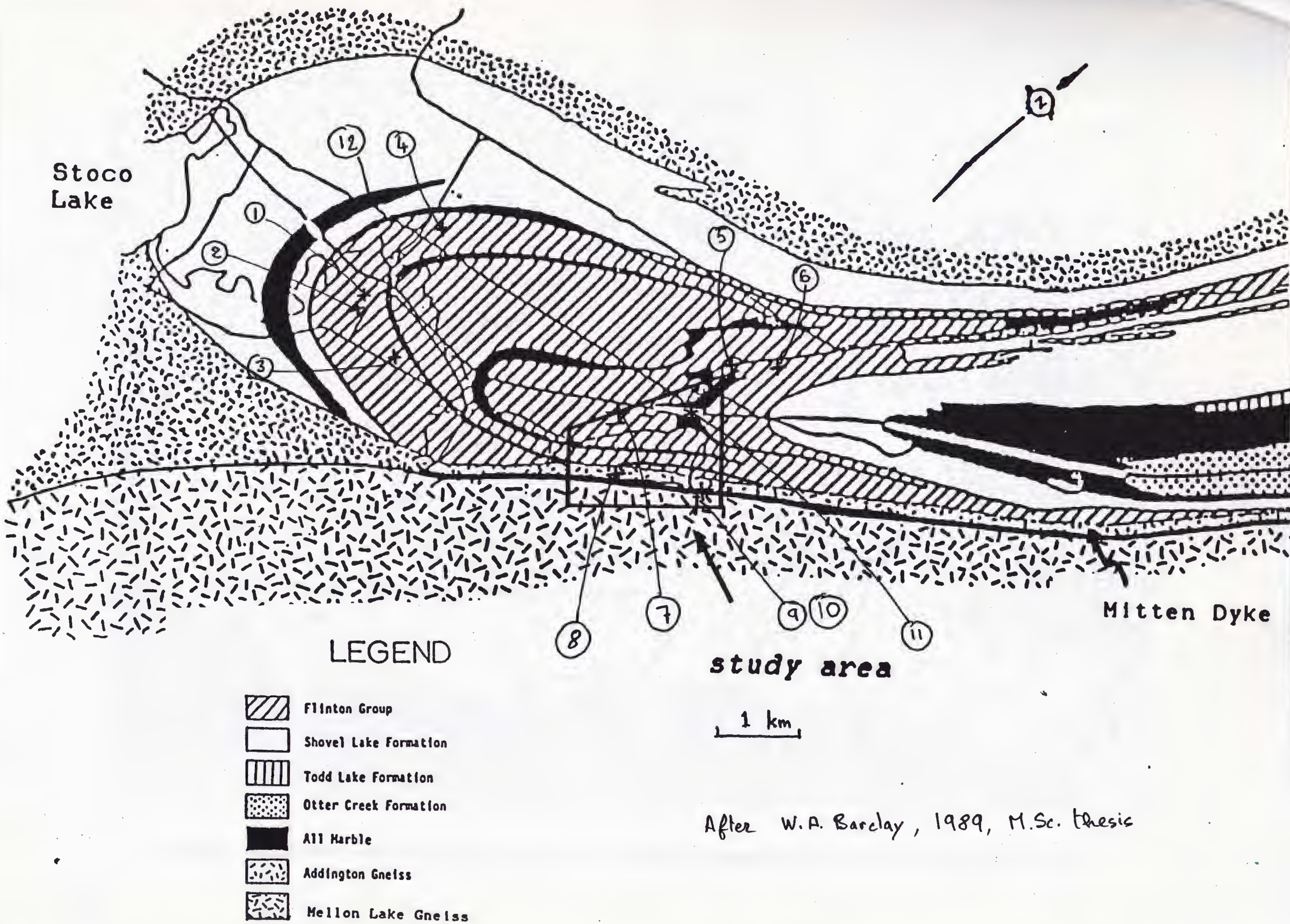
(a) Start in the SE end of the area and examine a rusty schist, a ~ 10-m wide granite dike, and the contact between the two. Note angular relationship between contact and layering in the schist, and the similarities of both orientations to those observed in Stop 11 (which we can see in the distance).

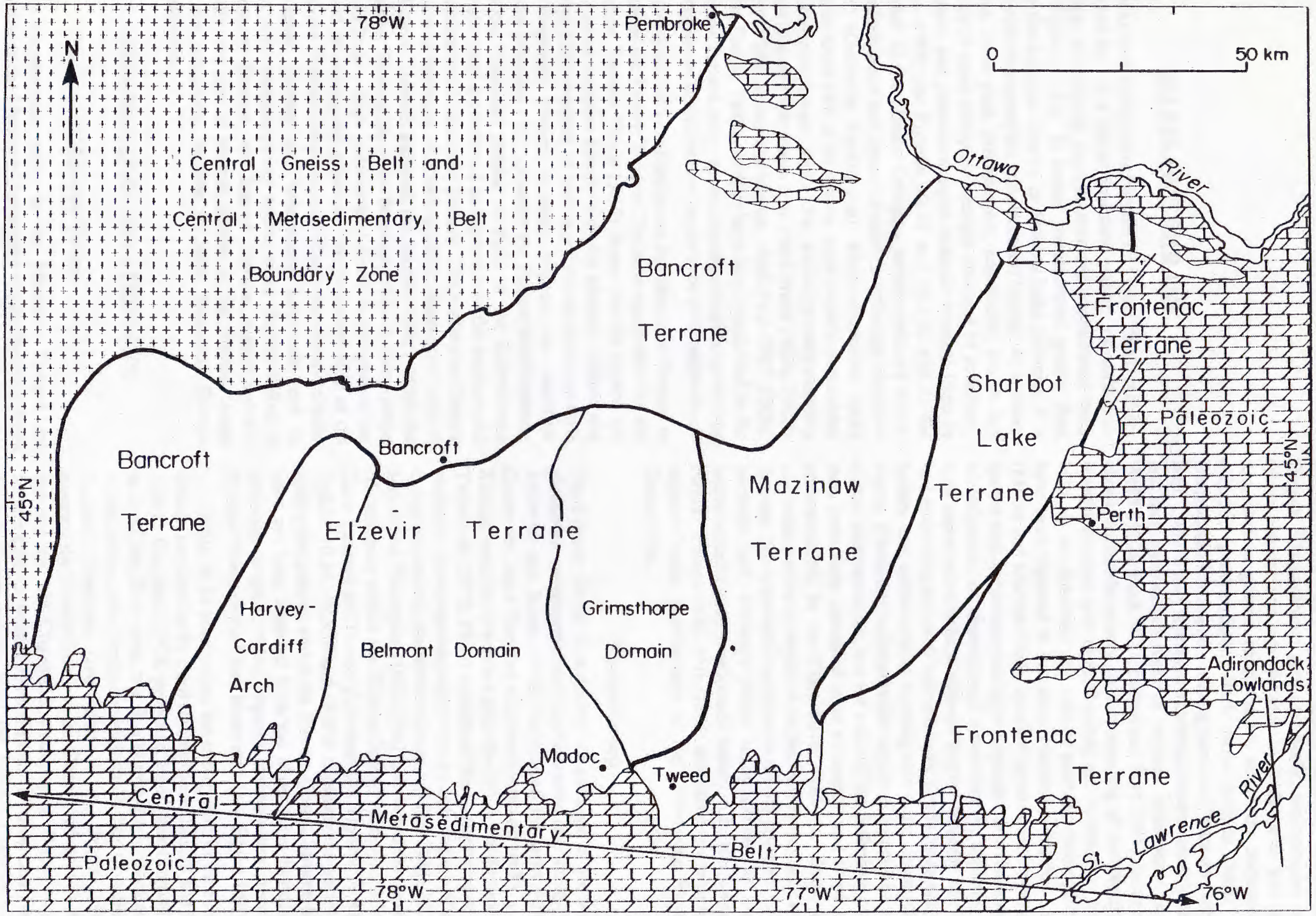
(b) Next examine the impure marble with schistous laminae and pegmatite veins/dikelets which makes up most of this outcrop. Both laminae and veins are very very well exposed by differential weathering. They are folded and/or boudinaged, depending on their orientation relative to that of the strain, and on the timing of their emplacement. The apparent strike of bedding in the marble is variable, because of folds on various scales. As one walks toward the NE, however, toward the high ridge behind the big barn, on the other side of the intersection, the minor folds decrease in amplitude and come to conform with the major fold geomorphologically expressed by the ridge. The minor folds have variable plunges.

(c) SE of the junction, granite dykes cut the marble and are boudinaged within it. Calc-silicate minerals at the contact between granite and marble are rare, suggesting perhaps that skarns have been sheared off during the deformation.

Part IV:

12. The outcrop which summarises the geology of the area, near the road cut in calcite-dolomite-diopside-tremolite marble hill, along main road back to Tweed. (Paul F. Williams)


Stoco Lake
Mitten Dyke
study area
1 km
LEGEND
Flinton Group
Shovel Lake Formation
Todd Lake Formation
Otter Creek Formation
All Marble
Addington Gneiss
Mellon Lake Gneiss
After W.A. Barclay, 1989, M.Sc. thesis



a

Figure 19.51a. Lithotectonic terranes and domains of the Central Metasedimentary Belt in Ontario as used in this report.

MAZINAW TERRANE

The Mazinaw Terrane (*see* Figures 19.51 and 19.82) is characterized by: 1) a linear structural pattern of northeast-trending, alternating, metaplutonic and supracrustal belts (*see* Figure 19.82); 2) higher metamorphic grades than either the adjacent Elzevir or Sharbot Lake terranes; 3) a polycyclic metamorphic and structural history; 4) a slower cooling from peak metamorphic conditions, which has resulted in more complete isotopic resetting of rubidium-strontium and potassium-argon isotopic systems (data *in* Cosca 1989; *see* Figures 19.16 and 19.17); and 5) the presence of the Flinton Group, a metamorphosed supracrustal sequence that unconformably overlies the Mazinaw Group (Grenville Supergroup). Metasedimentary rocks underlie about 25% of the terrane. metavolcanic rocks about 20% and metaplutonic rocks 55%. Although many mapping studies (Smith 1958; Smith et al. 1969; Wolff 1982a, 1982b; Bright 1986a; Moore and Morton 1986; Pauk 1987, 1989a, 1989b; Easton 1988a, 1990e) have been conducted in the Mazinaw Terrane, only recently has the complex history and distinct nature of this terrane begun to be fully understood.

Previous workers considered the Mazinaw Terrane to be part of the Elzevir Terrane. The time-space chart for the Mazinaw Terrane (Table 19.17) shows that the terrane has had a more complex history than the adjacent Elzevir and Sharbot Lake terranes. as well as a different structural style and magmatic history. In some respects, the Mazinaw Terrane provides a link between the Elzevir and Frontenac terranes. Tonalites, similar to those of the Grimsthorpe Domain, may have served as basement to the supracrustal succession in the Mazinaw Terrane. The Addington Granite appears to represent Methuen alaskitic magmatism seen in the Elzevir Terrane, although the granitic bodies within the Mazinaw Terrane are more heterogeneous in composition and include a high sedimentary xenolith component not typical of anorogenic granites. The structural style of the Mazinaw Terrane, however, is similar to that observed in the Frontenac Terrane and Adirondack Lowlands, and its polycyclic metamorphism seems more closely linked to Frontenac Terrane, not Elzevir Terrane, metamorphism. Finally, the age of detrital zircons in the Flinton Group ties the Mazinaw Terrane in with Frontenac Terrane magmatism.

STRATIGRAPHIC RELATIONSHIPS

Mazinaw Group

Most supracrustal rocks in the Mazinaw Terrane are assigned to the Mazinaw Group (new term). Two main belts of supracrustal rocks are present (*see* Figure 19.82), one located in the Kaladar area, including the Clare River synform, the other stretching from Bishop Corners northeast to Clyde Forks through Fernleigh, Ardoch and Ompah. Metavolcanic rocks of mafic, intermediate and felsic composition are present in both belts and are described in greater detail below, under the section on volcanic geochemistry. Both belts include mafic to intermediate hornblende-bearing gneisses ("para-amphibolites"), which may represent interbedded metavolcanic tuffs and metasedimentary rocks of volcanic provenance (e.g., Wolff 1982a; Pauk 1987; Easton and Ford 1991). Siliceous clastic metasedimentary rocks consisting of metawackes, metalitharenites and metasiltstones are common in both belts, as are dolomitic and calcitic marbles. Stromatolites have been reported in dolomitic marbles at Marble Lake (Moore and Morton 1986). Primary textures and structures are not well-preserved making stratigraphic correlation and identification of depositional environments difficult. In both supracrustal belts, it appears that the volcanic edifices had associated aprons of volcaniclastic debris (para-amphibolites and metawackes) interfingered with clastic sediments. Carbonate sedimentation occurred coincident with clastic sedimentation, but the bulk of the carbonate rocks may have been deposited after volcanism and associated sedimentation. In the Clare River synform, clastic and carbonate metasedimentary rocks become more abundant to the east. No systematic variation is readily apparent in the northern belt, although major accumulations of metavolcanic rocks seem to be located mainly in the Harlowe and Flower Station areas, with clastic and carbonate metasedimentary rocks predominant in the intervening Plevna–Ompah area.

Flinton Group

The Flinton Group is a distinctive package of metaconglomerate, metamorphosed quartz arenite and metapelite. Ambrose and Burns (1956) introduced the term Flinton Group for some (but not all) conglomeratic rocks in the Kaladar area, but it was not until the 1970s that Moore and Thompson (1972. 1980) formalized the stratigraphy of this sequence of metasedimentary rocks by defining the Flinton Group and several distinctive formations (*see* Table 19.14). Moore and Thompson (1972, 1980) also noted the presence of carbonate within the Flinton Group, namely, the Myer Cave, Lessard and Fernleigh formations. Figures 19.83a to 19.83f, 19.84, 19.85a and 19.85b illustrate the variety of rock types present within the Flinton Group. Stromatolites have not been reported from the Myer Cave Formation, although Bright (1986a) reports stromatolites from the Flinton Group in the Clare River area (*see* Figure 85b).

Table 19.14 summarizes the stratigraphy of the Flinton Group, as modified by Ford (in prep.) and Easton and Ford (1991). Chappell (1978) proposed 2 additional formations in the Clare River area, the Beatty and Bogart formations which are roughly correlative with the Bishop Corners and Lessard formations (*see* Table 19.14). Bright (1986a) mapped the Flinton Group in the Clare River area in detail but did not use Chappell's stratigraphy. Hounslow and Moore (1967) present several major element analyses of Flinton Group schists from the Marble Lake–Fernleigh Belt, however, they are not tied to stratigraphy and appear to

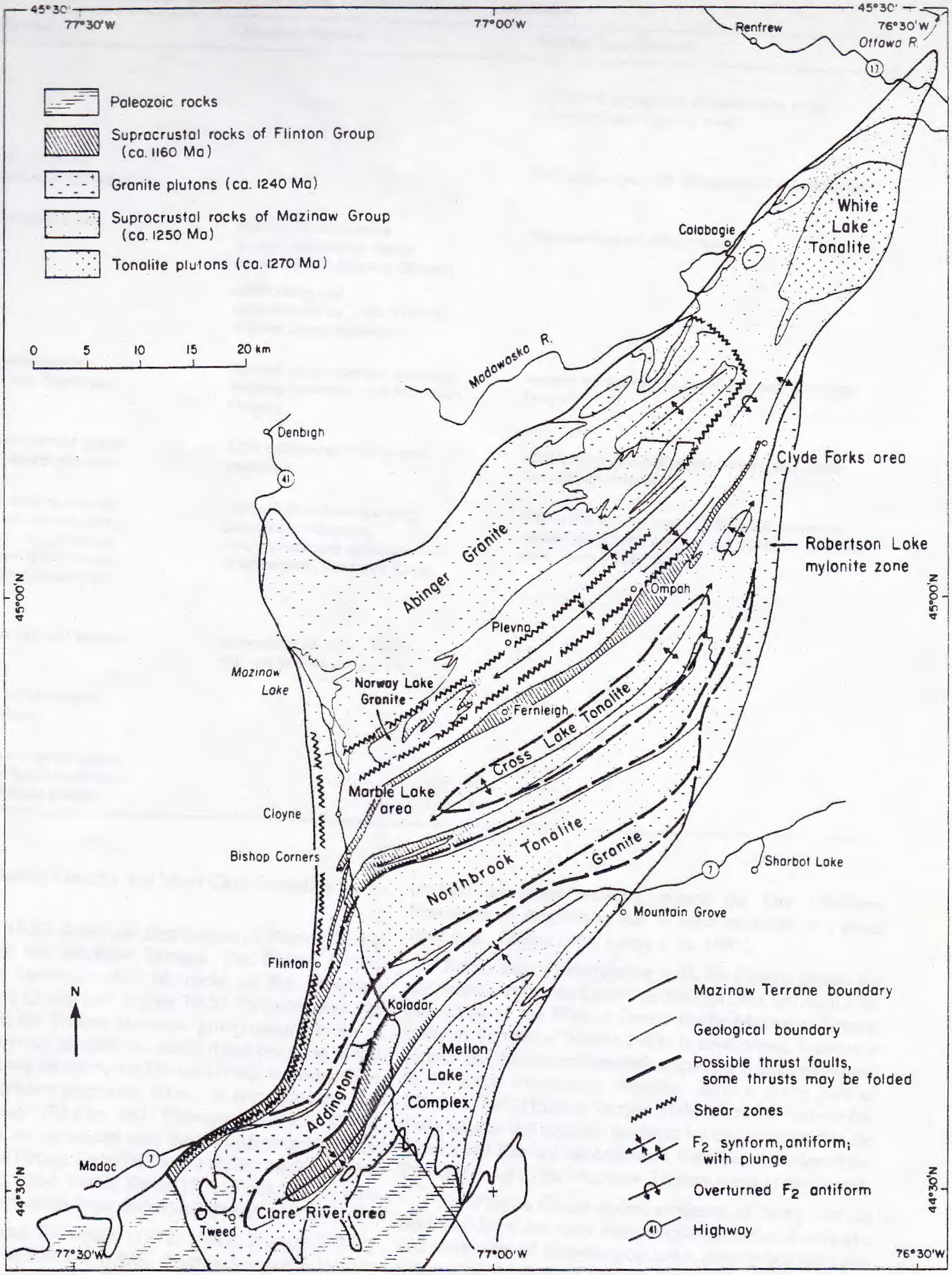


Figure 19.82. Sketch map showing structural trends within the Mazinaw Terrane and adjacent terranes.

Table 19.17. Comparative geologic histories of the Elzevir, Mazinaw and Sharbot Lake terranes

Elzevir Terrane	Mazinaw Terrane	Sharbot Lake Terrane
		<1050 million-year-old mylonitization along Robertson Lake mylonite zone
1090 million-year-old Skootamatta Suite intrusions		1090 million-year-old Skootamatta Suite intrusions?
Ottawan orogenic effects minimal?	regional metamorphism to upper amphibolite facies associated with Ottawan Orogeny	Ottawan orogenic effect minimal
	uplift, rifting and sedimentation ca. 1160–1150 Ma (Flinton Group deposition)	
regional metamorphism coincident with Elzevirian Orogeny	regional metamorphism, thrusting, shearing associated with Elzevirian Orogeny	regional metamorphism coincident with Elzevirian Orogeny
1250 million-year-old granite and gabbro-diorite plutonism	1250 million-year-old? granite plutonism	1250 million-year-old? gabbro-diorite and granite-granodiorite plutonism
1260–1250 million-year-old bimodal tholeiitic volcanism, volcaniclastic and carbonate sedimentation (Hermon and Mayo groups, Grimsthorpe Group)	1260–1250 million-year-old? calc-alkalic volcanism, volcaniclastic and carbonate sedimentation, (Mazinaw Group)	1260–1250 million-year-old? tholeiitic volcanism, volcaniclastic sedimentation and carbonate sedimentation (Sharbot Lake Group)
1270 million-year-old tonalite plutonism	basement of ca. 1270 million-year-old tonalite and gabbro	? ?
1290–1270 million-year-old gabbro plutonism		
>1290 million-year-old gabbro and tholeiitic basalt magmatism (Canniff and Basal groups)		

represent Bishop Corners, and Myer Cave formation metapelites.

Figure 19.82 shows the distribution of Flinton Group strata within the Mazinaw Terrane. The Flinton Group locally lies unconformably on rocks of the Mazinaw ("Grenville") Group (*see* Figure 19.82 for localities). As noted above, the Flinton-Mazinaw group unconformity is locally preserved. In addition, mafic dikes common in the Mazinaw Group do not cut the Flinton Group, and other than a few late syenite pegmatite dikes, no intrusions cut the Flinton Group (Moore and Thompson 1980). These observations are consistent with the unconformable nature of the Flinton Group. Detailed studies of the conglomerates (Walton et al. 1964; van de Kamp 1971; Psutka 1976) also suggested derivation from underlying plutonic rocks.

Moore and Thompson (1972, 1980), Harnois (1987) and Harnois and Moore (1988) describe a regolith locally preserved beneath the Flinton Group. The regolith is preserved as a staurolite-garnet schist, termed the Ore Chimney Formation (Moore and Thompson 1980), which is part of the Mazinaw Group. The extent of the regolith is unclear, as some workers regard the Ore Chimney Formation as originating due to metasomatism in a shear zone (e.g., Dillon 1985; Laing et al. 1987).

Rocks directly correlative with the Flinton Group are not present within the Elzevir or Sharbot Lake terranes. This restriction of the Flinton Group to the Mazinaw Terrane (Elzevir Terrane *of* Moore 1982) is anomalous, because it means that this unconformable sequence is only preserved in the most structurally complex, highest grade part of Moore's (1982) Elzevir Terrane. If the Mazinaw Terrane has had a unique and separate geologic history compared to the Elzevir and Sharbot Lake terranes, then the restriction of the Flinton Group to the Mazinaw Terrane is easily explained.

The Flinton Group shows evidence of being derived from both local and more distal sources. South of Sand Lake and southwest of Shawenegog Lake, granite boulders and feldspathic litharenites rim the Norway Lake Granite, the probable source of the boulders (Smith 1958; Smith et al. 1969; Easton and Ford 1991). Easton and Ford (1991) found that basal conglomerates were more extensive than previously recognized, and contain interbedded pelitic and

Figure 19.83a. Quartz arenite (left) and quartz arenite cobble conglomerate (right) of the Bishop Corners Formation, Flinton Group, Highway 41 north of Northbrook. Hammer handle is 30 cm long.

Figure 19.83b. Stretched cobble conglomerate within the Lessard Formation, Highway 41, north of Kaladar. Hammer handle is 30 cm long.

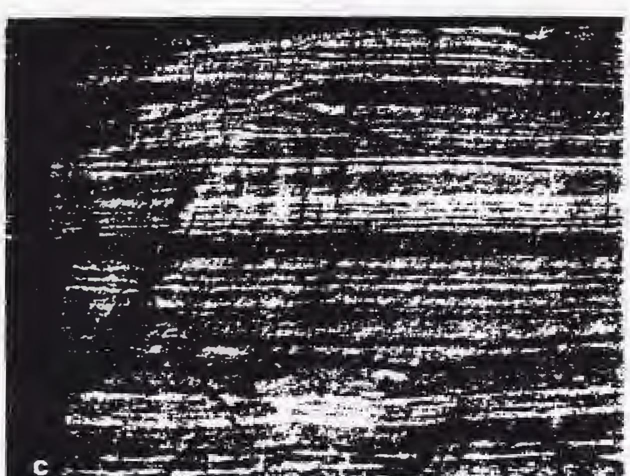


Figure 19.83c. Thinly laminated sulphide-bearing black schist of the lower Myer Cave Formation, Flinton Group, Highway 506 east of Marble Lake. Field of view is 20 cm across.

Figure 19.83d. Poorly laminated sulphide-bearing black schist of the upper Myer Cave Formation, Flinton Group, Highway 506 west of Fernleigh. Hammer handle is 30 cm long.

Figure 19.83e. Carbonate breccia unit with black shale matrix, Myer Cave Formation, Flinton Group, Highway 506 east of Marble Lake. Hammer handle is 30 cm long.

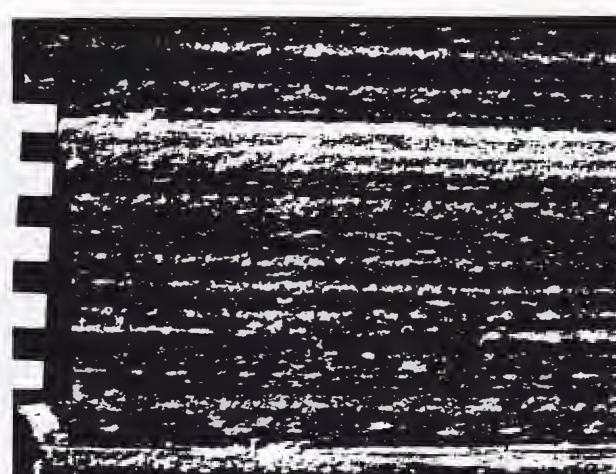


Figure 19.83f. Thinly bedded calcareous silts and shales of the Fernleigh Formation, Flinton Group, with microcline porphyroblasts. Scale bar is 9 cm long.

Figure 19.84. Kyanite porphyroblasts within biotite-kyanite-staurolite schists of the Bishop Corners Formation near Fernleigh. Hammer head is 17 cm across.

Figure 19.85a. Thin- to medium-bedded calcitic marbles of the Myer Cave Formation, Flinton Group, Highway 506 east of Fernleigh. Hammer handle is 30 cm long.

semi-pelitic schists, including distinctive biotite-kyanite schists. The conglomerates are in unconformable contact with the granites, marbles and amphibolites in the area, thus, the conglomerates are correlated with the Flinton Group. Significantly, these Flinton Group occurrences are not in the typical synclinal setting that Moore and Thompson (1972, 1980) document for the Flinton Group, and the Flinton Group thus may not be as infolded with Mazinaw Group strata as previously believed (cf., Rivers 1976; Chappell 1978). Moore and Thompson (1980) also note conglomerates containing tonalite and granite clasts adjacent to the Northbrook Tonalite and Addington Granite, near Flinton and Kaladar respectively, are not associated with major synclines. Moore and Thompson (1972, 1980) also describe major facies changes in Flinton Group rocks suggesting significant local source controls. Kinsman and Parrish (1990) suggest a more distal source region, as 1157 million-year-old zircons in Bishop Corners Formation metamorphosed quartz arenites are similar in age to Frontenac Terrane plutons. Frontenac Terrane metamorphosed quartz arenites may also be the source of quartz arenite cobbles within the Flinton Group. Alternatively, the Abinger Granite (Rb-Sr age of 1185±25 Ma, Bell and Blenkinsop 1980) and the Norway Lake Granite, which locally shed detritus into the Flinton Group, might be the source of the 1157 million-year-old zircon population reported by Kinsman and Parrish (1990).

Moore and Thompson (1972, 1980) suggested a fluvial environment for the coarse clastic units of the Flinton Group and a shallow-marine setting for the pelitic and carbonate units, both being deposited in an extensional environment. Easton and Ford (1991) postulate a fluvial-lacustrine environment for the Flinton Group in the Mazinaw Terrane, with deposition controlled by rift valleys resulting in longitudinally oriented alluvial-fan and braided stream trunk systems (Figure 19.86). This model explains the preservation of the Flinton Group in synformal, linear belts; localized faulting present along the Flinton unconformity; and localized mineralization. Alternating pelite-carbonate

Figure 19.85b. Domal stromatolite (right) within dolomitic marbles of the Flinton Group, Clare River area, Highway 41 south of Kaladar. Field of view is 20 cm across.

beds within the Fernleigh Formation may reflect annual variation in sediment supply into rift valley lakes, and the ubiquitous and abundant presence of tourmaline throughout the Flinton Group (e.g., Thompson 1972; Easton 1988a) would be expected in an arid depositional environment. The Myer Cave Formation megabreccia unit probably formed by collapse of a cliff into a lake, allowing for the incorporation of Myer Cave Formation and Mazinaw Group blocks in a pelitic matrix. The depositional setting outlined above is similar to that described by Lefebvre (1989) for the Zambian copper deposits, as are the metal associations (gold, copper, lead, zinc and arsenic). The metallogenic implications of this model are outlined *in* Easton and Fyon (this volume).

GEOCHEMISTRY OF METAVOLCANIC ROCKS

Two chemical suites are represented in the Mazinaw Terrane, a tholeiitic suite and a calc-alkalic suite, studied in the Harlowe–Marble Lake area by Sethuraman (1970), Sethuraman and Moore (1973), Condie and Moore (1977), Ayer (1979), Moore and Morton (1986) and Harnois (1987). Chappell (1978) and Bright (1986a) discusses the chemistry of metavolcanic rocks within the Clare River synform in the southern part of the terrane.

The tholeiitic succession straddles the Grimsthorpe Domain–Mazinaw Terrane boundary, and thus, some of the analyzed tholeiitic rocks are from the Grimsthorpe Domain. The calc-alkalic succession, however, lies entirely within the Mazinaw Terrane. Thus, continuity of the 2 successions is debatable; more likely, the 2 sequences are tectonically juxtaposed (Easton and Ford 1991). Neither metavolcanic sequence has been dated.

Condie and Moore (1977) and Sethuraman and Moore (1973) classify the tholeiitic sequence as low-potassium tholeiites, similar to modern arc and rise tholeiites and Archean low-potassium tholeiites (*see* Figure 19.62). Little chemical variation is observed in the tholeiitic sequence. The tholeiites are nepheline normative and contain chlorite (Sethuraman and Moore 1973). Both features could be secondary (Moore and Morton 1986). Mafic volcanic rocks near Flinton are also tholeiitic (Wolff 1982a), in agreement with the results of Sethuraman and Moore (1973). The rocks near Flinton, however, are most likely part of the Grimsthorpe Domain.

The calc-alkalic sequence consists of andesites, dacites and minor rhyolite and are similar to modern calc-alkalic suites (Condie and Moore 1977; *see* Figure 19.62). Fragmental rocks are common in the intermediate and felsic part of the sequence. Dacite dikes cut the tholeiitic sequence and the Elzevir Tonalite and have been interpreted as feeders to the calc-alkalic volcanic succession (Moore and Morton 1986; Lumbers et al. 1990). Possibly correlative dacite dikes, adjacent to the Elzevir Tonalite, have been dated at 1229^{+11}_{-4} Ma (Connelly et al. 1987).

Ayer (1979), in the southern Mazinaw Lake–Pringle Lake area, reported a tholeiitic basalt-andesite-rhyolite complex, which he considered to be a volcanic centre. Rocks in this area are generally east-trending and are structurally discordant with the north- to northeast-trending calc-alkalic metavolcanic sequence to the south (Moore and Morton 1986). Easton and Ford (1991) remapped much of the area and reported the presence of few recognizable volcanic textures within the sequence. Easton and Ford (1991) described these rocks as laminated, quartzofeldspathic gneisses that may in fact be recrystallized mylonites (*see also* Rivers 1976). Some rocks adjacent to the Norway Lake Granite were reinterpreted by Easton and Ford as flattened conglomerates of the Flinton Group. Thus, the extent of the Mazinaw Lake–Pringle Lake volcanic complex may be less than originally indicated by Ayer (1979) and Moore and Morton (1986).

Bright (1986a) reports basaltic and dacitic-rhyolite metavolcanic rocks in the Clare River synform (Shovel Lake Formation *of* Chappell 1978) and in the Salmon River area. Metavolcanic rocks from both occurrences are bimodal, consisting of tholeiitic basalts and tholeiitic to slightly calc-alkalic dacite and rhyolite.

In the western half of the Darling area, mafic, intermediate and felsic metavolcanic rocks are metamorphosed to upper amphibolite facies and are intensely deformed by the adjacent Robertson Lake mylonite zone and a high-strain zone along the eastern margin of the Bartraw dome (Easton 1988a). Limited geochemistry on this sequence suggests a calc-alkalic affinity, however, abundant carbonate in these rocks makes interpretation problematic (Easton 1988a). These rocks continue south into the Lavant area (Pauk 1989b), but no geochemistry is available from that area.

In summary, the Mazinaw Terrane volcanic sequences are largely calc-alkalic, unlike the dominantly tholeiitic Elzevir Terrane metavolcanic rocks.

PLUTONISM

It has been assumed previously that the plutonic rocks of the Mazinaw Terrane, such as the Northbrook/Cross Lake Tonalite and the Abinger Granite, were equivalent to the Elzevir and Weslemkoon tonalites and the Methuen Suite in terms of age, composition and origin. However, apart from the Addington Granite which has been dated at 1245^{+15}_{-10} Ma (van Breemen and Davidson 1988b), no U-Pb zircon ages are available to confirm this assumption. Skootamatta Suite plutons, which are common in the Elzevir and western Frontenac terranes, are notably absent from the Mazinaw Terrane. Unlike the Elzevirian domains, the Mazinaw Terrane is characterized by elongate, northeast-trending lenticular plutonic patterns, exemplified by the Addington Granite and the Northbrook Tonalite (*see* Figure 19.82).

In the central part of the terrane, a repetitious pattern of elongate tonalite masses separated by zones of layered amphibolite, para-amphibolite and siliceous clastic metasedimentary rocks, exemplified by the Northbrook and Cross Lake tonalites, is also suggestive of a tectonically influenced shape. The tonalite plutons are associated with the metavolcanic rocks of the terrane. The Mazinaw Terrane tonalites differ from Grimsthorpe Domain tonalites in being more granodioritic in composition (e.g., Wolff 1982a; Kamilli 1974), being more deformed and being more migmatitic. Most of these differences could be attributed to the higher grade metamorphic conditions seen in the Mazinaw Terrane. The geochemistry of the tonalites has been studied by Lumbers et al. (1990). Somers (1984) studied the White Lake Tonalite located in the extreme northeast corner of the terrane.

The best evidence for tectonism being responsible for pluton form is found near Coxvale, where the northern and southern lobes of the Cross Lake Tonalite are separated by a high-strain zone developed in both tonalite and supracrustal rocks, and where an increasing strain gradient can be observed in the tonalite as the contact is neared. The form of the Cross Lake Tonalite is also suggestive of a folded thrust contact, rather than an unmodified intrusive contact. Lumbers et al. (1990) report an age of *circa* 1270 Ma from the Cross Lake Tonalite, suggesting that the tonalites may have served as basement to the metavolcanic sequence.

The granitic bodies within the Mazinaw Terrane are characterized by a fine- to medium-grain size and considerable compositional heterogeneity ranging from granodiorite and quartz monzonite to syenogranite. Monzogranites and monzonites predominate. Muscovite is common in all phases of these leucocratic granitic rocks, which are typically weakly foliated to gneissic. Bright (1986a) presents geochemical data on the Mellon Lake Complex. Moore (1967), Wolff (1982a), Carter (1981) and Easton (1988a) present geochemical and modal data for the Addington Granite, and Lumbers et al. (1990) give an average analysis for the Abinger Granite. Lumbers et al. (1990) include both the Addington and Abinger granites in the alaskite (Methuen) suite, even though they are considerably more heterogeneous and include a greater proportion of metasedimentary xenoliths than Methuen Suite granites in the Elzevir Terrane.

In the northern part of the terrane, the Abinger and Norway Lake granites have tectonized northern margins and are separated by narrow belts of supracrustal rocks. The shape of the granitic plutons, in part, is largely a reflection of tectonism. This is also consistent with gravity models across the Mazinaw Terrane (Real and Thomas 1987), which indicate that the plutons are extremely thin (approximately 1 km thick).

The granitic plutons are associated with the dominantly metasedimentary belts within the terrane. The granitic bodies commonly contain inclusions and inclusion trains of country rocks. Given the lithologic variation between phases of the granite separated by inclusion trains, the granite bodies may have been emplaced as a series of sills. Wolff (1982a) reached the same conclusion regarding the emplacement of the Addington Granite. The Abinger Granite intrudes tonalite intrusion breccias along Highway 41, indicating that this body is younger than the tonalites. The granite is more deformed than the intrusion breccia it intrudes, perhaps due to syndeformational emplacement.

The Mellon Lake Complex (Bright 1986a; Sheffield intrusion *of* Wolff 1982a) has also intruded an older, migmatitic, nebulitic tonalite, and both Bright (1986a) and Wolff (1982a) have suggested that it may, in part, have served as basement to the supracrustal sequence in the Clare River area. The Bartraw dome (Karboski 1980; Easton 1988a) in the northeast of the Mazinaw Terrane is cored by migmatized tonalite and rimmed by a medium-grained, alaskitic granite. On structural grounds, Karboski (1980) suggested that the Bartraw dome may have also served as a basement complex.

The Elbow Lake meta-anorthosite (Lumbers 1982; Lumbers et al. 1990) on the southeast shore of Norcan Lake bears an unknown relationship to the anorthosite suite rocks in the Central Metasedimentary Belt. Anorthosite slivers also occur south of this body in the tectonized margin of the Bartraw dome (Easton 1988a).

Rb-Sr whole-rock ages on the Abinger and the Mellon Lake bodies are 1185±25 and 1120±90 Ma respectively (Bell and Blenkinsop 1980), although the preliminary age on the Mellon Lake Complex was 1250±130 Ma (Bell and Blenkinsop 1979). These ages could have been reset during regional metamorphism, as the Rb-Sr age on the Addington Granite 1035±42 (Krogh and Hurley 1968) is some 200 million years younger than the U-Pb zircon age on the

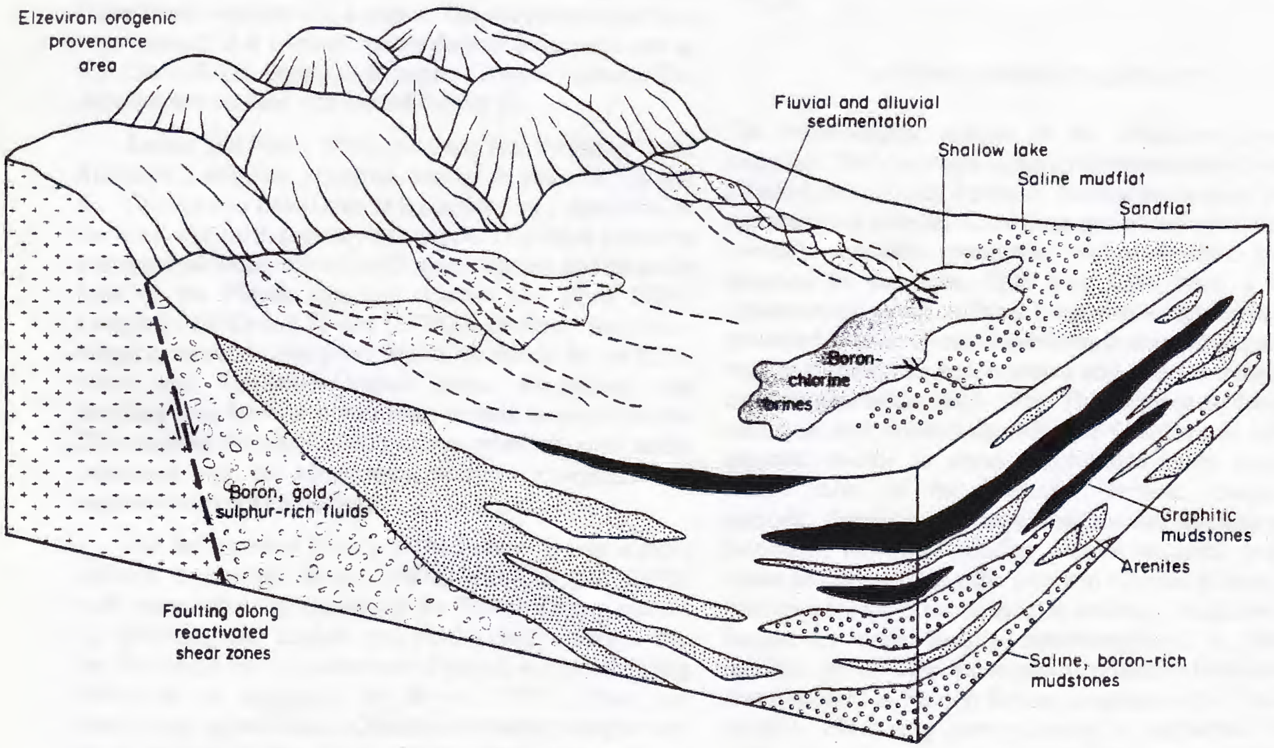


Figure 19.86. Sketch showing a possible depositional model for the Flinton Group.

same body. However, the presence of 1160 to 1150 million-year-old zircons in the Flinton Group, which seems to be largely of local provenance, may indicate that the Rb-Sr ages approximate the age of emplacement of the granitic component of these bodies. If so, then 2 ages of granite magmatism may be present in the area: a *circa* 1240 million-year-old magmatism comparable in age with the Methuen Suite (e.g., Addington Granite), and a *circa* 1160 million-year-old magmatism similar in age to Frontenac Terrane magmatism but of alaskite suite affinity (e.g., Abinger and Norway Lake granites).

STRUCTURAL HISTORY

Several structural studies have been conducted in the western and central part of the Mazinaw Terrane, most notably Venkitasubramanyan (1969), Thompson (1972), Rivers (1976), Chappell (1978) and Wolff (1982a). In general, 3 main periods of deformation resulting in folding have been recognized. These 3 periods of deformation are generally considered to have affected all supracrustal rocks in the Mazinaw Terrane, including the Flinton Group. D_1 produced F_1, northeast-trending isoclinal folds. An earlier deformation event (D_0) is recorded only in fragments within tonalite intrusion breccias in the area (Moore and Thompson 1980). D_2 also produced northeast-trending folds (F_2), generally coaxial to the F_1 folds; tighter structures; and local overturning of stratigraphy. The F_2 folds are readily evident on geologic maps of the area and include the Bishops Corners, Plevna and Ompah synclines.

Chappell (1978) suggested that thrusting in the Clare River area occurred late in D_1 or was post-D_1, and that the Clare River synform was a nappe. The décollement surface was located in a narrow metasediment band adjacent to the Clare River Group–Addington Granite contact. The décollement surface was folded during D_2.

Easton and Ford (1991) conclude that thrusting in the Mazinaw Lake area occurred during, or prior to, D_2 and F_2. The Norway Lake Granite is bounded by a shear zone to the south and north and may be a nappe. This shear zone also truncates the north limb of the Ompah syncline and the south limb of the Plevna syncline (Easton and Ford 1991). Chappell (1978) and Rivers (1976) both noted that metamorphic minerals overgrow the D_1-S_1 fabric in the Clare River and Fernleigh–Ompah areas, suggesting that thrusting may have occurred prior to peak metamorphism. This suggests that heating of crust was relatively slow, and is consistent with the observation that metamorphism and plutonism are not coincident.

The deformation history of the Flinton Group is more difficult to unravel. Rivers (1976) and Chappell (1978) both questioned the relation of the Flinton Group and the F_2 synclines, and Easton and Ford (1991) confirm that the Fernleigh belt is a northeast-dipping, southeast-facing homocline as suggested by Rivers (1976). Thus, the localization of the Flinton Group in synformal troughs may be an artifact of deposition; consequently, the synclines post-ulated by Moore and Morton (1986) do not exist. Depending on the timing of deformation, the Flinton Group may have been largely unaffected by the F_2-D_2 event. It was affected mainly by D_3, a period of coaxial, northeast-trending open folding and cleavage development and by D_4, a period of broad, northwest-trending open folds (F_4).

Connelly (1985, 1986) and Connelly et al. (1987) used structural evidence and geochronologic data in the Flinton area to suggest that the main fabric forming event in the region affected the Elzevir Tonalite, "Tudor metavolcanics" and the Flinton Group, and suggested that the Flinton Group was tectonically juxtaposed against the Elzevir Tonalite during the final stages of its crystallization. Certainly in the area studied, the Flinton Group is in tectonic contact with mafic metavolcanic rocks and the Elzevir Tonalite (Easton, unpublished data). If the major Grenville metamorphic event occurred at *circa* 1230 to 1180 Ma, then the major deformation fabric would be in all units, as noted by Connelly (1986), negating the argument that the Flinton Group be the same age as the Mazinaw Group (Connelly 1986).

The structural history of the Mazinaw Terrane is similar to that outlined by deLorraine and Carl (1986) for the Frontenac Terrane. Namely, a D_1 event resulting in northeast isoclines (F_1) and migmatization (M_1); D_2 thrusting and broad, northeast-trending folding (F_2); D_3 north-northeast-trending tight to isoclinal folds (F_3) resulting in local overturning and finally; D_4, resulting in gentle, northwest-trending folds (F_4). Wynne-Edwards (1967a, 1967b) outlined a similar structural history, although the more intense metamorphism in the Frontenac Terrane in Ontario apparently completely transposed the F_1 and F_2 events.

METAMORPHIC HISTORY

The metamorphic history of the Mazinaw Terrane is complex. Two, perhaps 3, regional metamorphic episodes affected these rocks; 2 prior to Flinton deposition. The first metamorphic event affected the metavolcanic and metasedimentary xenoliths preserved within tonalite intrusion breccias in the area. This may have been a contact metamorphic event. At least 1 regional metamorphic event preceded Flinton Group deposition. It was accompanied by high metamorphic temperatures and intense deformation including thrusting and folding. This event caused migmatization in and around the Abinger Granite and probably attained middle to upper amphibolite-facies conditions across most of the Mazinaw Terrane. During this episode, thrusting of metaplutonic masses into their current lenticular, northeast-trending pattern occurred and shear zones developed along the northern margins of most metaplutonic masses in the terrane. In addition, Mazinaw Group limestones were locally metamorphosed to dolomite marbles, which were subsequently eroded to form dolomite clasts present in Flinton Group conglomerates. This metamorphic event may have occurred in the period 1230 to 1180 Ma and may have been even older, perhaps coincident with Elzevir Terrane magmatism (*circa* 1240 Ma).

Much of the structural framework of the area was already established prior to Flinton Group deposition, as all shear zones have been overprinted by the latest regional metamorphism, which affects the Flinton Group. This latest regional metamorphism was one of lengthy heating with only limited deformation, mainly regional folding and cleavage development, and was followed by a period of slow cooling. The result of this lengthy heating episode has been extensive recrystallization, resulting in a great textural similarity, and lack of primary structures, in all rocks in the domain. A variety of porphyroblasts (*see* Figure 19.84), including biotite, plagioclase, magnetite, garnet, sillimanite, kyanite and staurolite, developed in rocks of appropriate bulk composition and overprint the structural fabric in the rocks. Conditions were upper greenschist to middle amphibolite facies (cf., Chappell 1978). In the Clare River area, Bright (1986a) and Chappell (1978) both locally document a lower metamorphic grade assemblage overprinting a higher grade assemblage.

BASEMENT

Two basement-cover relationships may be observed in the Mazinaw Terrane: 1) the Flinton Group metasedimentary rocks were deposited on a deformed and metamorphosed Grenville supracrustal and plutonic basement: and 2) tonalitic rocks may have served as basement to the Mazinaw Group. In the latter case, the Abinger Granite intrudes ultramafic and metagabbroic, and metatonalitic, rocks similar to those found in the Grimsthorpe area to the west (Easton and Ford 1991). In addition, slivers of ultramafic and metagabbroic rocks are found in the belts of layered quartzofeldspathic gneisses separating supracrustal-granite packages in the northern Mazinaw Terrane (Easton and Ford 1991). These mafic and ultramafic rocks are similar to rocks in the Grimsthorpe Domain, greater than *circa* 1270 million years (Lumbers et al. 1990) in age. This relationship may also imply that the Northbrook, Cross Lake and other tonalite bodies in the Mazinaw Terrane are similar in age to the Elzevir and Weslemkoon tonalites. It is critical to the tectonic history of the area to know if the volcanic rocks in the Harlowe–Marble Lake area are similar in age to those in the Belmont Domain (ca. 1250 Ma and hence younger than the tonalites) or are indeed older than the tonalites. To the south, Bright (1986a) has suggested that the Mellon Lake Complex of tonalite and younger granitic phases may also served as basement to supracrustal rocks in the Clare River synform.

MINERAL DEPOSITS

Many mineral deposit types are absent or rare in the Mazinaw Terrane, most notably contact-metasomatic iron deposits and copper-nickel deposits, both of which are associated with Lavant Suite gabbro-diorite plutons, which are uncommon in the Mazinaw Terrane. The Radenhurst–Caldwell deposit (Carter et al. 1980; Easton 1988a) consists of massive magnetite hosted in mafic metavolcanic and synvolcanic metagabbroic rocks and is probably the best described deposit of this type. It contains elevated light REE contents (1 to 2% LREE, Easton 1988a), an atypical feature of Central Metasedimentary Belt magnetite deposits.

Two broad groupings of metallic mineral deposits are present in the terrane. The first consists of sulphide mineralization (pyrite, pyrrhotite, chalcopyrite, arsenopyrite, magnetite and bornite) associated with metavolcanic rocks, generally as disseminations or within interflow metasedimentary rocks (e.g., Blithfield pyrite deposit, Carter et al. 1980; Easton 1988a). The second consists of quartz-carbonate-sulphide vein and sulphide mineralization found in close proximity to the Flinton Group unconformity and within the Flinton Group itself. Most significant mineral deposits in the terrane are of the second type.

Vein mineralization associated with the Flinton Group unconformity varies with host rock composition. In mafic metavolcanic host rocks, gold-arsenopyrite-pyrite mineralization is typically associated with quartz ± carbonate veins (e.g., Addington Mine, Papertzian 1984, Dillion 1985, Laing et al. 1987, Harnois 1987, Harnois and Moore 1989; Ore Chimney Mine, Meen 1944, Papertzian 1984, Moore and Morton 1986; O'Donnell occurrences, Moore and Morton 1986). Metawackes also host similar gold-bearing vein systems (e.g., Boerth and Webber properties and the James Mine, Smith 1958, Pauk 1987). Carbonate rocks also host gold-pyrite-sphalerite-galena-bearing, dolomite-quartz vein systems near the unconformity (e.g., Cook, Helena, Stead, Camgar, Dome and Pay Rock gold occurrences, Papertzian 1984, Moore and Morton 1986).

Easton and Ford (1991) noted that most gold occurrences in the northern Mazinaw Terrane are associated with an older shear zone near the Flinton Group unconformity. This older shear zone may have been a critical factor in siting mineralization and influencing subsequent Flinton Group deposition. Although most of these deposits are located near the unconformity with the Flinton Group, the variability in mineralization types and host-rock and vein compositions, suggests that the specific genetic process varied considerably between deposits. Weathering near the unconformity surface may have locally concentrated metals which were latter mobilized into veins during metamorphism. Faulting and fluid movement associated with the extensional environment in which the Flinton Group was deposited may have been an important factor in mineralization, especially as nondetrital tourmaline is closely associated with Flinton Group metasedimentary rocks and many of these deposits. In addition, fluids associated with the Flinton Group may also have aided in metallogenesis, with the unconformity serving as a fluid channel.

Mineralization is found in 4 settings within the Flinton Group itself. 1) In the Marble Lake area, black pyritic-graphitic schists of the Myer Cave, and, locally, the Fernleigh formation contain stratiform pyrite-chalcopyrite-sphalerite mineralization. In the Clare River synform, similar mineralization has been described by Bright (1986a) from the Kaladar Mines Ltd. property and the Amiko mica property. 2) In the Clare River synform, pyrite-pyrrhotite deposits are closely associated with basal muscovite-schists of the Flinton Group (Bright 1986a). Examples include the Canada and Hungerford mines near Sulphide (Bright

1986a) and the Ontario Sulphur and Donahue Creek deposits (Carter 1984). 3) Bright (1986a) reports strata-bound zinc mineralization with minor pyrite and chalcopyrite in dolomitic marbles in the Clare River area, possibly equivalent to the Myer Cave Formation (e.g., Spry Zinc occurrence, Carter 1984; possibly the Glenshire property, Carter 1984, Bright 1986a). 4) Best known in both areas are crosscutting sulphide-quartz veins in the dolomite horizon of the Myer Cave Formation containing pyrite-chalcopyrite-sphalerite-galena mineralization. Examples include the Barrie Syndicate and Mazinaw Base Metals occurrences (Moore and Morton 1986) and the Ganda Silver mines occurrence (Pauk 1987). This mineralization is probably the result of remobilization of Kuperfiescher-type, copper-zinc mineralization related to the original deposition of the Flinton Group and represented by types 1 and 3. The Zambian copper belt (e.g., Lefevbre 1989) is similar in many respects to the postulated environment for Flinton Group deposition and mineralization both within, and adjacent to, the Flinton Group. The low sulphur isotope content of the Spry zinc deposit (2.6‰, Sangster 1991) is consist with it being derived from basinal brines that flowed through adjacent metavolcanic strata (Sangster 1991). It is noteworthy that lead is more abundant in Mazinaw Terrane sulphide occurrences than elsewhere in the Central Metasedimentary Belt. This may be a reflection of the fact that supracrustal rocks of the Mazinaw Terrane were deposited on a basement complex of tonalite and gabbroic rocks. Although this basement is only slightly older than the overlying supracrustal rocks, it may have served as a source of lead.

The Clyde Forks copper-antimony-silver-mercury-barite occurrence (Nikols 1972; Carter et al. 1980; Carter 1984) represents a unique deposit type in the Central Metasedimentary Belt. It is a conformable deposit hosted in dolomitic marbles and quartzofeldspathic metasedimentary rocks of the Mazinaw Group, close to the trace of the Clyde River fault. Nikols (1972) proposed that mineralization resulted from hydrothermal springs discharging into a carbonate basin. Flinton Group rocks are present about 1 km east of the deposit, and it is not known if hydrothermal activity related to Flinton Group deposition played a role in the genesis of this deposit (in favourable host rocks).

Uranium- and thorium-bearing pegmatite veins occur in 3 concentrations in the easternmost Mazinaw Terrane: 1) the Cross Lake area (Ford and Charbonneau 1979; Ford 1982), 2) near Barnett Chute (Easton 1988a) and 3) the Salmon River area (Bright 1986a). Compared to Bancroft Terrane pegmatites, the Mazinaw Terrane pegmatites have distinctly lower regional equivalent thorium values (less than 5 ppm; Ford 1982). The concentration of pegmatites in these areas may reflect partial melting of metaplutonic rocks (and possibly adjacent paragneisses, cf., Ford 1982) related to upper amphibolite metamorphism (all are located in large plutonic complexes in areas of peak metamorphic conditions). The lower thorium content may reflect their derivation from tonalitic to granodioritic metaplutonic rocks (Mellon Lake Complex, Cross Lake Tonalite, Bartraw dome). As in the case of other Grenville uranium vein deposits, gamma-ray spectrometry has identified all potential targets in the Central Metasedimentary Belt, and the most favourable deposits are those found in pegmatites hosted in adjacent metavolcanic rocks (cf., Bright 1986a; Pauk 1987). Pauk (1987) describes the exploration history of the higher grade Cross Lake occurrences. Pauk (1987) reports higher uranium values at surface then at depth, probably due to supergene enrichment during weathering. Locally, molybdenum mineralization is present in association with the pegmatites, most commonly in the less radioactive pegmatites.