This is to certify that the

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Occurrence of Sills in
Extensional Environments

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Marco Antonellini

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OCCURRENCE OF SILLS IN
EXTENSIONAL ENVIRONMENTS

By

Marco Antonellini

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ABSTRACT

OCCURRENCE OF SILLS IN EXTENSIONAL ENVIRONMENTS

By

Marco Antonellini

Upper crustal extension in the Midcontinent Rift System consists of simple shear.

Horizontal sheet intrusions follow the detachment plane through the basement and into the sedimentary sequence of the Northern Lake Superior Region.

Two different kinds of shear strain are connected with the emplacement of the intrusions in the sedimentary sequence. The first kind of shear strain is connected to the regional extension and is organized along bedding-parallel surfaces along which the sills seem to have preferentially intruded. The second kind of shear strain is related to the sill intrusion itself and is responsible for drag folds and structural elements rotations in close proximity of the boundary layer and the country rock. These secondary structures define a northerly direction for magma emplacement.

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INTRODUCTION

The associations of sills with extended terrains is widely documented in the geological literature. Examples include the North Sea, the western United States, the Russian Triassic Rift System and the Oslo Rift (Caston 1981,Thorpe and McDonald 1985, Gayduk 1987, Vdovykyn 1980, McCarthy and Thompson 1988, Schonwandt and Petersen 1983, Morgan and Ramberg 1987). Emplacement of sills in soft sediments at spreading centers has been described by Einsele (1985) in the Gulf of California. Recently McCarthy (1988) has suggested that the reflectors in the lower crust of the Basin and Range Province as well as those in the Western North Atlantic (Fig. 1), could be sills emplaced during the extension of the lithosphere.

These reflectors in the lower continental crust as well as the "dipping events" in the oceanic crust have been subjected to intensive study by seismologists (Pockalny and Fox 1989, Mutter and Buhl 1989) who tend to relate them to intrusive mafic bodies, lava flows or extensional detachment planes. Seismic layering of the lower crust is a characteristics of paleorifts (Morgan and Ramberg 1985) and

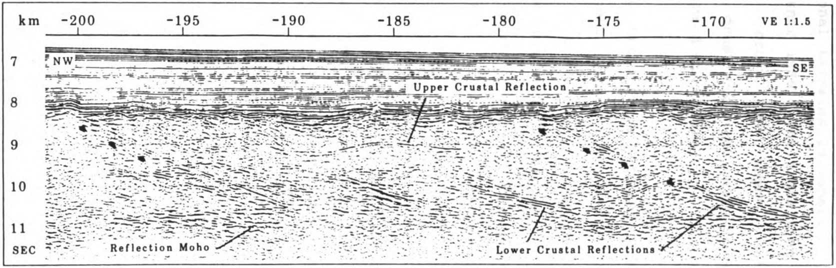


Figure 1. Reflection section from the North Atlantic. Dipping reflectors are visible in the crust. The lower crust has a laminated aspect given by horizontal reflectors (from McCarthy and Thompson, 1988)

of extensional terrains in general (Gaijewski and Prodehl 1987, Meissner 1986). The transition between layered lower continental crust and upper, seismically transparent continental crust was classically "considered" or "confused" with the Conrad discontinuity.

In this study it is proposed that sills take advantage of detachment planes developing in the crust during extension and are emplaced along them. The basic assumption is that a simple shear mechanism, in the form of a crustal detachment plane (fig. 2) associated with shear zones, controls the extensional deformation during rifting (Wernicke, 1985; Wernicke and Burchfiel, 1982 and Lister and others, 1986). The classical interpretation of sills place them in a compressional environment (Anderson, 1951). However, this theory, mainly based on mechanical arguments, is not consistent with the geologic evidence. To shed some light on this ambiguity it is necessary to review factors believed to be important in controlling sill intrusion and the theoretical mechanisms of emplacement proposed during the last four decades.

Anderson (1951) introduced the concept that emplacement of sheet-like intrusions was controlled by stress conditions; a sill or a dike will intrude the country rock perpendicular to the least principal stress direction so that the magma can open up fractures along planes across

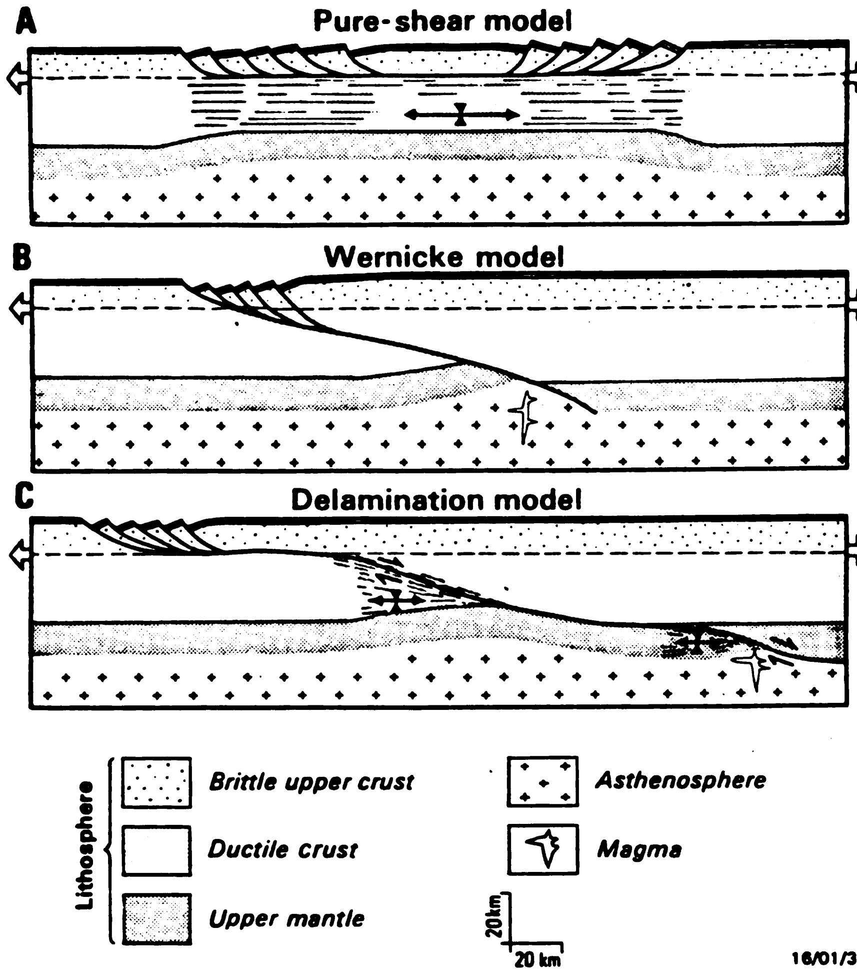


Figure 2. Three models for continental extension (from Lister and others, 1986)

which the pressure is low. Furthermore, to explain the association of dikes and sills (an apparent contradiction) he envisaged a mechanism in which the magmatic pressure could change the stress conditions in the country rock if the three principal stresses were close in value. In his model the direction of the least principal stress can rotate 90 degrees during the emplacement of the intrusion. These concepts have a sound mechanical basis, are ultimately correct in the framework of a theoretical model and have become firmly established in the geologic literature (Delaney and others 1986, Nakamura 1977, McCarthy and Thompson 1988, Zoback and Zoback 1980).

However in strongly anisotropic rock, principal stresses may not be the controlling factor. Mudge (1968) reviews the occurrence of concordant sills in 54 localities worldwide and concludes that three factors seem to affect the depth at which intrusion occurs: 1) a well defined parting surface such as a bedding plane or an unconformity, 2) the lithostatic pressure of the overburden and 3) the presence of a fluid barrier over the intrusion. The latter together with magmatic pressure seem to constrain the depth of emplacement to a range of 3000 to 7500 feet. At depths shallower than 3000 feet, shales behave in a ductile manner and will not act as a fluid barrier. The deeper limit of 7500 feet is considered by Mudge to be controlled by the magmatic pressure; however, it could be an artifact introduced by limiting the observations to exposed flat lying sedimentary sequences.

Overburden control was previously considered by Bradley (1965) in his "floatational" theory of sill intrusions. His theory is based on a simple hydrostatic model (fig. 3a): high density magma intrudes below low density sediments and the low density sediments float over the magma. To account for the variations in overburden pressure caused by surface relief the sills will follow a compensation surface at depth which is complementary to the topography (fig. 3b). Along the compensation surface the magma pressure and the lithostatic pressure are the same. The sills that follow bedding planes during intrusion maintain their level of compensation through a stepping process which give rise to transgressive sills (Bradley 1965, Meyboom and Wallace 1978). Bradley states that a well defined parting surface must also be present and that there must be a high enough fluid pressure to aid the steam injection mechanism of the magma.

Gretener (1969), following Anderson's (1951) principles, explains the intrusion of sills in a layered sequence introducing the concept of a stress barrier. He considers the formation of a horizontal sheet intrusion to be a major reorientation in the direction of magma emplacement of a vertical dike beneath sedimentary strata in which the minimum principal stress is vertical. These strata represent a stress barrier to the ascending magma. He rejects the presence of the parting surface as well as the

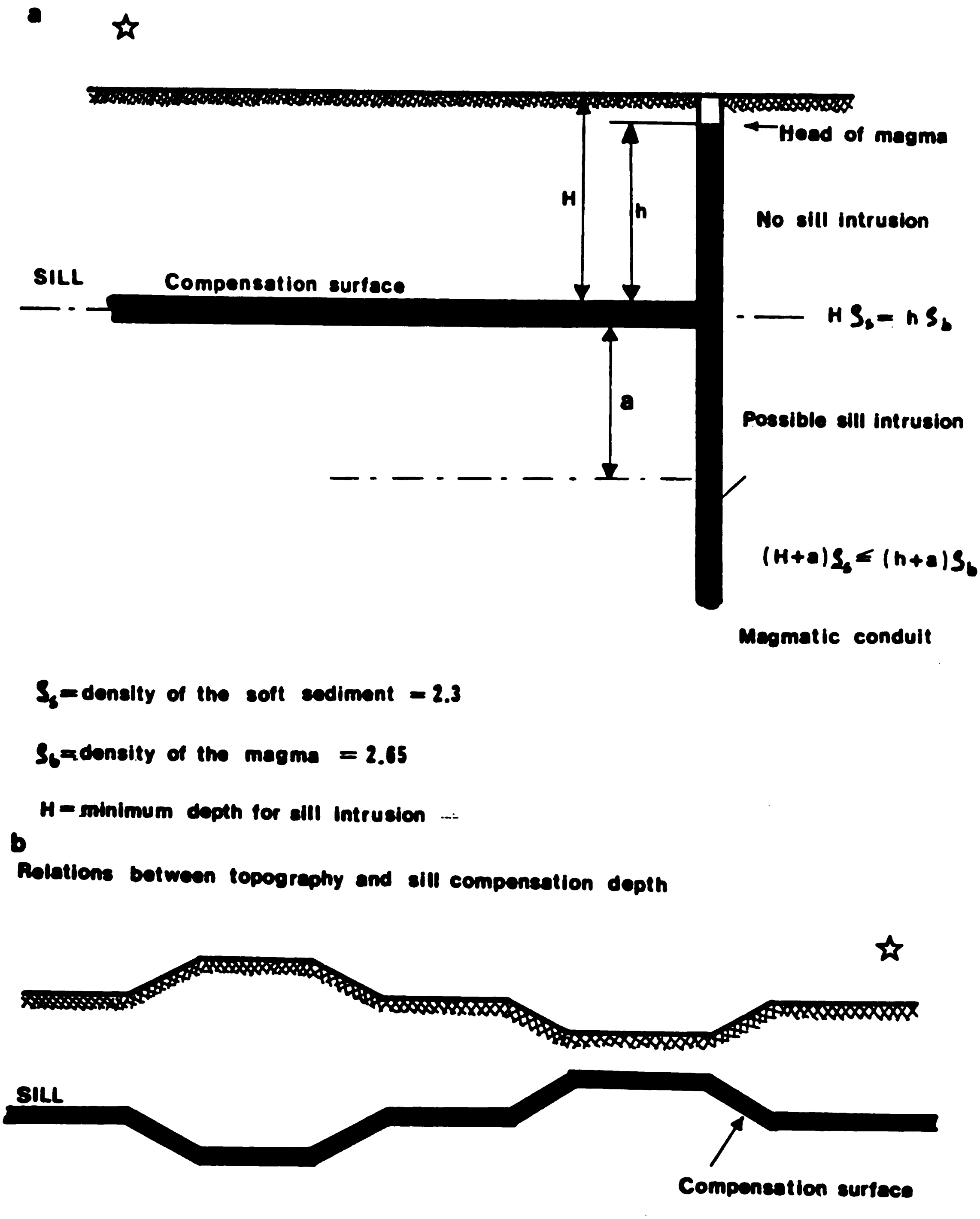
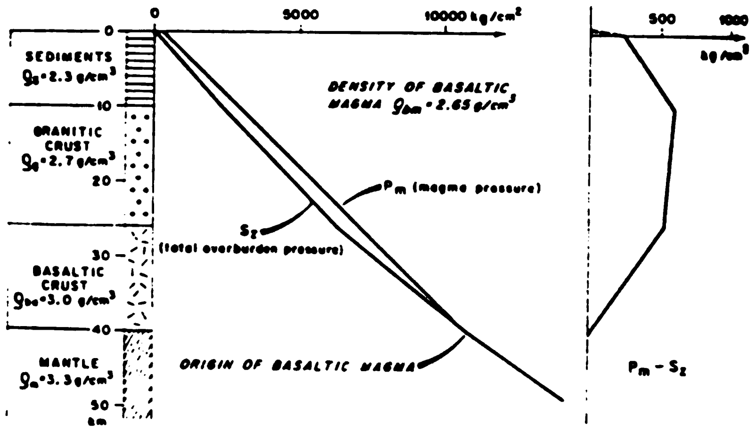


Figure 3. Schematic representation of the principles of Bradley's (1965) floating theory. Hydrostatics governs sill emplacement at any time

overburden as controlling factors for the intrusion. Such constraints should lead to the presence of sills along major unconformities between the crystalline basement and the supracrustal sedimentary cover. The parting surface is well developed along these unconformities and the differential pressure between overburden and magma would be at a maximum. However sills are rarely found in this situation. He proposes a model related to the effective Poisson's ratio. Let x and y (fig. 5) be the horizontal principal stress directions and z the vertical one. Continued stress build up parallel to x will cause an increase in the y-stress according to the effective Poisson's ratio of the rock. The vertical direction, z will provide the easiest direction of release providing that horizontal surfaces act as stress barriers to the rise of magma (fig.4 and 5).

Pollard (1973) derived a model for sheet intrusions based on elastic deformation around a pressurized cavity in an homogeneous medium. He modeled the stress field, based on Anderson's ideas using the finite element method. He was able to account for many of the observed features around the intrusions, especially at the terminations and in the zone of interference between different magma bodies. Three mechanisms for sheet intrusions are considered in his study, extension faulting, brittle faulting and ductile faulting. He is able to mathematically constrain the width of the stress field around the intrusion but only for a homogeneous country rock (fig. 6). He shows experimentally (fig. 7) that a


0
5000
10000 kg/cm²
SEDIMENTS
ρ_s=2.3 g/cm³
GRANITIC CRUST
ρ_g=2.7 g/cm³
BASALTIC CRUST
ρ_{bc}=3.0 g/cm³
MANTLE
ρ_m=3.3 g/cm³
km
DENSITY OF BASALTIC MAGMA ρ_{bm}=2.65 g/cm³
P_m (magma pressure)
S_z (total overburden pressure)
ORIGIN OF BASALTIC MAGMA
500
1000
kg/cm²
$P_m - S_z$

b

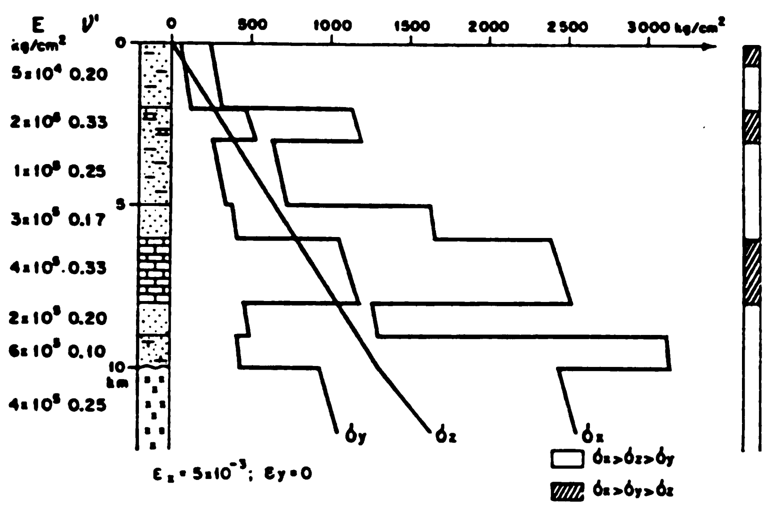

E kg/cm²
ν'
5x10⁴ 0.20
2x10⁵ 0.33
1x10⁵ 0.25
3x10⁵ 0.17
4x10⁵ 0.33
2x10⁵ 0.20
6x10⁵ 0.10
4x10⁵ 0.25
0
500
1000
1500
2000
2500
3000 kg/cm²
km
σy
σz
σx
σx>σz>σy
σx>σy>σz
εx = 5x10⁻³; εy = 0

Figure 4

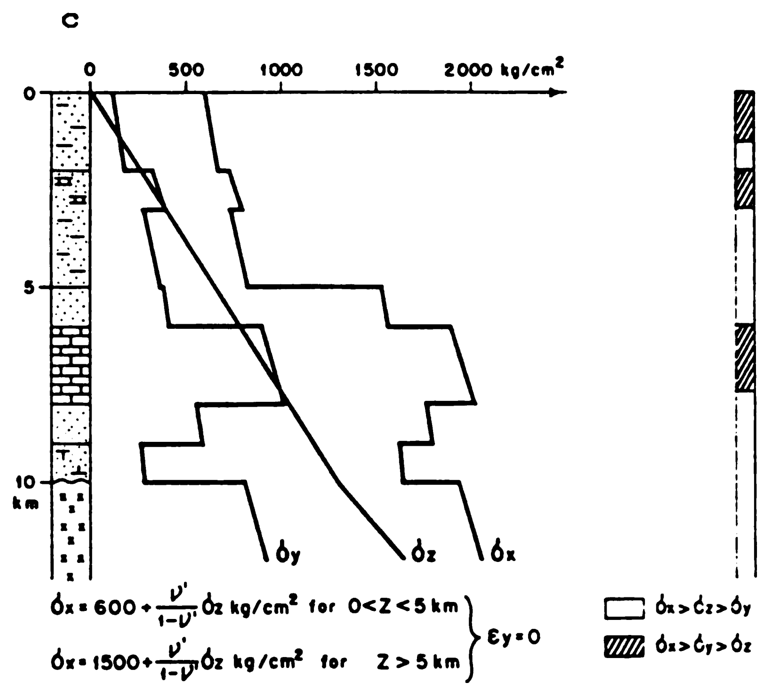


Figure 4. Effective Poisson ratio and sill intrusions. (a) Relations between overburden pressure (Sz) and magma pressure (Pm) in the crust and sedimentary cover. (b) State of stress in the basement and sedimentary sequence for $\epsilon x = 5x10^{-3}$ (positive compression) and $\epsilon y = 0$ (confinement in the y direction). (c) State of stress in the basement and the sedimentary sequence for the values of horizontal stress given in the figure (from Gretener, 1969).

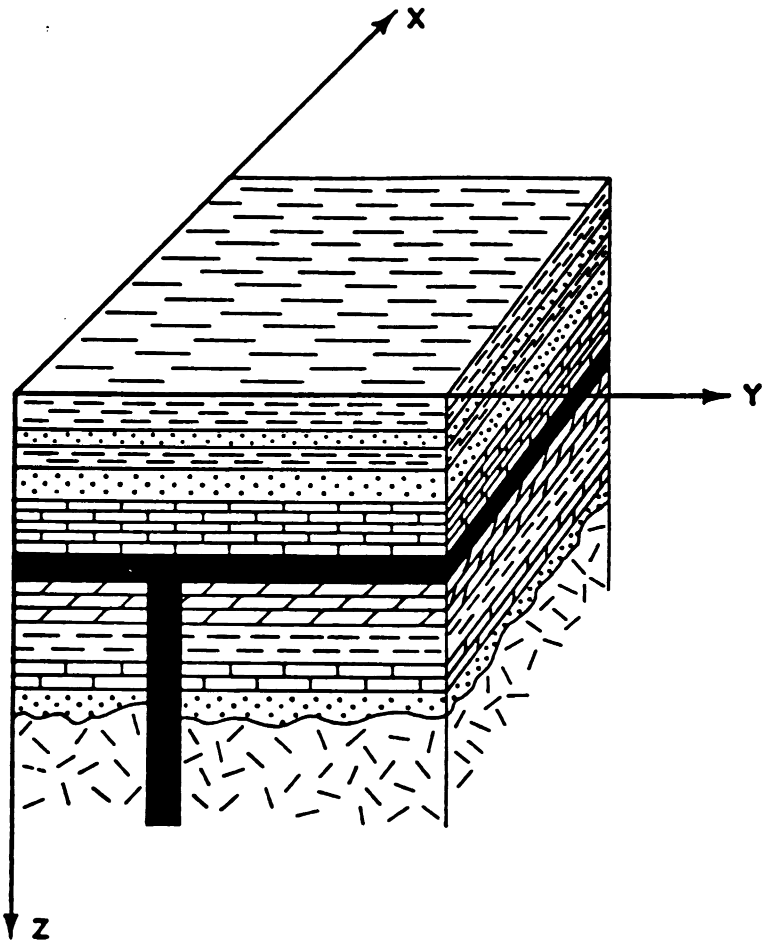


Figure 5. Orientation of the reference axes for the values given in figure 4 (from Gretener, 1969)

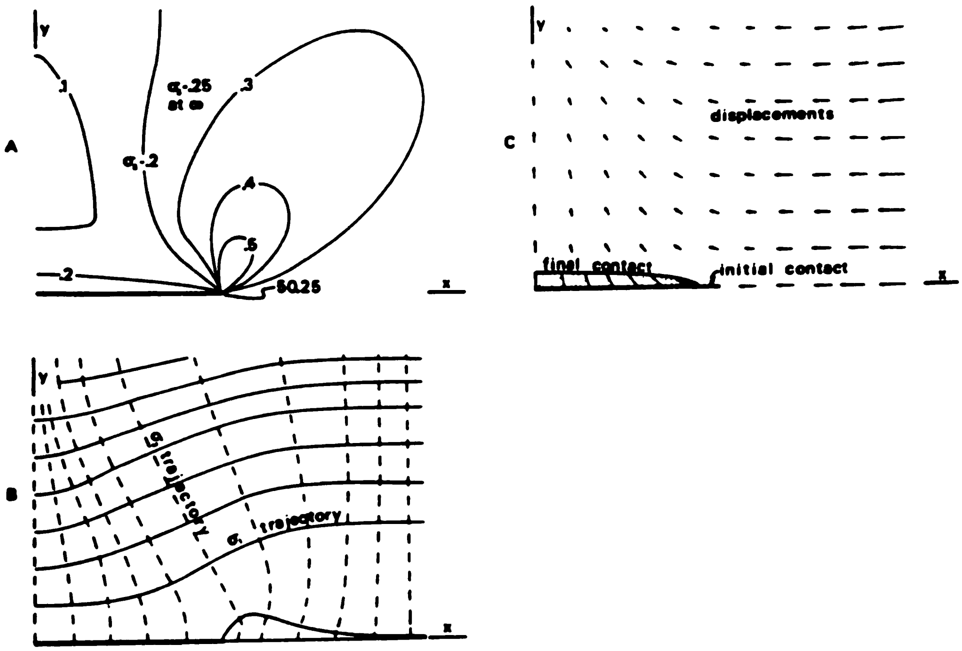


Figure 6. One half of the sheet intrusion is represented by the heavy line parallel to the x axis. (a) Contour map of the principal shear stress. (b) Trajectories of the principal normal stresses σ_1, σ_3. (c) Displacements are represented by the relative length of the arrows (from Pollard, 1973)

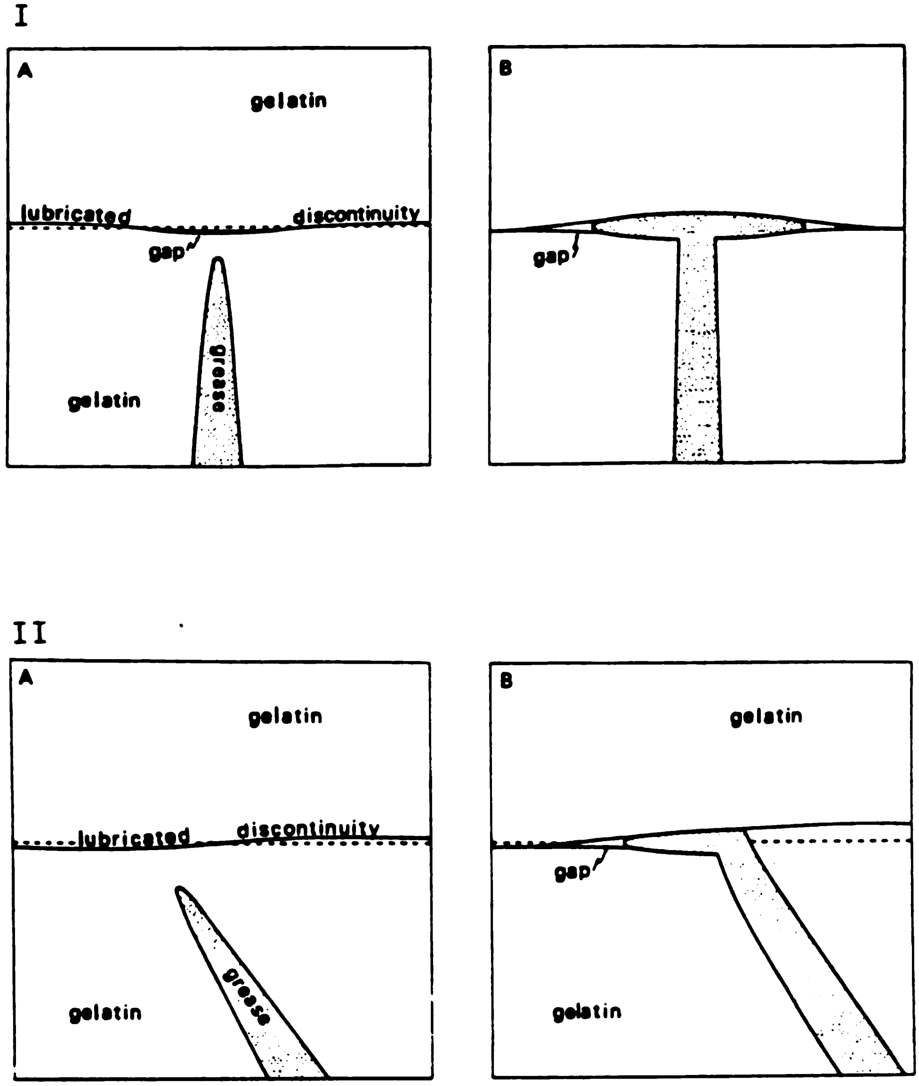


Figure 7. Experimental model for intrusions. I- Sketch of an experiment in which grease was intruded into gelatin at right angle to a lubricated discontinuity. II- Same experiment but this time the grease was intruded into gelatin at an angle of less than 90 degrees to a lubricated discontinuity (from Pollard, 1973).

the presence of a discontinuity is important in driving the flow of the intrusion.

Renewed interest in the problem of the mechanics of sill emplacement recently was stimulated by McCarthy and Thompson (1988) who relate the horizontal lower crustal seismic reflectors in the extended regions of the Western United States to layered magmatic intrusions. They propose that the injection of sills in the lower crust of an area undergoing rifting is possible because there the stresses are relaxed due to the rheology of the rocks under ductile conditions. For sill emplacement in the upper brittle crust they invoke a mechanism analogous to that of Anderson's (fig. 8).

Emerman and Marrett (1990) question the view that sheet like intrusions form because country rock fractures more easily along a plane that is perpendicular to the axis of least compressive stress. They consider an inviscid flow of low viscosity for the magma sheet to be of prime importance and suggest that preexisting anisotropies in the country rock control the pattern of intrusion. This point will be discussed in the chapter concerned with the shear stress at the walls of sills. The ratio of the volume of an intrusion to its surface area, which is a direct function of the viscosity, is also considered by Emerman and Marrett (1990) and McCarthy and Thompson (1988) as way of differentiating between intrusions of large batholiths and stocks of viscous sialic magma or salt from the sheet intrusions of low

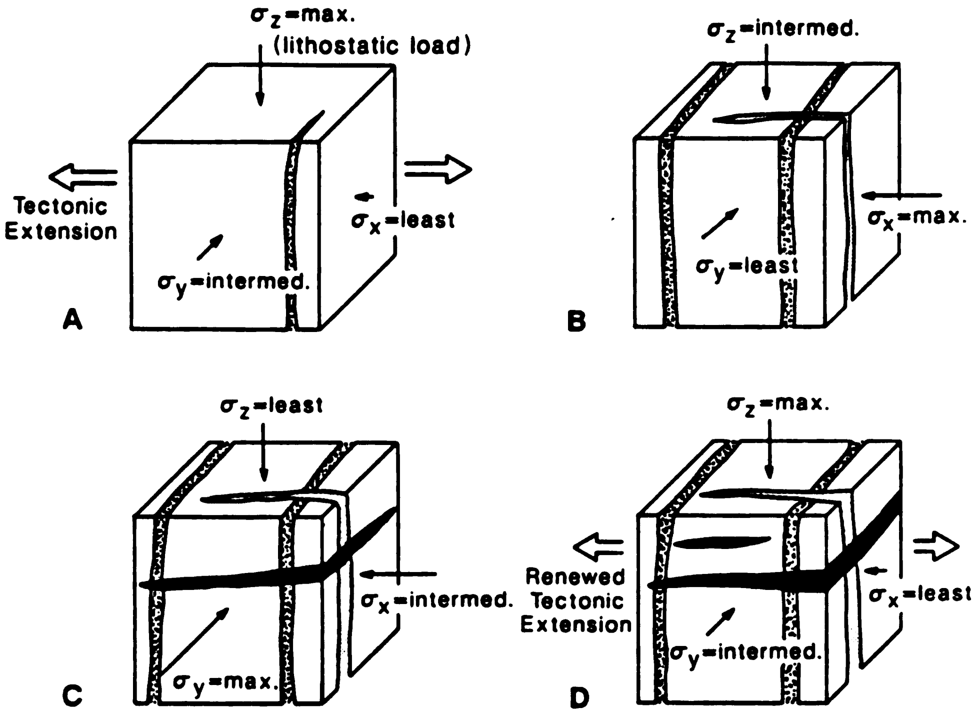


Figure 8. Schematic explanation for intrusion of horizontal sheets fed by dikes in an extending terrain. The theoretical basis is that discussed by Anderson (1951). There is not much difference between the magnitudes of the principal stresses at the beginning. Vertical intrusions can increase the horizontal normal stresses and decrease the vertical one so that horizontal intrusions can occur (from McCarthy and Thompson, 1988)

viscosity basaltic magma.

As we can see from the historical development of the ideas about the mechanics of sill emplacement, different authors have focused their attention on various specific aspects of the problem and developed theoretical models to explain them. However no attempt has been made to include all these factors in a single model. The hydrostatic principles of Bradley (1965) are important in explaining the range of depths at which intrusions can occur and the interconnection between magmatic pressure and lithostatic load. Anderson's (1951) model explains fracture development in homogeneous country rock without preexisting fractures. On the other hand the role of preexisting anisotropies and discontinuities has been recognized by most of the authors (Pollard 1973, Mudge 1968, Bradley 1965, Emerman and Marrett 1990) as a determinant factor in driving the emplacement of the magma. I will take this last consideration as the mechanical basis for the development of the model presented in this thesis together with the observation that mafic intrusions are not extremely common in compressional tectonic environments, whereas they are almost the rule in the extensional ones.

To conclude I would like to review ideas on the driving forces involved in magmatic intrusions.

First of all we can define the vertical climb force of the magma, Fz, as

$$Fz = (\rho - \rho')\ g\ V$$

where ρ is the weighted mean density of the country rock, ρ' is the magma density, g the acceleration of gravity and V is the total volume of magma per unit width of crack length within the feeder dike. It is apparent that a controlling parameter is the contrast in density between the country rock and the magma (fig. 9). The intrusion zone is located where $\rho = \rho'$ and the magma becomes neutrally buoyant. The lower the density of the magma the larger is the climb force; the temperature and the contrast in density between its solid and liquid state are important factors to consider (fig. 10).

It is possible to derive the magma driving pressure both for a Newtonian and a Bingham magma which rises opening cracks through the lithosphere. According to Weertman (1964, 1971b) for a Newtonian fluid the magma driving pressure is

$$Pd = Pt - Pl = [\rho_1\ g\ (2\ Lc + z_1) - \rho'\ g\ h\ ] - \rho_z\ g\ z(h)$$

where Pt is the total magma pressure, Pl is the lithostatic pressure (ρ_z g z), z is the thickness of the overburden, z_1 is the depth of the intrusion, Lc the critical crack length (for $|L| > |Lc|$ the crack is closed), z and z(h) are evaluated at Lc + z_1 - h and h is the height of the observer above the bottom of the crack.

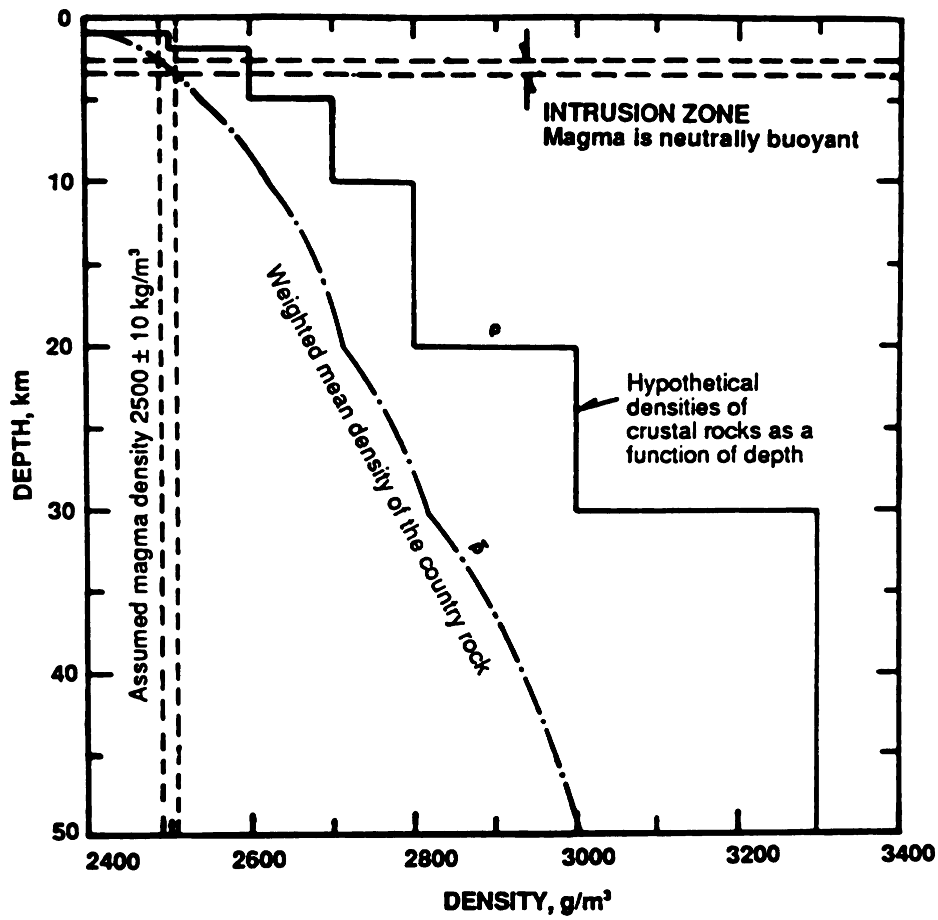


Figure 9. Weighted mean density as a function of depth for an assumed range of densities within the lithosphere. The intrusion zone shown assumes a silicic magma with a liquid density of 2500 Kgxm^{-3} (from Corry, 1988)

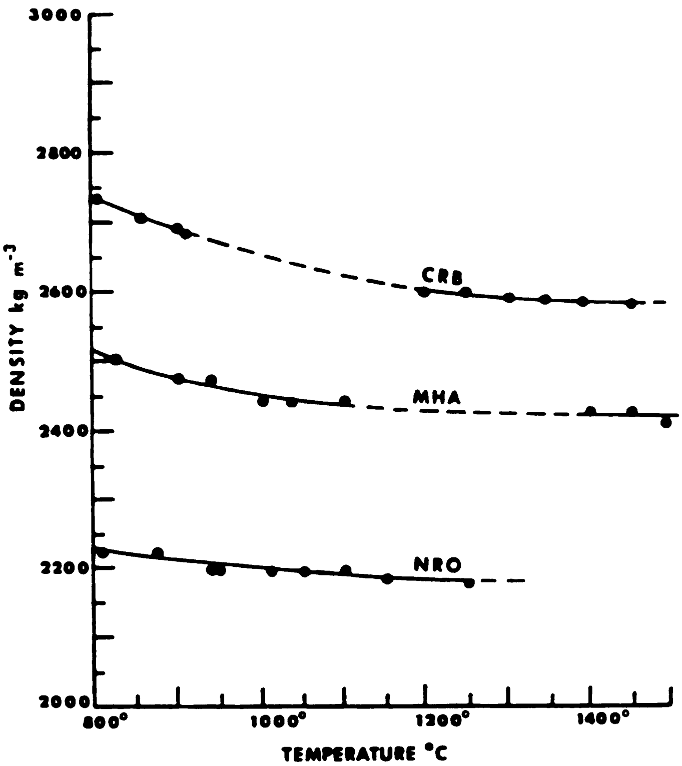


Figure 10. The variation of density as a function of temperature for three types of igneous rocks: Columbia River Basalt (CRB). Mount Hood Andesite (MHA), and Newberry Rhyolite (NRO) (from Corry, 1988)

This study concentrates on basaltic magma and the Newtonian assumption is used throughout.

The local tectonic (deviatoric) stress could be sometimes an important driving mechanism for magma emplacement but it is impossible to quantify and its effect cannot be considered.

HYPOTHESIS TO TEST AND AREA OF STUDY

The hypothesis being tested in this study is that during simple shear continental rifting, magmatic intrusions can take advantage of the detachment plane and rise through the crust. Once the detachment plane enters the sedimentary sequence it is deflected along the bedding planes providing easy passage for the sill intrusions (fig. 11 and 12).

The concept proposes that in the lower crust and in the basement below the sedimentary cover, magma tends to concentrate along the detachment surfaces (Cambray 1988 a and b) and then to step up along releasing bends. This process creates horizontal intrusions in the continental crust and greatly contributes to its thickening and growth. A similar process has been recently documented to occur for sialic intrusions by Hutton (1988) and Hutton and others (1990) along a pull-apart of a vertical shear zone at Strontian (Scotland) and in the form of granitic horizontal sheet intrusions along an active extensional shear zone in the Proterozoic continental crust of South Greenland.

The area chosen to test this hypothesis is the Midcontinent Rift System (MRS), which has particular advantages over other extensional areas. First of all in the MRS the pre-rift sequence is directly exposed and is not buried by a thick rift fill sequence as in others rifts or passive continental margins of the world. Second, the MRS includes a well-developed volcanic and magmatic association in which the Duluth complex of Minnesota and the Logan sills

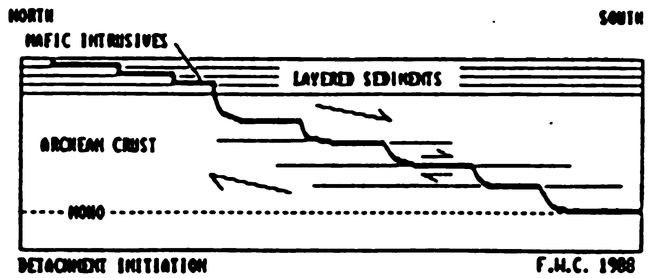


Figure 11. Schematic model of simple shear stretching of the crust with mafic intrusions that intrude parallel to the flats of the detachment plane (from Cambray, 1988)

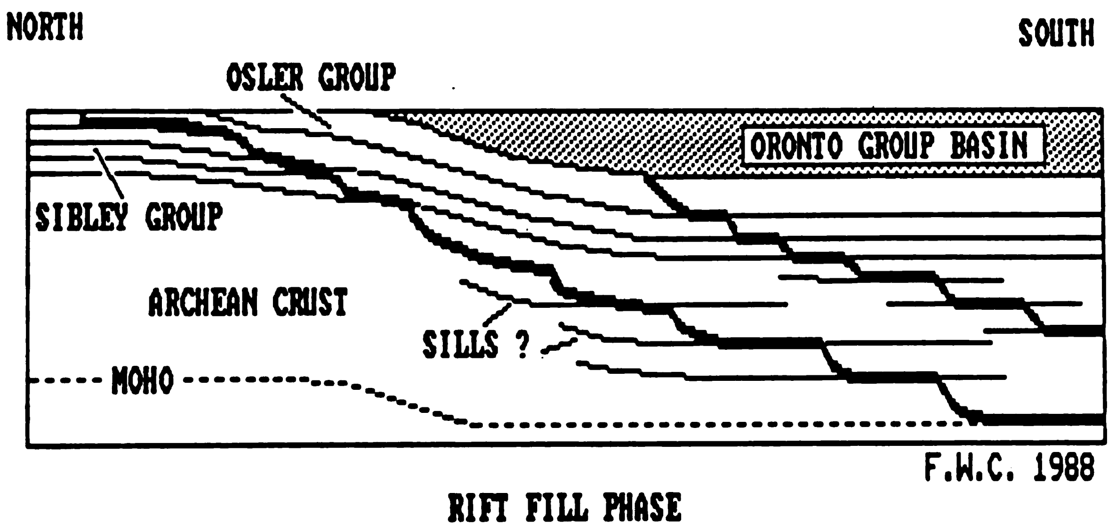


Figure 12. Model of figure 11 applied to the Midcontinent Rift System (from Cambray, 1988)

of Northern Ontario probably represent the outcrop of the detachment plane (Cambray 1988).

Critical tests for the hypothesis are the relations between shear strain present in the sedimentary sequence and its association with sill intrusions. The study has concentrated on the timing of events, whether they are related to intrusion or shear associated with the detachment. To do this it has been necessary to consider the rheological conditions of the sedimentary strata, the width of the area affected by deformation close to the sill and the width of the shear deformation supposed to be connected to the detachment plane. The relative importance of these elements will be addressed in the description and elaboration of the structural elements in the locations studied. To understand these relations it is necessary to define the detachment plane as envisaged in this work and to briefly review the geological and stratigraphic setting of the area of study.

The detachment plane

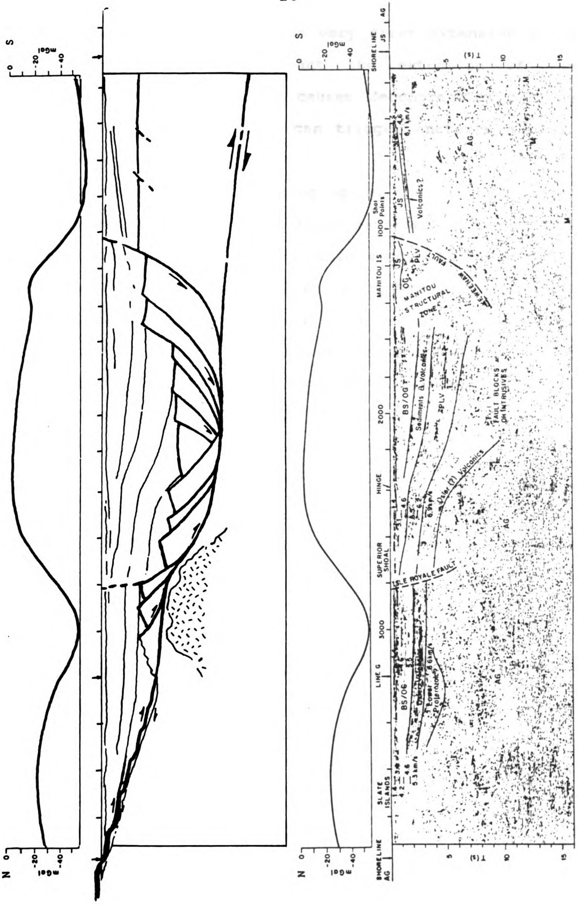
The detachment plane considered in this study is south dipping and is not coincident with that envisaged in the geological reconstruction of Cannon et al. (1989). These authors present an interpretation of line A (fig. 15) of the 1986 GLIMPCE seismic refraction survey. This line lies to the west of the Thiel fault and is on-shore, structurally continuous with the Nipigon area. In their reconstruction they outline a north dipping detachment plane as an extension of the Keweenaw fault, which defines a large crustal half graben structurally differentiated in two basins divided by a central horst.

I consider this interpretation to be inconsistent with some of the geological evidence. In particular the structure is not a simple half-graben and the presence of a horst in proximity to the Isle Royale Fault (fig. 13, 14 and 15) causes geometrical problems that can be solved only with the presence of faults on both sides of it.

The presence of this horst and of a structurally differentiated sub-basin to the north in this area is hard to fit with the geology of the Northern Lake Superior shore. There are no major faults even in the seismic sections which could represent the north side of this horst.

Furthermore the gravity low seems to be too pronounced to be just the result of a small basement horst. It is more consistent with subcrustal intrusions of low density material, similar to those exposed on St. Ignace (fig. 24). The overall scenario of a south dipping detachment plane is represented in fig. 13 superposed on the interpretation of line A made by Cannon et al. (1989). The situation is very similar to the geometries of faulting proposed by Wernicke (1981, 1982, 1985) for the metamorphic core complexes of the Western United States, with the only difference that the detachment can bottom down in the lower crust instead of the mantle. The presence of low density intrusives is the result

Figure 13. Tentative interpretation of the crustal geometry for the Midcontinent Rift System. The detachment plane is south dipping. The interpretation of the reflectors in the sedimentary sequence is the same as that of Cannon et al. (1989) except for a few minor changes in proximity of the Isle Royale Fault consisting in a different interpretation of some north dipping reflectors



S
N
mGal
SHORELINE
JS
AG
T (s)
Shot Points
MANITOU IS
HINGE
SUPERIOR SHOAL
LINE G
SLATE ISLANDS
KEWEENAW FAULT
MANITOU STRUCTURAL ZONE
ISLE ROYALE FAULT
Volcanics?
Sediments & Volcanics
FAULT BLOCKS OR INTRUSIVES
Osler (?) Volcanics
Lower Proterozoic
PLV
BS/OG

of magma emplacement during the very first extension along the detachment plane. In fact the extension of the lithosphere and its thinning causes decompression melting and increased heat flux that can trigger anatexis also in the upper crust.

A detachment plane stepping up on the northern shore ties very well with the shear strain recorded in the surface geology of the Nipigon area in this study. Furthermore it fits the lack of faults in Northern Lake Superior and explains the presence of intrusions in a ramp of the plane. This ramp could be original or the result of bending during unloading and magma emplacement. In coincidence with the ramp and perhaps at the time of its formation the Isle Royale Fault may have become active and cut off the northern part of the detachment plane which remained inactive during the subsequent major extension in the deep southern basin. The Isle Royale fault can represent the excisement of the first detachment plane by a second younger one, a process very well documented by Davis and Lister (1988) in the Whipple Mountains of Arizona.

Cambray's proposal of a south-dipping detachment plane as interpreted in this thesis (fig. 13) is directly opposed to that of Cannon et al. (1989). Geophysical and geological evidences suggest this as a working hypothesis on which to direct future work.

Geological setting of the study area

The Midcontinent Rift System

The Midcontinent Rift System (fig. 14) is a Middle Proterozoic U-shaped structure that extends from Kansas through the Lake Superior Region and beneath the Michigan basin. Its extent has been defined on the basis of its geophysical signature; the rocks of the Keweenawan supergroup that were deposited in the rifting episode outcrop only in the Lake Superior Region, everywhere else they are buried by Paleozoic rocks.

The magnetic and gravity anomalies which characterize the extent of this feature are caused by the large volume of mafic igneous rocks, both volcanic and intrusive, associated with the rifting. In the last few years a seismic reflection study, known as GLIMPCE (Great Lakes International Multidisciplinary Program on Crustal Evolution) has added new insights into the geometry of the rift and its mechanism of formation (Cannon et al. 1989).

Some of the characteristics of this rift, like the thickened continental crust beneath it (50 Km) and the presence of layered intrusives are characteristic of other paleorifts present in different parts of the world as described by Morgan and Ramberg (1985).

The volume of volcanic and associated sedimentary rocks deposited in the rift is remarkable; 20 Km of volcanics and interflow sediments plus 10 Km of sediment filling. The

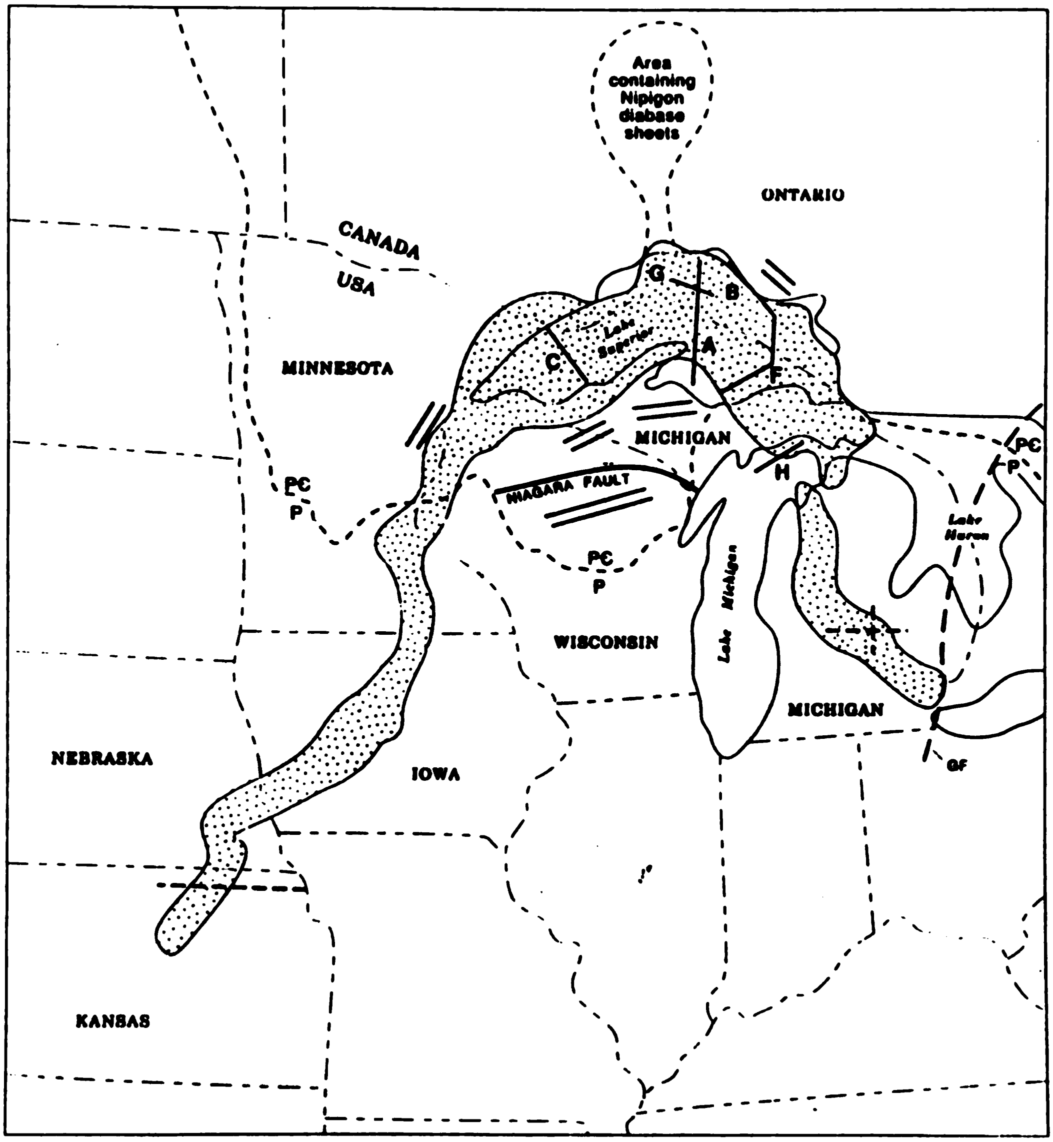


Figure 14. Map showing location of the positive Bouguer gravity anomaly (shaded) produced by rocks of the MRS and some major geological features of the region. Heavy solid lines are GLIMPCE profiles (from Cannon et al., 1989).

volcanics alone can reach 30 Km in thickness! The stretching which reduced the continental crust to 1/4 of its original thickness, almost causing the formation of an ocean floor, is thought to have been accomplished mainly by simple shear extension along a detachment plane in the upper brittle crust (Cannon et al. 1989, Cambray 1988 a and b). However, given the unusual thermal condition of the mantle below this region ductile extension of the crust cannot be ruled out (Cannon et al. 1989). Anomalous temperature in the mantle beneath the rift could have been responsible for ductile stretching at depth and for the amount of volcanics produced. In fact it has recently been calculated by McKenzie and White (1989) that a rise in mantle temperature of only 100-150 degrees C, in association with adiabatic decompression related to a plume, is enough to trigger the production of the large quantities of melt which form the mafic rocks now seen as the volcanic continental margins along the North Atlantic and the Deccan Plateau.

In summary the two outstanding characteristics of the MRS are the overthickened continental crust (about 10 Km thicker than the adjacent stable continental platform) and the large volume of rift fill volcanics possibly related to the presence of a nearby hot spot (Hutchinson et al. 1989). A hot spot in this region was also postulated by Mitchell and Platt (1978) on the basis of the orientation of the dikes around Lake Superior.

These features must be taken into consideration in any study of the structural evolution of the rift.

When considering the structure of the Superior Region of the MRS one cannot help but be surprised by the lack of faults, which are characteristic of most modern rifts. The main feature is the Lake Superior syncline which has very gentle limbs in the northern shore and steeper limbs in the south. There are a few reverse faults with a NE-SW trend (Isle Royal/Douglas and Keweenaw faults) which effect a shortening that is associated with an uplift of 3-4 Km. These reverse faults are thought to have originally been normal faults during the extension phase which switched their sense of movement during the later compressive phase producing a classical example of inversion tectonics.

Many tectonic models have been proposed for the evolution of this rift, some of them will be briefly mentioned in the following pages.

The structure of the rift itself is more complex than one might imagine from a casual examination of the GLIMPCE reflection study. The depocenter of the basin is not central and continuous along the synclinal axis but shifts from one side to the other (see fig. 15) as it passes across some linear zones (NW-SE to N-S trending) separating different sub-basins. These zones appear to be accommodation zones or transfer faults (Cannon and others 1989, Behrendt and others 1988). These features together with the lack of a developed system of faults and a thick continental crust led some

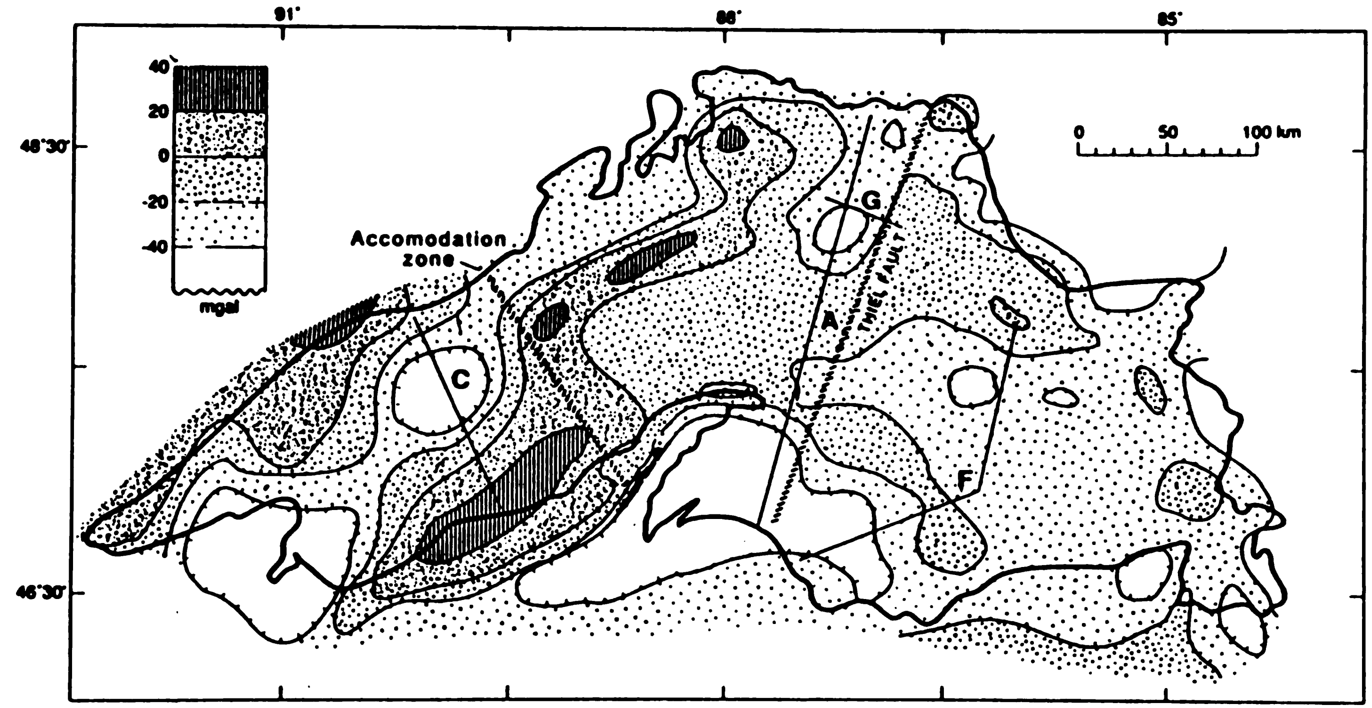


Figure 15. Generalized Bouguer gravity anomaly map of Lake Superior with two proposed locations for accommodation zones along prominent breaks in the gravity pattern of the rift (from Cannon et al., 1989)

authors (Cambray 1988, Cambray and Fujita 1989) to apply models in which the extension is accomplished through a detachment plane (Wernicke 1981, 1985) cutting through the whole crust down to the lithosphere. Given the inhomogeneities arising during the stretching of the crust, the detachment plane can change dip from one side of the rift to the other so that some features called accommodation zones (see fig. 16) or transfer faults are necessary to decouple the movement of the planes dipping in opposite ways (e. g. Rosendahl 1987, Lister and others 1986, Bosworth 1987).

The geometry of the rift requires strike slip motions along the North-South trending arms. The transcurrent motion results in the formation of pull-apart structures during the opening of the Keweenawan basin and in the formation of flower structures, related to transpression, during the inversion of the rift (fig. 17, Cambray 1988).

This work assumes that the detachment hypothesis is valid and leaves aside the problems of rifting driving forces. However, in order to explain some characteristic features detected in the area of study, a brief mention to the tectonic models applied to this rift is necessary.
Tectonic models for the evolution of the Midcontinent Rift System reflect the general question concerning the active or passive role of the mantle in the formation of rifts.

The plume-generated triple-junction model (Burke and Dewey, 1973) implies an active role of the asthenosphere in

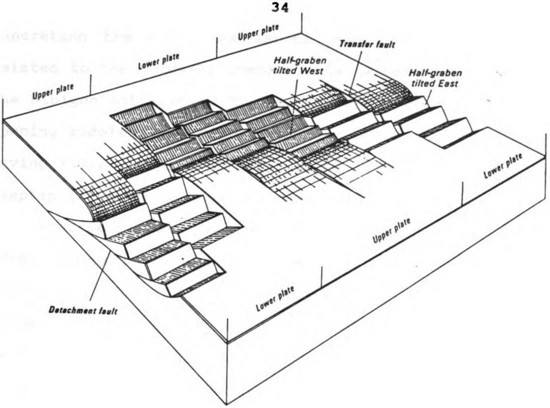


Figure 16. Simple shear model of extension with accommodation zones separating detachment planes dipping in opposite directions (from Lister and others, 1987)

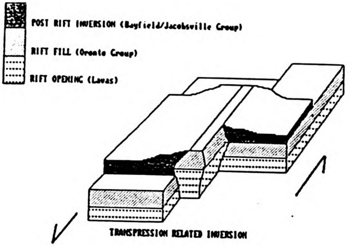


Figure 17. Transpression related inversion causing the formation of flower structures along a N-S trending arm of the MCRS (Kansas) (from Cambray, 1988)

generating the rift; the failed arms are believed to be related to the Coldwell Complex, the Kapuskasing uplift and the Nipigon intrusives. On the other hand collision related opening models (impactogens) such as those of Donaldson and Irving (1972), McWilliams and Dunlop (1978), and Gordon and Hempton (1986) don't need an active role of the mantle.

Precise radioactive age determinations (Paces and Davis 1988) demonstrated that opening and filling of the rift overlaps in part with the Grenville orogeny. However the orogeny extends over a much greater time range. The relations between the orogeny and the Keweenawan rifting are unclear; a possible interpretation has been given recently by Cambray and Fujita (1989) who propose that the reason for the extension is related to a collision induced ripoff. In this model an asymmetric, subducting continental margin is responsible through the process of slab pull for the opening of a rift on the subducting plate once part of the subduction zone becomes locked by the continental impact (fig. 18).

Other models imply large scale crustal growth faults accomplishing the tilting of the strata and allowing the rise of the magma, or explain the subsidence in the basin through a process of dike intrusions over a mantle upwelling, that loaded the lithosphere and caused it to sag.

On the Northwest side of Lake Superior the Pre-Keweenawan Sibley Formation and the intrusive Logan Sills

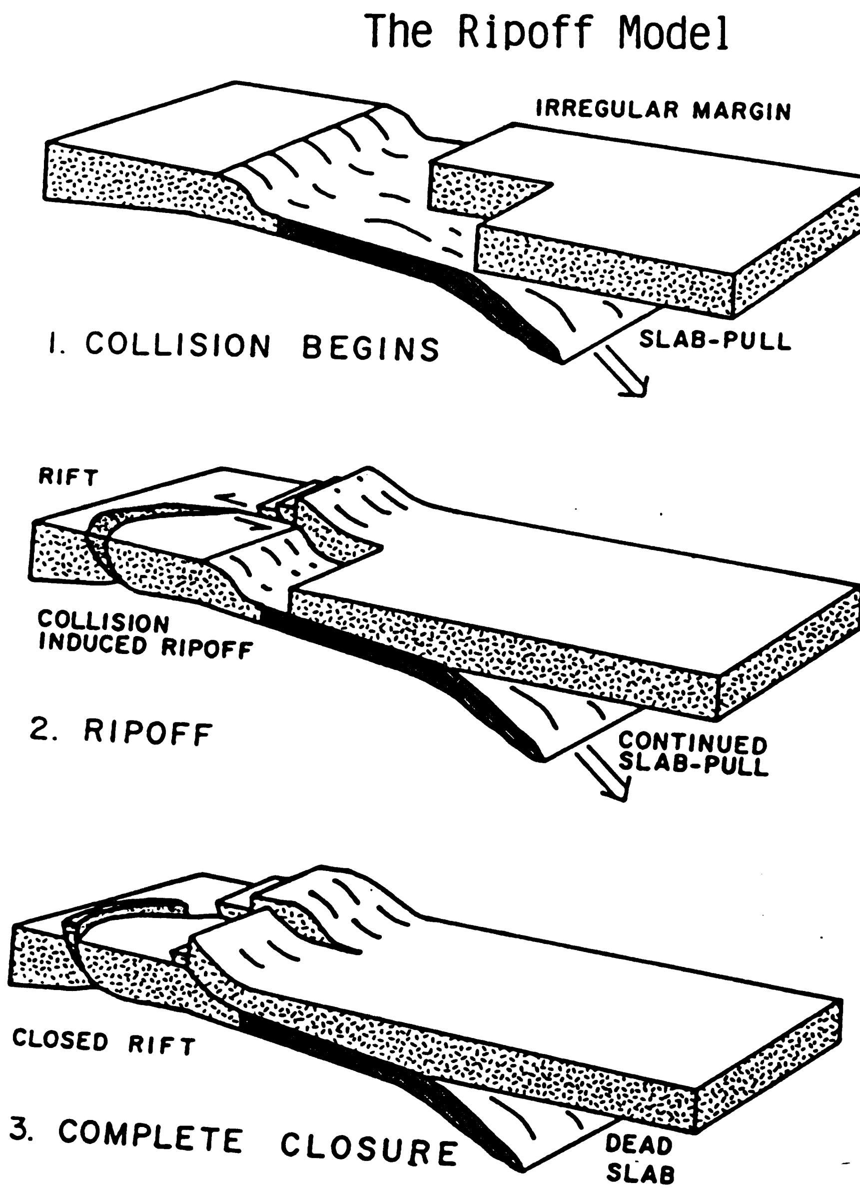


Figure 18. Collision induced ripoff model (from Cambray and Fujita, 1989)

dip gently to the south. There are no large scale faults in these rocks and any deformation in them is in narrow bands following the margins of the sills. It is proposed that this deformation is consistent with the idea of extension being accomplished by a detachment plane cutting through the crust and being deflected along bedding planes which then become shear zones (Cambray 1988 a and b). The existence of bedding plane discontinuities in the sedimentary rocks confines the shear strain. This, and the possibility that at depth the deformation is concentrated along a single discontinuity represented by the detachment plane (the simple shear model adopted in this study), are the reason for the lack of widespread deformation in the form of normal faults. The availability of magma at depth due to anomalous temperatures of the mantle (hot spot) allowed the extrusion of enormous quantities of magma (Hutchinson and others 1989) which followed the discontinuities provided by the detachment plane and the bedding surfaces which had been transformed into shear zones. This is the working hypothesis on which the thesis is based.

A brief description of the stratigraphic record in the Midcontinent Rift System and in Northern Ontario is given below to help in understanding the structural setting.

Stratigraphy of the Midcontinent Rift System

The comparative stratigraphic record for the Keweenawan rift system is represented in figs. 19 and 20. In recent publications authors have tended to call the whole rift fill sequence the "Keweenawan Supergroup".

The formations of the Keweenawan Supergroup are better exposed in Northern Minnesota, Wisconsin and Michigan while in Southern Minnesota, Iowa, Kansas and Nebraska they are mainly known from the well-log records. Except for these two latter states where the formations are very poorly defined, everywhere else a correlation among the depositional sequences of the rift sub-basins is possible to some extent. The first Keweenawan record is represented by the basalts of the rift floor. On top of this we have the Oronto group which is defined along the southern shore of the Lake Superior by three formations; the basal Copper Harbor Conglomerate, the Nonesuch Formation and the Freda Sandstones. Part of the Solor Church Formation of Iowa and Minnesota is equivalent to these formations. The Oronto Group rocks can be considered a thick rift-fill sequence with intercalated minor volcanics (Dickas 1986), the source of sediment coming from the margins of the subsiding basin. A fining upward trend as well as an increase in maturity are characteristics of these clastics; very likely the depositional environment was comparable to a transgressive-regressive alluvial fan (Elmore and Daniels 1980) associated

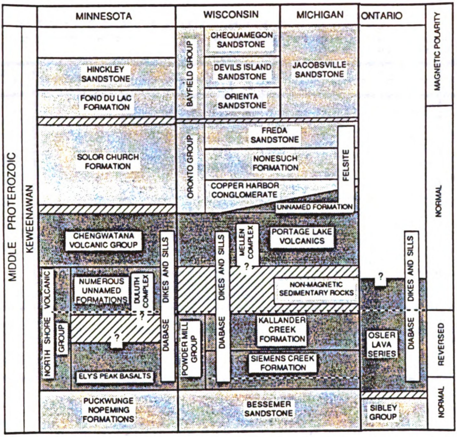


Figure 19. Comparative stratigraphy of the Lake Superior Region (from Weiblen et al., 1989).

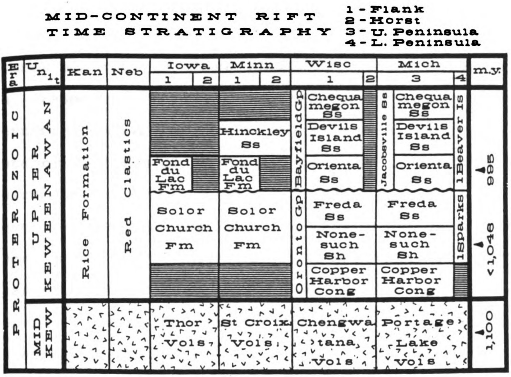


Figure 20. Comparative stratigraphy for the Mid-Continent Rift (from Dickas, 1986).

with synrift fluvial and lacustrine conditions.

Above the Oronto Group we have the sandstones of the Bayfield and Jacobsville Group (Fond du Lac and Hinckley Formation in Iowa and Minnesota). The relatively higher textural maturity of these sandstone represents a puzzle, with respect to their provenance. Sedimentologists alternatively consider them as the product of the erosion of Oronto group rocks or sediment input from weathering of an exposed older granitic terrain.

However the Bayfield and Jacobsville rocks represent a sharp change in the tectonic conditions; according to Morey and Ojakangas (1982) they mark the change from a pure extensional regime to a vertical process of accommodation of faulted blocks.

None of these formations is present along the Northern shore of Lake Superior as it is evident from the comparative stratigraphic column (fig. 19); here in fact we are on the shoulder of the rift and hence the Pre-Keweenawan stratigraphic record is still preserved.

Stratigraphic record of northern Lake Superior

The rocks outcropping on the northern shore of Lake Superior are not as intensely deformed and tectonized as those of the Upper Peninsula of Michigan. The rocks exposed are Proterozoic and they can be assigned to two main tectono stratigraphic units (Franklin, 1982), the early Proterozoic Aphebian Basin and the middle Proterozoic Keweenawan Supergroup (Helikian Midcontinent rift system).

Aphebian Basin

In the northern part of Lake Superior the rocks of the Aphebian Basin are represented by the sedimentary strata of the Animikie group which here have very simple and less differentiated stratigraphic characteristics compared to the corresponding rocks of the southern and central part of the lake that were affected by the Penokean orogeny. The rocks of the Animikie group are best represented in the north western region of the Lake by the Gunflint and Rove Formations:

Gunflint Formation

The Gunflint and Rove Formations outcropping near Thunder Bay represent a transgressive shelf sequence of shale, greywacke and iron rich sedimentary deposits. The average thickness of the 2000 Ma years old Gunflint Formation is 122 meters (Franklin, 1982) and it extends from the Gunflint Lake to NE of Thunder Bay. Its deposition was cyclical. A basal conglomerate member is overlain by two others members composed of chert and tufaceous shale. Goodwin (1956) concluded that the Gunflint Formation was deposited in a shallow euxinic basin with limited circulation to an open sea. The successive foundering of this basin was associated with intense volcanic activity.

Rove Formation

The Rove formation seems to conformably overly the Gunflint Formation, but it is also interpreted as a lateral distal facies of the Gunflint. Its thickness is variable ranging from 380 to 610 meters and more. The formation consists of three lithologic units: a basal black pyritic shale, an interbedded argillite and graywacke (transition sequence) and a quartzitic graywacke. The lower argillite is dominantly exposed in Ontario.

Keweenawan Supergroup in Northern Ontario (Helikian Basin)

The Helikian sequence belongs to the Keweenawan Supergroup which consists of a lower and upper sedimentary section and a middle volcanic sequence. The latter is represented by the Duluth Complex, the Coldwell alkali complex, the Prairie Lake carbonatite and the Logan diabase sill complex.

The lower sequence is represented in Ontario by the quartzose sandstones and carbonate deposits of the Sibley Group equivalent to the Puckwunge Formation and the Bessemer quartzite present in other areas of Lake Superior (NE Minnesota, West Michigan and North Wisconsin). The upper sedimentary sequence is not represented in the area of this study.

The Sibley Group

The Sibley Group is considered to be Pre-Keweenawan in age or very early rifting stage (Franklin, 1982). The Rb-Sr radiometric age is 1300 Ma (Franklin, 1982). This group can be divided into three formations: the Pass Lake at the bottom, the Rossport, and the Kama hill formation at the top.

The Sibley Group was deposited in a N-S trending depositional basin that was deeper to the south. An initial stage of rapid subsidence is marked by the deposition of fanglomerates at the margins of the basin. A series of transgressive (north directed) and regressive episodes controls the deposition of the Pass Lake and Rossport Formations. The top of the Rossport formation with its clay- and dolomite-rich deposits, typical of a constant-depth basin, marks the end of these episodes. The Kama Hill formation records the final stages of filling of the basin in almost subaerial conditions, in fact its depositional setting is that of a mudflat environment. A brief description of the individual formations follows.

Pass Lake Formation

The Pass Lake Formation is composed of a basal conglomerate overlain by quartz arenites. In the south it overlies The Rove Formation and progressively thins to the north coming to rest directly on the Archean Basement. The lithology ranges from conglomerate to quartz arenite generally poorly sorted and with subrounded grains. Most of the quartz grains have a prevalent undulose extinction (>5 degrees). Usually the feldspar is a minor constituent. The matrix consists of fine mica and clay, while the cementing material could be carbonate or chert (Franklin 1982).

Rossport Formation

It overlies the Pass Lake Formation with a discomformable contact in the south, while in the north it steps directly on to the basement. Its characteristics are the brick red color and the high dolomite content. Its thickness is about 100 meters. It could be subdivided, at least in the Southern part of the basin, into three members: lower dolomite, central chert-carbonate (stromatolite to the north) and upper dolomite. The lower and upper members exhibit only minor lateral facies changes in the E-W direction and both members tend to become more clastic to the north (Franklin 1982).

Kama Hill Formation

It represents the final stage of filling of the basin. The type section is right in the area of study (fig. 21) under the powerline. Its maximum thickness is 50 meters and the boundary with the Rossport formation is where the carbonate disappears and there is a distinct change into a deep red purple color. The grains are silt- to clay-sized

Figure 21. Kama hill Formation. Standard section under the powerline at Kama hill.

and the bedding (visible only in thin section) is very thin. Fissility is pronounced. An average content of hematite of about 4% gives the purple color.

In the lower part of the formation (as well as in the Rossport) there are white spherical "reduction" spots. Stromatolitic beds, mud cracks and ripple marks are common features (Franklin and others 1980).

Osler Group, Duluth Complex and Logan Sills

The Osler Group (Ontario) is composed of rocks of the Middle Keweenawan that from their geochemical characteristics, could be considered dominantly continental flood basalts. The basaltic rocks are prevalently olivine-tholeiite with a minor amount of quartz-tholeiite. The Osler Group has a reversed magnetic polarity.

In strict association with the Osler Group is the Duluth Complex extending from Minnesota to Northern Ontario. It is 1100 Ma years old (zircon age, Sutcliffe 1987) and is an order of magnitude larger than the Archean Stillwater Complex. The Duluth Complex (see fig. 22) consists of a series of anorthositic intrusions later intruded by a series of troctolitic Gabbroic and granitoid intrusions; the latter are characteristically layered and occur as separate distinct sheets. The only ultramafic units in the Duluth Complex are limited sequences of feldspathic dunite formed

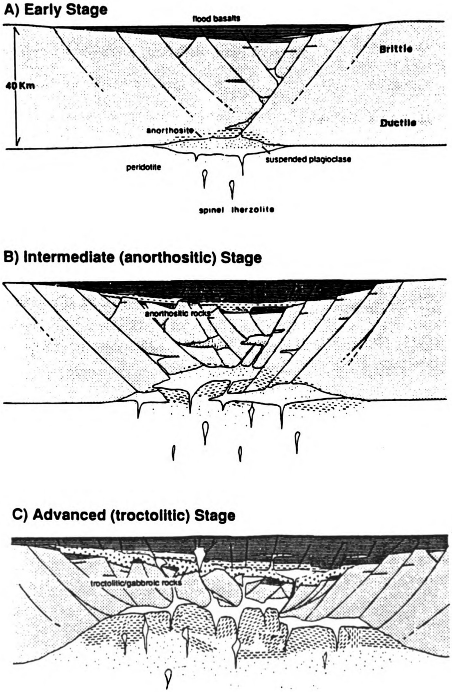


Figure 22. Magmatic development of the Duluth Complex (from Weiblen et al., 1989).

at the base of the troctolitic intrusion. The structural setting of the Duluth complex is fairly simple with a gentle dip of 10-15 degrees to the south east.

The Logan sills occur throughout the Thunder Bay area and northward to Nipigon Lake. The sills at Kama Hill are of the Logan suite. Their thickness ranges from less than one meter to 100 meters. They cut through Aphebian and Helikian rocks at a low angle and gently dip towards the Lake (see fig. 23). The sills are quartz-tholeiite in composition and only the thickest are fractionated. Granophyric zones are common and thought to be connected to assimilation of the country rock.

Figure 23. Gentle dip to the south of the sills along the northern shore of Lake Superior (west view from Kama hill).

DATA ACQUISITION AND ANALYSIS

Data acquisition in Northern Ontario

Field work was carried out during Summer 1989. Useful outcrops were limited to the road cuts or to the shoreline cliffs (see location map in fig. 24) because of the thick vegetation and talus. The most significant outcrops representing contacts between the Logan sills and the Sibley Group Formations are at Kama hill (location 1); most of the work has been concentrated on this particularly well exposed section.

The method of study consisted of recording the relations between shear strain and sill intrusions at any visited locality paying attention both to the macrostructural elements and to mesoscopic scale structures. Samples for microfabric study were collected where their study seemed to be promising. Some of these samples have been used for petrofabric-microfabric analysis; these results are summarized in a later chapter.

The stratigraphic section was measured at Kama hill in order to define the extent and the distribution of the shear strain in that part of the sedimentary sequence.

Southwest of Thunder Bay dikes and sills in the Rove Formation were visited; no exposed contacts between sills and the shales were found. Two miles east of Thunder Bay at

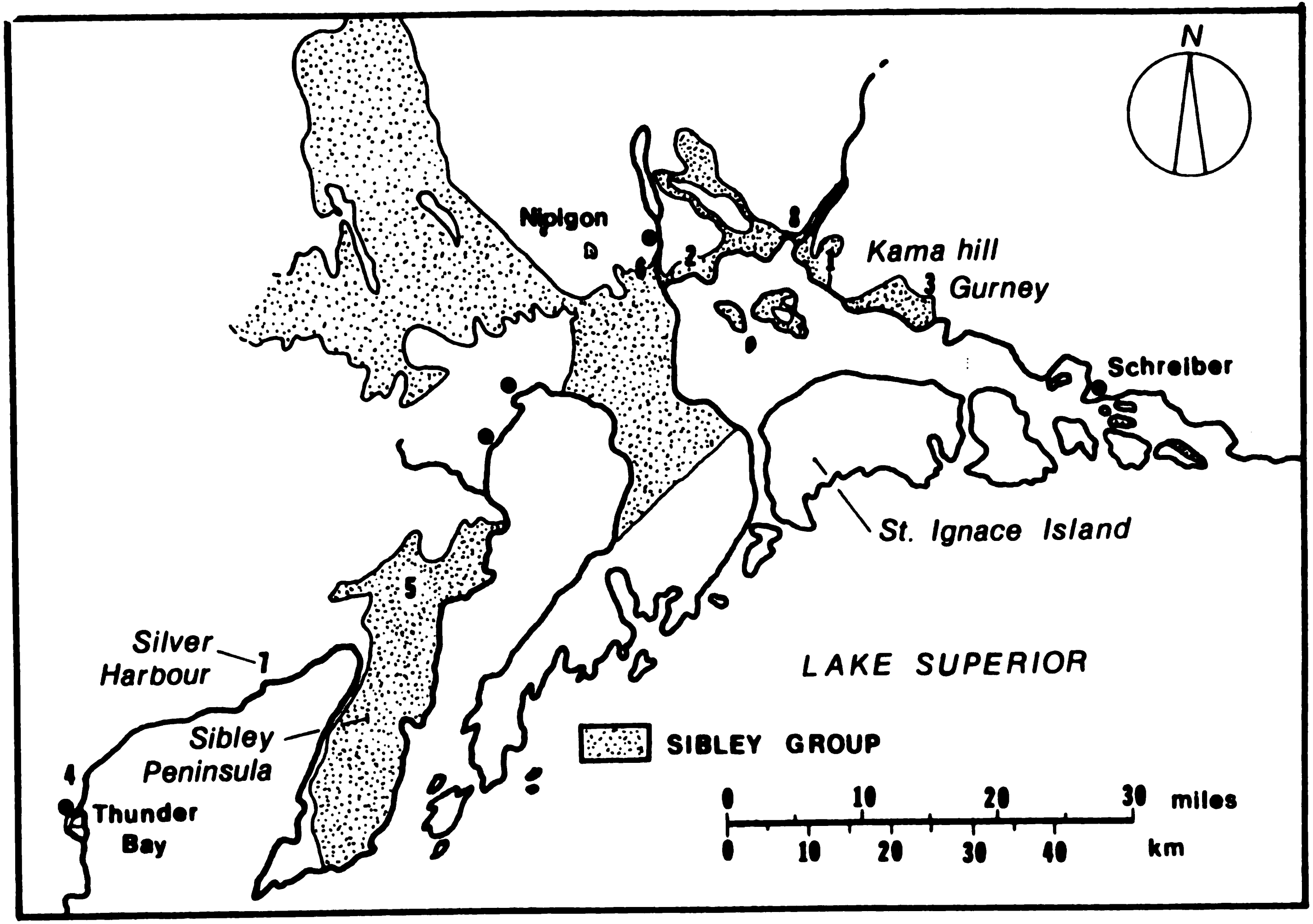


Figure 24. Index map of the locations. 1, Kama hill. 2, Junction highway 11/17. 3, Gurney. 4, Terry Fox lookout. 5, Pass Lake. 6, Nipigon Dike. 7, Silver Harbour. 8, Jackfish River sill.

Terry Fox scenic lookout the contact between a Logan Sill and the black shales of the Rove Formation was examined (location 4). Another similar contact was examined farther east at Silver Harbor (location 7) where the country rock appears to be Gunflint Formation (the shales of the Rove Formation are virtually indistinguishable from the shale members of the Gunflint Formation).

The search for sill country-rock contacts along Highway 11 North was not fruitful given the extent of the vegetation cover; an exposed contact was found between a inclined dike reorienting into the Red cuesta sill a few miles west of Nipigon (location 6). A massive 70 meters-thick sill exposed along the Jackfish River was examined but no contacts were located (location 8).

Other localities (intersection 11/17 (location 2), Gurney, a village west of Schreiber (location 3), and the Sibley Peninsula (location 5)) were visited in search of outcrops of the Sibley Group Formations containing shear strain phenomena, however these outcrops were not in close proximity to sill intrusions.

Kinds of shear strain observed

Several features in the Rossport Formation which have previously been described as sedimentary structures (Franklin 1982) are reinterpreted as phenomena related to shear strain in this study. The criteria used to make this interpretation are discussed in detail in the chapter on shear band analysis.

Two episodes of shear strain can be distinguished in some locations of the study area:

1- Shear strain in the sedimentary sequence probably related to a regional north-south extension along a detachment plane. This is recorded mainly by an organized shear-band pattern striking east-west at Kama Hill, Gurney (west of Schreiber) and in proximity of the intersection between Highway 11 and 17.

2- Shear strain in the sedimentary sequence connected to sill emplacement. This kind of shear was studied in close proximity to the thin sills exposed in the Kama hill outcrop and deforms features formed in the earlier episode.

The regional sense of shear responsible for the extension was constrained as top to the south (south-west) in at least three locations.

Extension was measured locally using the separation of boudins in the shear bands at the intersection 11/17 (fig. 25) and at Kama hill (reference to the photosection 2). In the first locality the extension resulted to be about 19% and 28% in the second one.

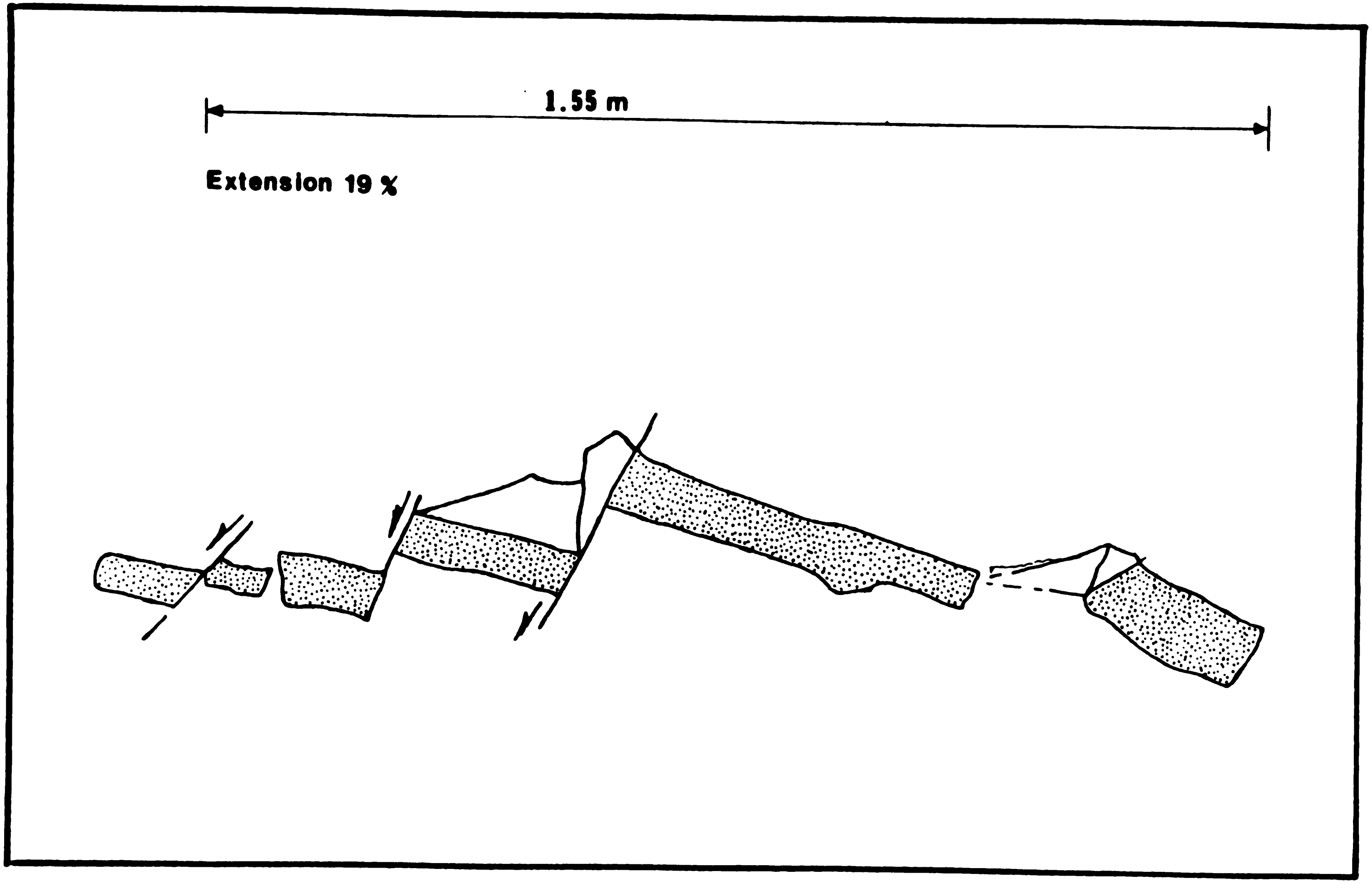


Figure 25. Extension measured on boudins with shear bands at location 2.

Structural relations between sills and shear zones

A- At Kama hill

The bulk of the field work and data analysis presented in this thesis was performed on this extremely well exposed outcrop (location 1). Given the importance of this location a stratigraphic section was measured (fig.26 and Appendix B); and it will be used as a reference when describing the structural setting of the outcrop.

A brief description of the local stratigraphy is necessary to understand the relations between the different structural elements; this is based on the work of Franklin (1982).

Stratigraphic setting of the Kama Hill roadside outcrop:

At Kama Hill the sandy red carbonates of the lowest member of the Rossport Formation outcrop in a broad anticline-like structure.

The characteristically red and white banded formation contains rounded isolated quartz clasts set in a carbonate matrix. The clasts increase in abundance towards both the top and the base of the member (Franklin, 1982); the uppermost part consists of interbedded quartz-arenite and sandy dolomite. The sandstone beds are white in color and are interbedded with sandy red dolostone. On top of these

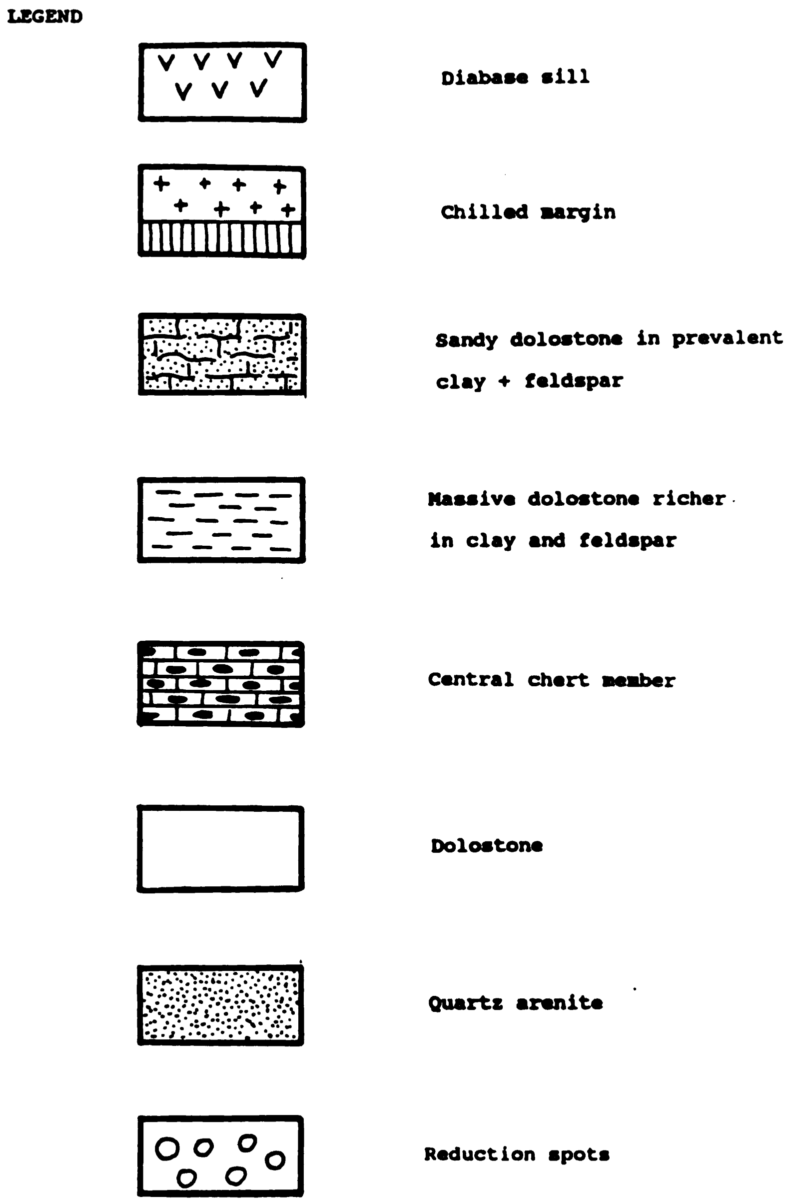


Figure 26. Stratigraphic section and structural relations at Kama hill. Location 1.

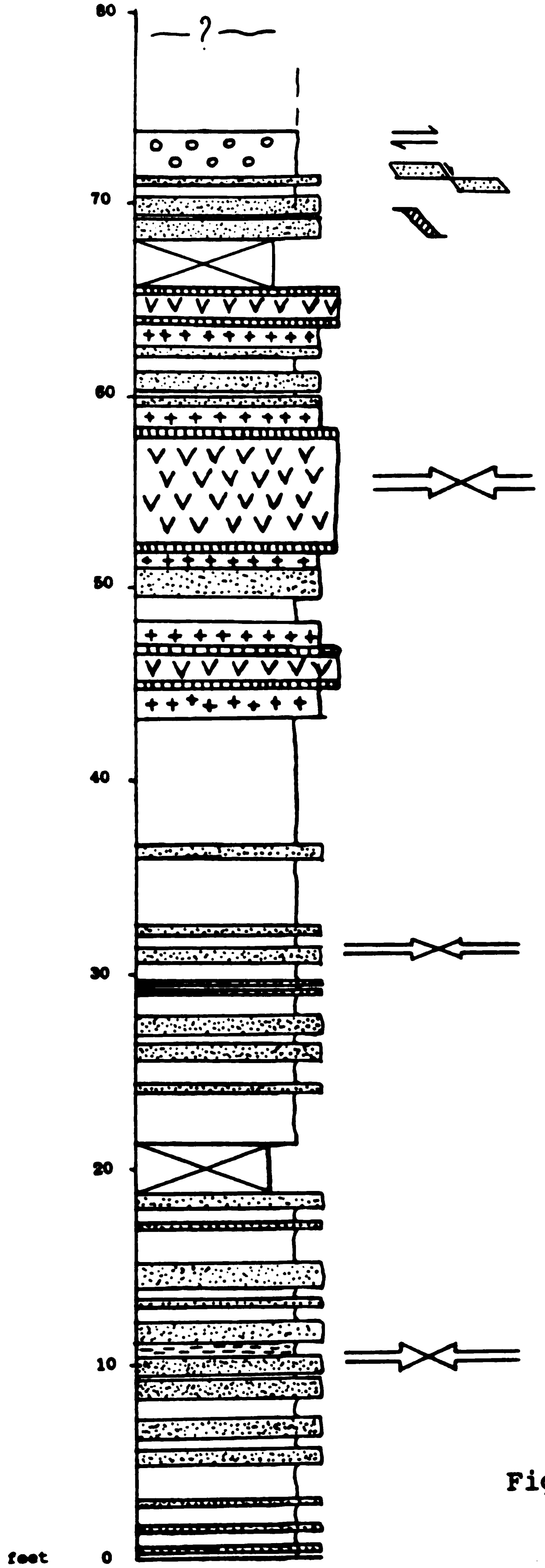
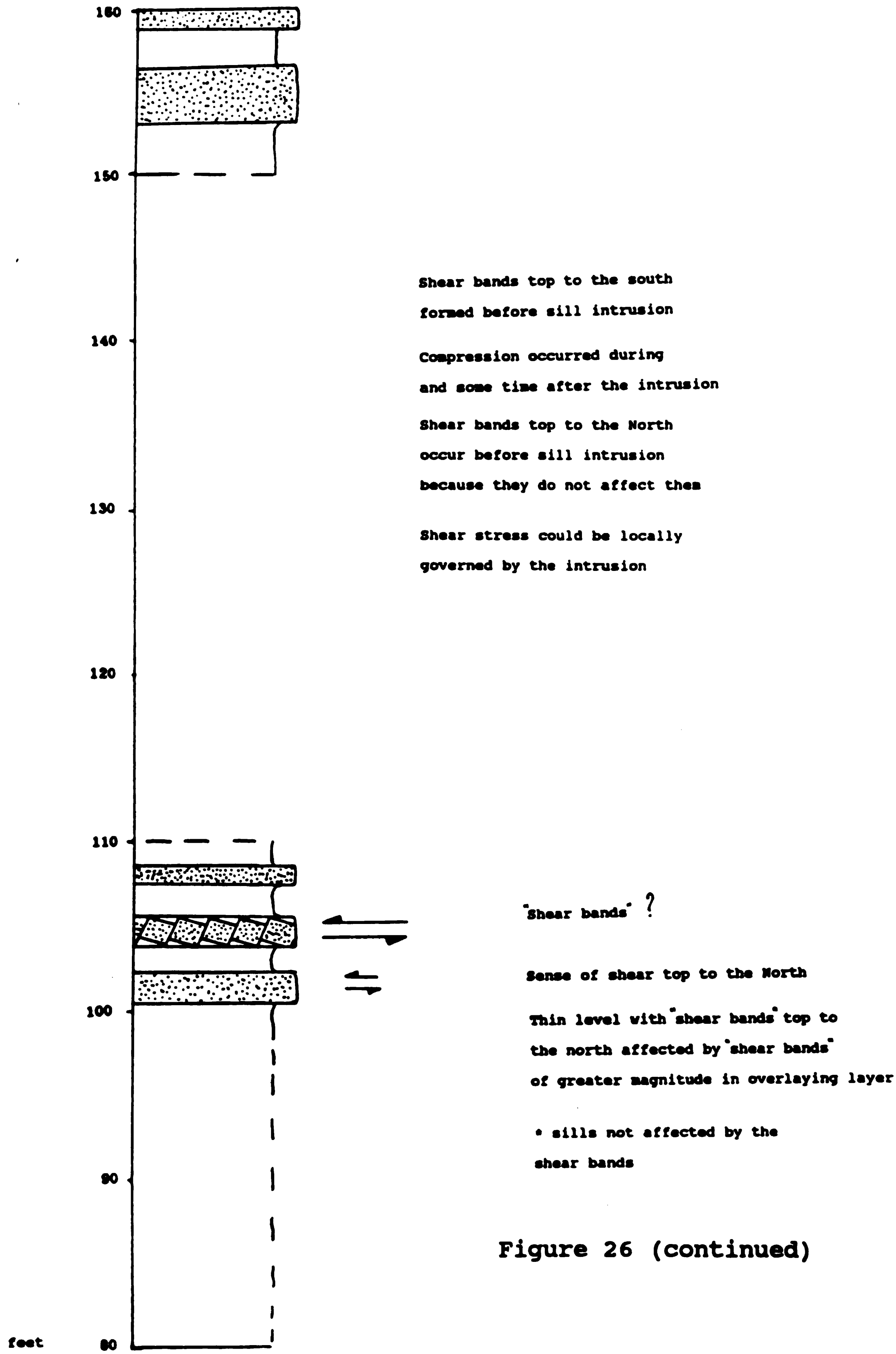


Figure 26 (continued)



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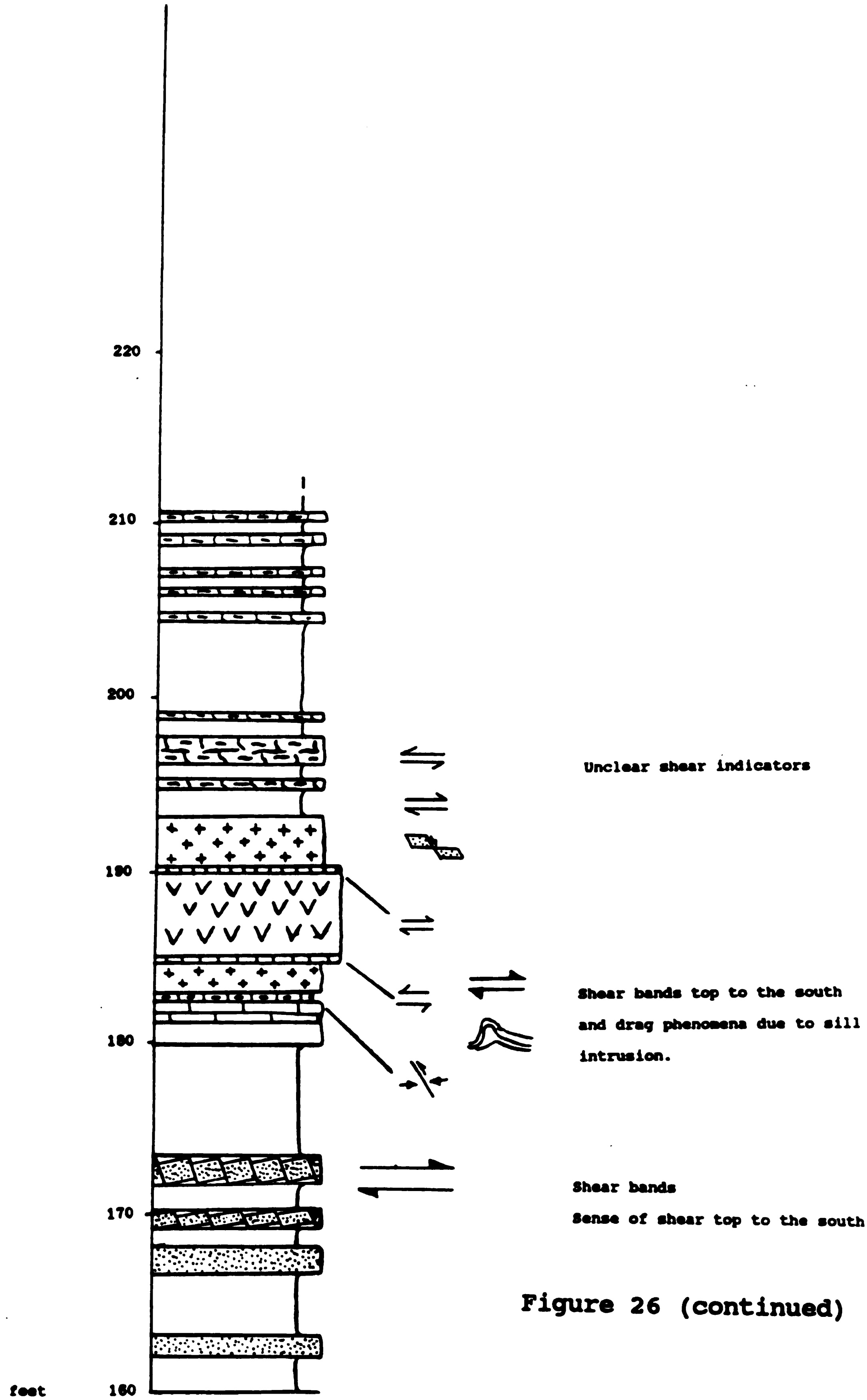


Figure 26 (continued)

layers there are the chert-carbonate deposits of the central member. The latter occur right below a 0.8 meters thick basaltic sill partially replaced by carbonate. Below the lower contact with the sill it is possible to recognize the chert stromatolite unit and a thin level of carbonate (about 15 cm in thickness, anthraxolithic).

Above the central member lies the red dolostone with reduction spots (cream colored, because of the lack of hematite coatings in the grains) representing the upper member of the Rossport Formation. Here the clastic content is minor, the bulk of the rock is made up of authigenic K-feldspar, illite and expandable clay.

Above the road cut on top of the talus slope, the Kama Hill formation is exposed immediately below the huge sill forming the cuesta. It is typically red in color owing to its high hematite content, it is also very fissile.

Kama hill - Lower section (Photosection no. 1)

For the description of the structural elements continuous reference is made to the Photosection no.1 and to the stratigraphic-structural section of fig. 26.

0 - 40 feet. The lower part of the section is affected by compressional phenomena at different scales (from a few centimeters to a few meters). A series of folds and small thrusts disturbs the sedimentary layers; the faults usually have offsets of less than a few tens of centimeters. The vergence is prevalently W-SW for the low angle thrusts and

North for the intermediate to high angle thrusts (fig. 27). The latter may have been early formed extensional features which have been reactivated and now form small backthrusts. The vergence of the folds is to the South-Southwest. The overall structural setting of this part of the outcrop is a broad anticline with an hinge line oriented SE-NW. From 30 feet on up we have a local thickening of the sequence due to drag along a low angle undulating thrust plane, which sometimes seems to be parallel to the dolostone layers; the vergence of the thrust is to the West-Southwest. The throw of the thrust is about 1 meter. A series of imbricate faults on a scale of a few centimeters is present in the sedimentary sequence just below this thrust.
At about 35 feet the first boudins with shear bands striking East-West and sense of shear top to the south, is observed; these shear bands are not widespread and occur in association with small scale folds similar to those previously described.

40 - 80 feet. In this part of the section we have the first appearance of sill intrusions at about 45 feet. Three sills are measured along the stratigraphic section: the first one at 45 feet is 1 foot and 6 inches thick, the second one at 55 feet is 3 feet thick and the third one at 64 feet is about 1 foot and 6 inches thick. Laterally, however, the situation is not so simple because the sills often pinch out and some outcrop for such a short span that they cannot be recorded on the stratigraphic section. The

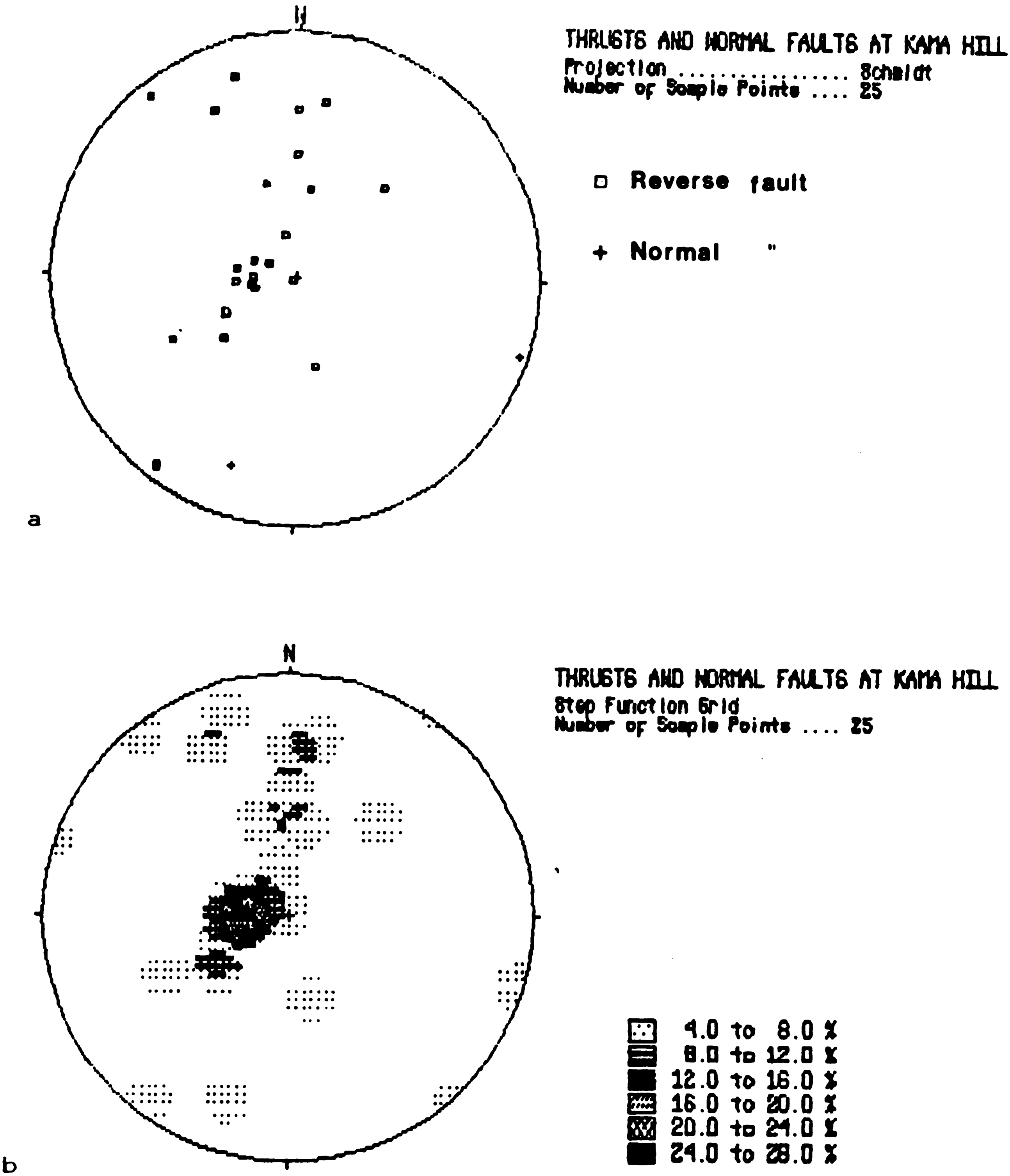


Figure 27. Thrusts and normal faults at Kama hill. (a) Stereoscopic representation of the poles of the planes. (b) Contouring of pole density.

structural setting of the sills is best described on the photosections 1, 2 and 3. The chilled margins of the sills are continuous along their extent and their thickness could be locally variable but tends to have a mean of about 40 centimeters.

At about 55 feet there is a thrust-ramp in the layers of the Rossport Formation. The vergence of the thrust varies, because of the undulose plane, from SW to W. The sill occupies the space created by the bending of the strata as it ramps from one layer to another and the chilled margin is continuous which would seem to suggest that the sill intruded during the thrusting (diagram in photosection 1). The thrusting continued for a little while after the sill intrusion, without causing their offset, but forming slickensides which were observed in the sample S 10. The slickensides indicate a sense of shear top to the SW. The sill IIb is affected by a small NW-SE trending normal fault with an offset of a few centimeters; this is one of the very few places where a sill is affected by a fault, probably the result of differential compaction.

At 70 feet there is a quartz arenite bed which is broken into a series of rotated boudins with E-W axes which have been interpreted as shear bands indicating a sense of shear top to the south. There are some shear bands with an ambiguous orientation. About 20 centimeters beneath this level there are extensional veins in the dolostone which indicate shear with a top to the south sense.

In the uppermost part of this section a sill steps up to the north through the sedimentary strata at a 30 degree angle. Boudins in the sediments below this sill have been affected by the intrusion of the sill. Some of the boudins have been shortened and stacked with a sense of movement opposite to that deduced for their formation (fig. 56). This demonstrates that shear strain was active before the sill intrusion. Completely spherical reduction spots are present in the dolostone suggesting that the shear strain was very localized.

Kama hill - Railroad section (Photosection no. 2)

In the section along the road from 80 feet up to about 150 feet there is a thick cover of vegetation and the bedrock is not exposed. However, below the highway by the lake shore there is an exposed section along the Canadian Pacific Railroad that represents in part the section which is hidden on the highway.

Two sills (probably equivalents of those outcropping at 58 feet and 65 feet in the lower section) are exposed at the base of the cut. The upper sill is about 30 centimeters thick and pinches out twice. The lower sill is more than 50 centimeters thick and appears to be continuous.

Above the sills in a stratigraphic level that, on the basis of the geometrical relations and sill correlation, is tentatively placed between 100 and 110 feet, the quartz arenite layers are affected by E-W to ENE-WSW striking normal faults. These structures occur on two different scales; one with offsets of a few centimeters and the other with offsets ranging from several decimeters to one meter. Thin quartz arenite layers containing the smaller faults have been affected by drag along larger scale faults (fig. 28 a and b) and seem to have undergone horizontal stratal shortening. The larger faults appear to have formed after the small-scale ones; for both, if one interprets them as shear bands, the sense of shear is opposite to that inferred by the kinematic indicators in the lower section.

At the southern end of the railroad outcrop small-scale faults with an antithetic sense of movement were found in association with the north dipping normal faults (fig. 29). The sills are never cut or affected by the faults, shear bands or related drag folds.

Kama hill - Upper section (Photosection no. 3)

The upper part of the section is exposed at the southern end of Kama hill. From geometrical considerations the first exposed layers are about at 150 feet from the bottom of the stratigraphic section.

150 - 180 feet. The most remarkable features in this interval are two levels with shear bands; one at 170 feet and the other at 172 feet; the second one is a key horizon. In the upper level the shear bands strike east-west and are consistent with a sense of shear top to the south; in the southernmost part of the exposure of this level the shear

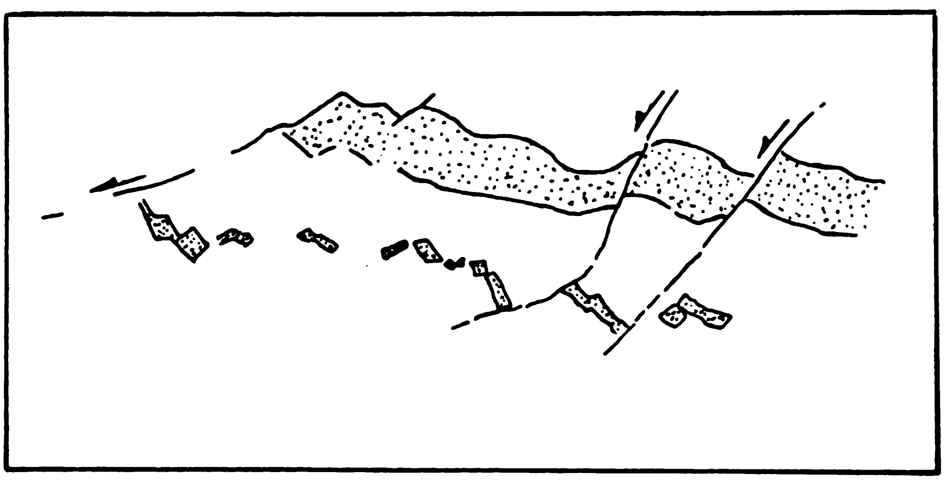


Figure 28. Normal faults (Shear bands?) along the Railroad section (Kama hill). (a) The reddish dolostone layers seem to be more ductile and the fault plane is deflected as it enters one of these beds. (b) Reverse drag of smaller boudins in connection with the movement of a large normal fault.

a

b

Figure 28

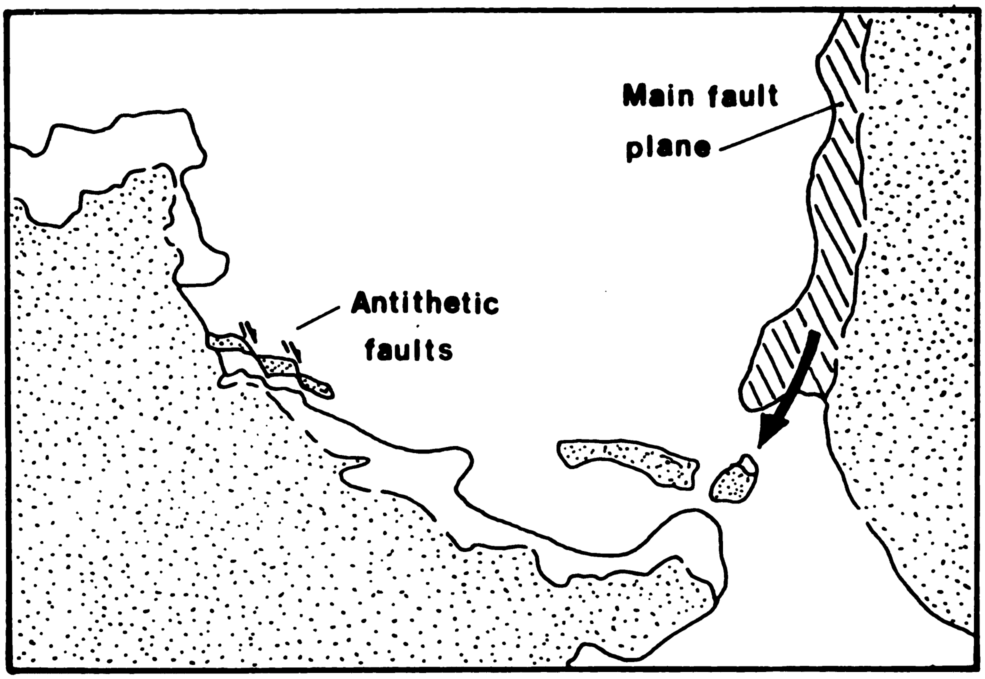


Figure 29. Antithetic normal faults (shear bands?) along the railroad at Kama hill. The arrow denotes the sense of motion along the fault plane.

Figure 29

bands rotate to strike NNE-SSW. The extension measured on this layer has yielded a value of 28%. Above the shear bands, in the red dolostone, two sets of calcite veins are recognized; it appears that the calcite has filled these veins growing perpendicularly to the walls.

180 - 200 feet. Starting from 180 feet beds of the central chert-stromatolitic member of the Rossport formation can be recognized. At 181 feet there is a thin carbonate-stromatolitic horizon and on top of this, banded chert. In the chert a system of microfaults (fig. 30) with an alternate normal-reverse sense of displacement is present. They are particularly well developed close to the sill intrusion. A series of drag folds with an overall vergence to the north (fig. 31) is present both in the chert and in a lithologically indistinguishable layer in the sill aureole. The stratal shortening of the chert and the drag folds seem to be consistent with a northward directed sense of emplacement for the magma. A layer of shear bands with east-west striking boudins and original sense of shear top-to-the-south is recognizable right at the lower contact of the sill; at location 40, some boudins have been rotated and shortened by the shear stress connected with the northward directed magma emplacement. A complete description of this latter phenomena is given in the chapter about shear stress modeling.

The sill at 187 feet has a continuous chilled margin everywhere and locally (location 34) pinches out; a model

Figure 30. Microfaults in the chert member of the Rossport Formation at Kama hill.

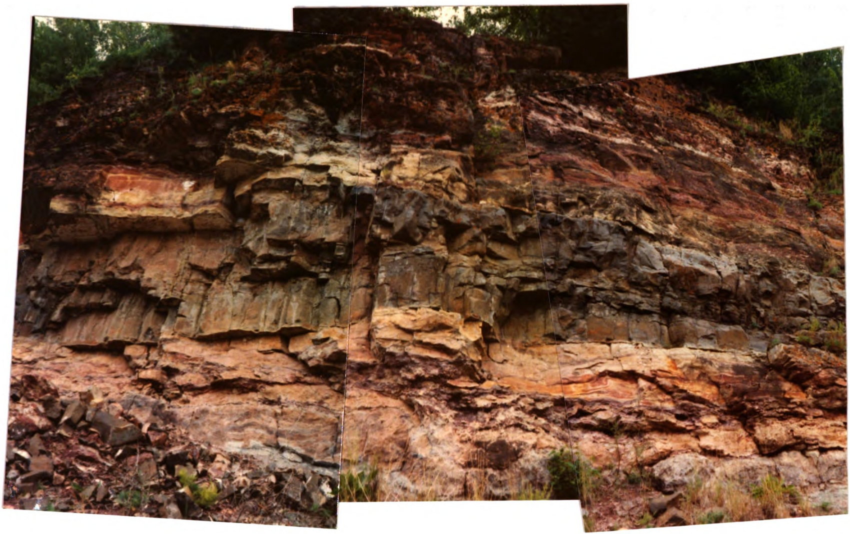


Figure 31. Drag folds in the chert member of the Rossport Formation at Kama hill.

that accounts for local pinch out of the sills without considering different directions of magma emplacement is sketched on the photosection 3. A laterally-not-continuous sedimentary layer is trapped by the sill following a parting surface (fig. 32) giving an impression of lateral termination for the intrusive sheet in cross section; this model also explains the apparent lack of macroscopic interference phenomena at two close sill terminations such as described by Pollard (1973). Right on top of the sill in the aureole there are a few drag folds consistent with a bottom to the north sense of shear as one would predict for these sills. Above the aureole the kinematic indicators are extremely ambiguous; at places there are east-west striking shear bands with sense of shear top to the south, that appear to turn into opposite sense of shear indicators just a few feet higher in the section. However, the overall sense of shear seems still to be top to the south. It must be mentioned that in this part of the section the lithology changes, and the layers become very rich in clay and poor in quartz arenite. Boudins appear to be cream-colored like the reduction spots in the lower part of the sequence, are not well defined and seem to be more stretched with respect to those outcropping below the sill.

> 200 feet. The upper member of the Rossport Formation, the Kama Hill Formation outcrops under the powerline, on top of that there is a 100 hundred meters thick massive diabase sill (fig. 33). Talus and vegetation cover prevents direct

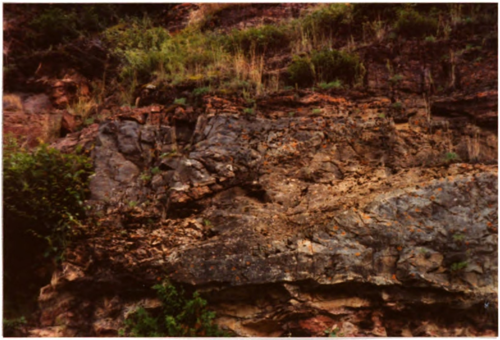


Figure 32. Sills pinch out. Chilled margin all around the intrusions. Some small drag folds in the sedimentary layers are visible (Kama hill).

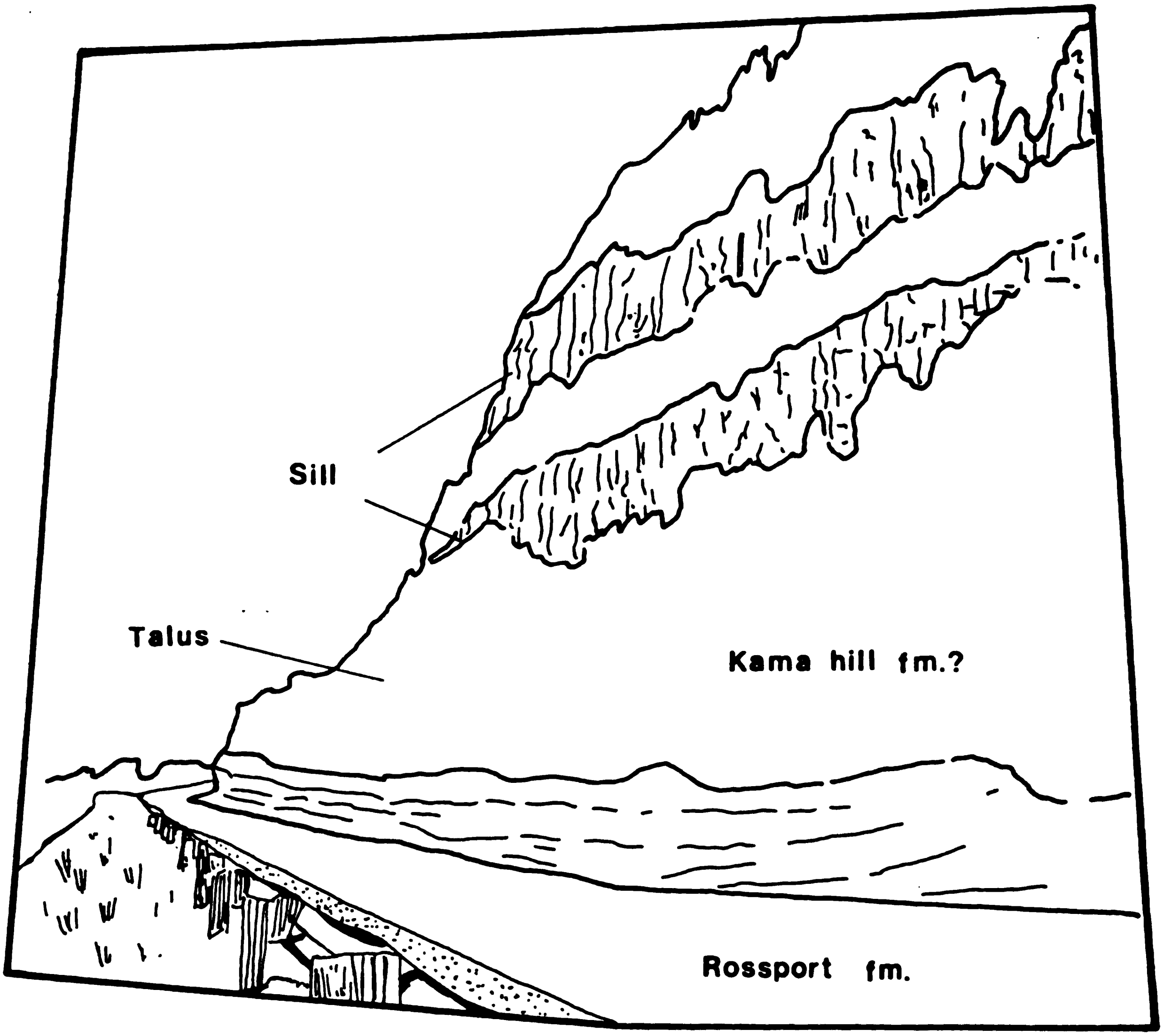
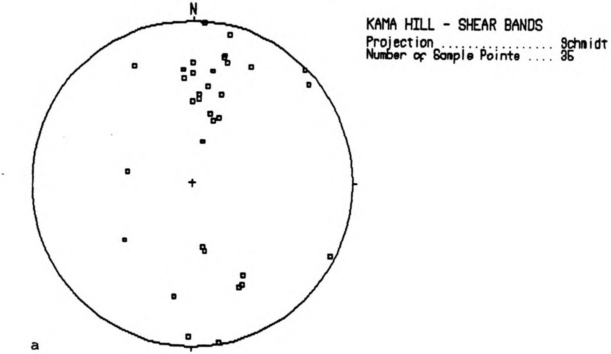


Figure 33. Southern outcrop at Kama hill. The thick sill on top intruded above the Kama hill formation.

Figure 33

observation of the contacts.

In summary, the relations between the different structures and the intrusions observed at Kama hill are as follows. The east-west striking shear bands connected with a top to the south movement (stereonet representation in fig. 34 a and b) were the first structures to form; extensional tension gashes are also consistent with this sense of shear. This was followed by sill intrusions almost concurrently with the compressional phenomena in the lower part of the section (stereonet representation in fig. 27 a and b). The vergence of the thrusts is to the W-SW. The intrusion of the sills created a certain amount of shear stress at the wall with the country rock (stratal shortening, anticlockwise rotation of boudins, drag folds, etc.); the shear strain observed at the wall and particularly the drag folds (stereonet representation fig. 35 a and b) is consistent with a northward-directed trend of emplacement of the magma. Sometimes the sills have intruded along layers in which boudinage and shear bands were previously developed. The sills are usually unaffected by the geological structures except for the thrust-ramp which seems to have driven their emplacement. Furthermore a few slickensides consistent with the sense of movement of the thrust have been observed in the sill below the ramp. The sills tend to follow the bedding during their emplacement (fig. 36 and 37).



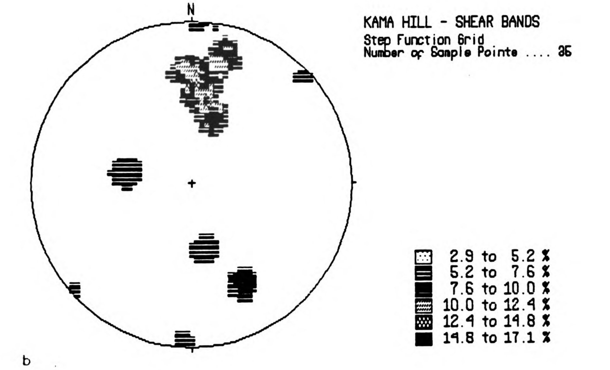
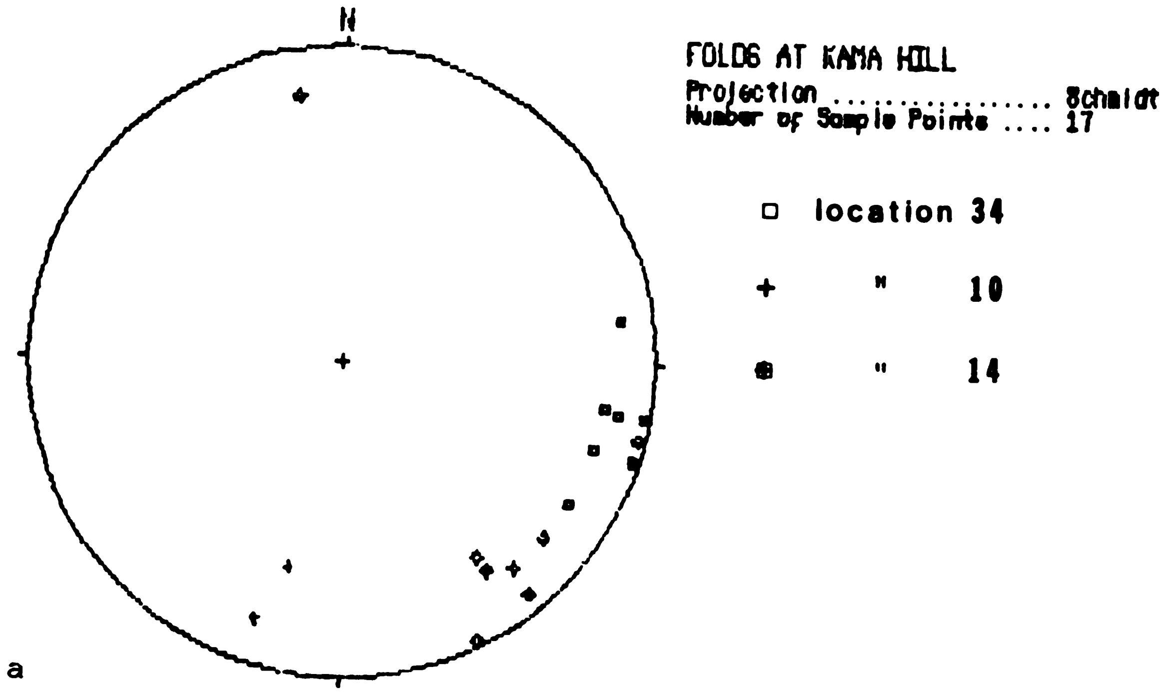


Figure 34. Orientation of the shear bands at Kama hill. (a) Stereoscopic representation of the poles. (b) Contouring of the density of the poles.



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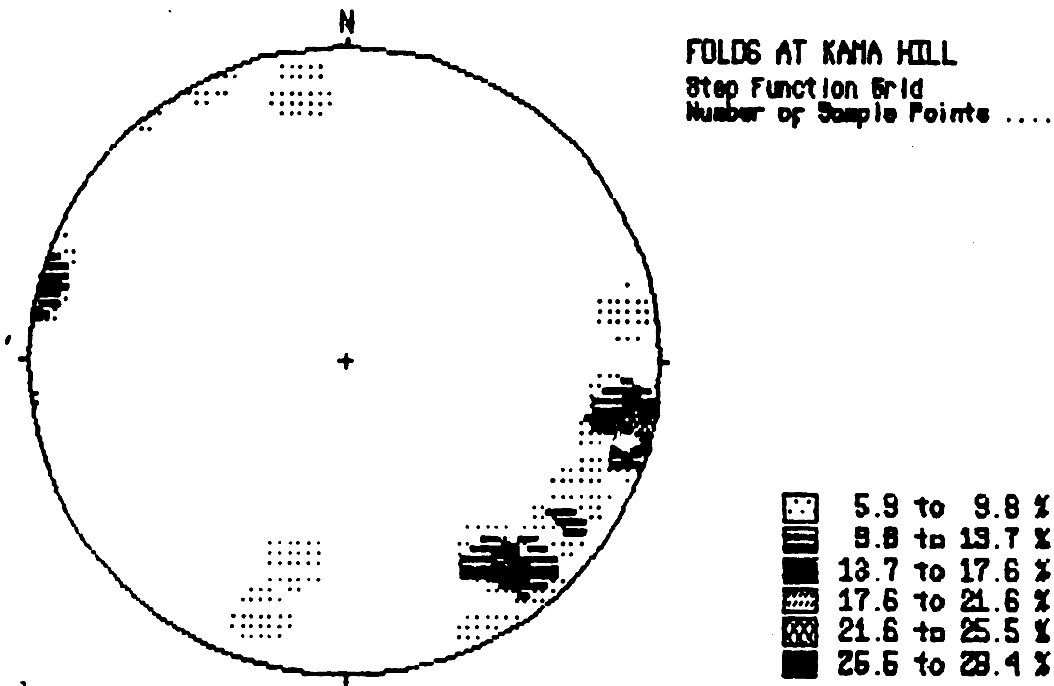


b

FIgure 35. Hinge lines of folds at Kama hill. (a) Stereoscopic representation. The different symbols represent measurements in different locations. (b) Density contouring.

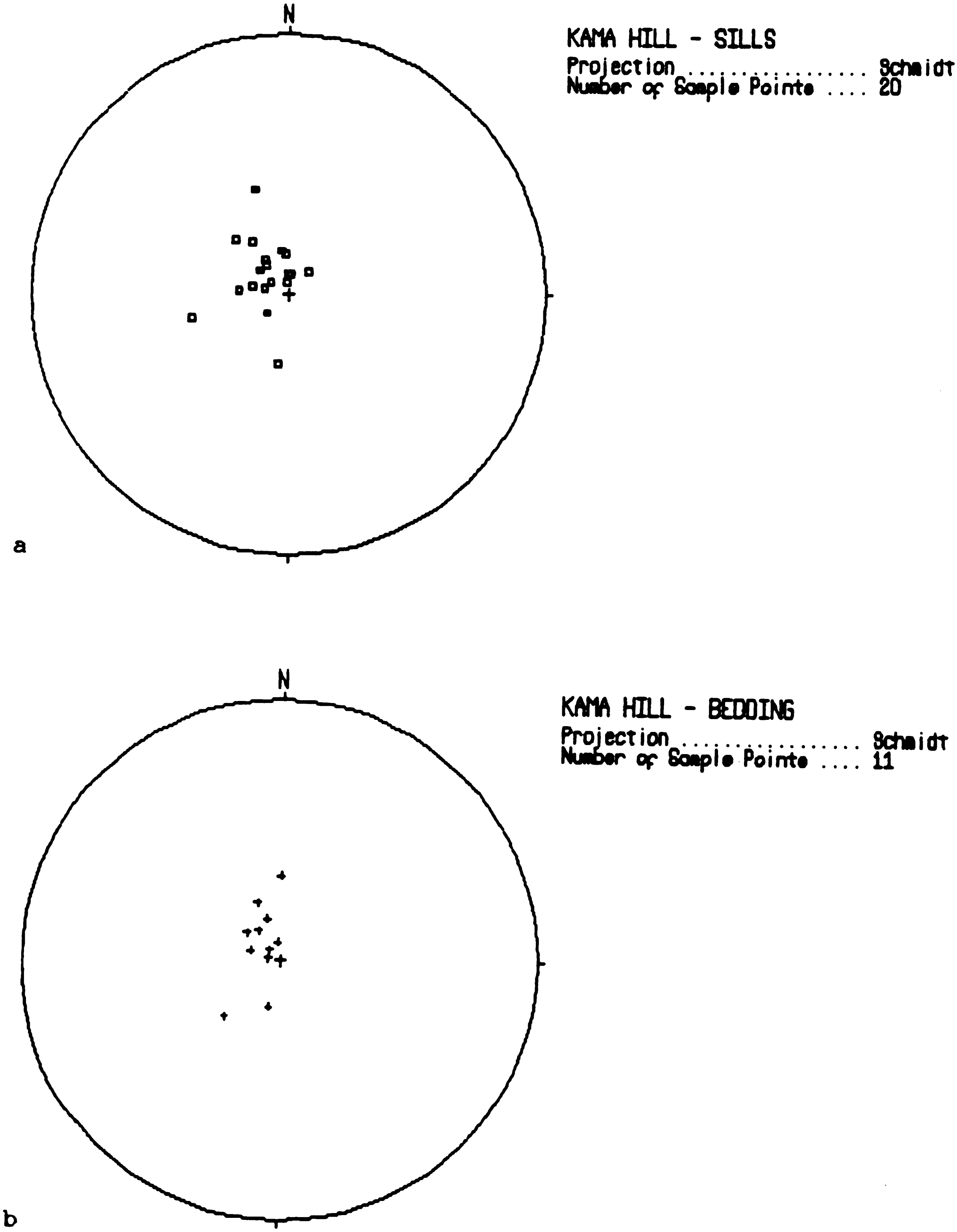
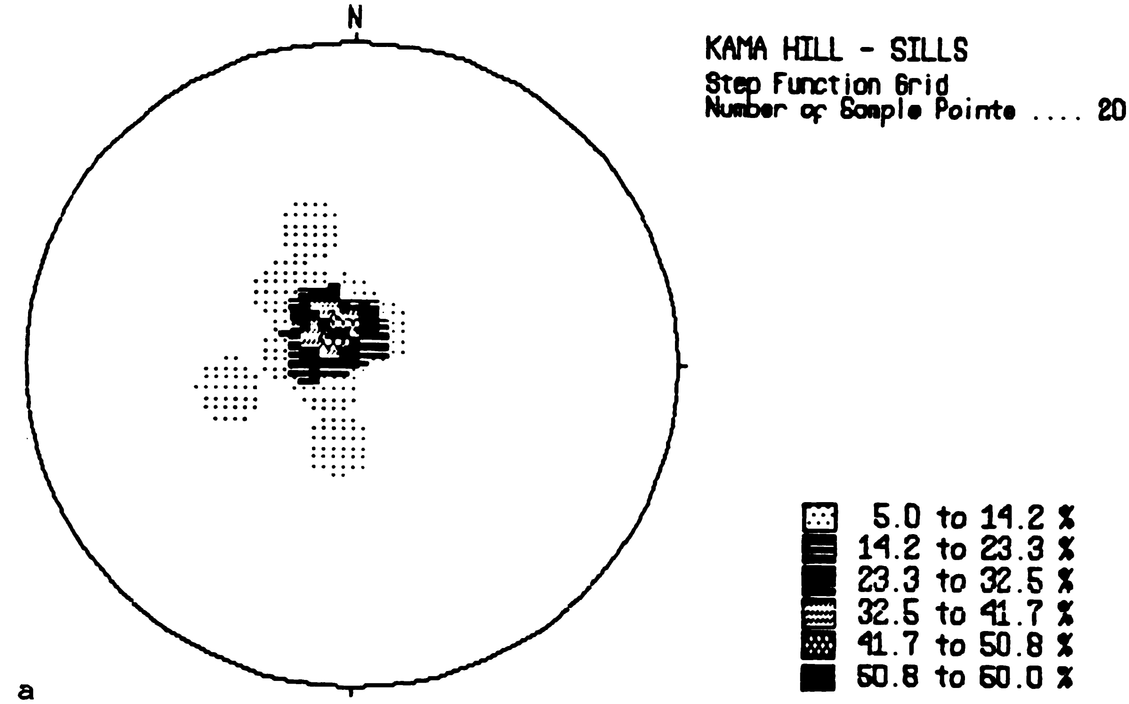


Figure 36. (a) Poles of the sills. (b) Poles of bedding (Kama hill).



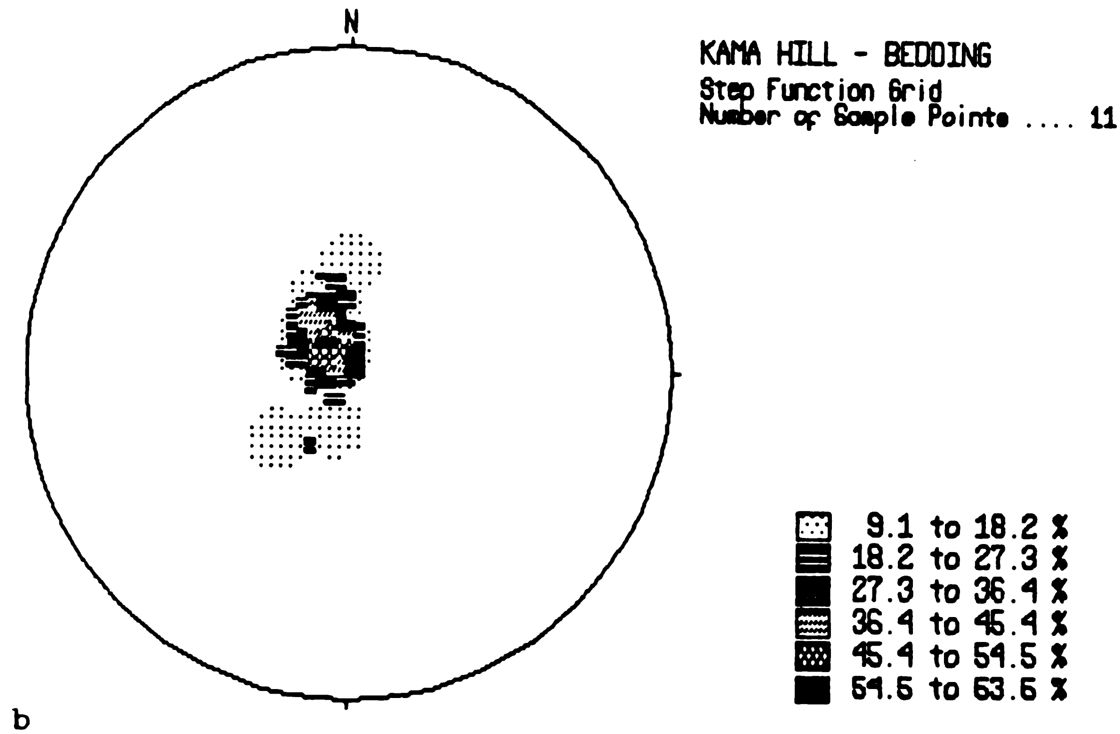


Figure 37. (a) Density contouring of poles of sills. (b) density contouring of poles of bedding.

B - Other outcrops

Evidence from other outcrops visited during the field work will now be reviewed; because of the lithologies exposed and the outcrop conditions, the amount of data is far from being comparable to that collected at Kama hill. The first two locations described are affected by shear strain phenomena probably connected to the regional extension.

Location 3 - West of Gurney (Schreiber). Roadcut in the Rossport Formation, where 30-40 cm thick quartz arenite layers are offset by NW-SE trending shear bands with a top to the SW sense of shear. Attitude of the strata: 40/20.

Location 2 - East of the Intersection 11/17 (5.8 miles after the junction 11/17 in direction Sault St. Marie). Roadcut in the Rossport Formation. The quartz arenite layers are intersected by E-W to WNW-ESE striking shear bands with sense of motion top to the south. The attitude of the strata varies from 80/13 to 130/16. An extension of 19% has been recorded (fig. 25) from the geometrical relations among the boudins of a quartz arenite layer.

Location 6 - Nipigon dike (2 miles south of Nipigon). A

25/40 trending dike steps up through the gneissic rocks of the basement and flattens into the Red Cuesta sill as it enters the sedimentary sequence (fig. 39).

Location 7 - Silver Harbor. Diabase sill in lower Gunflint Formation exposed along the Lake shore. The sill is gently dipping (4 degrees) to the south. The Gunflint Formation is layered but no shear phenomena were observed.

Location 4 - Terry Fox Scenic Lookout (2 miles north of Thunder Bay). Diabase sill in rocks of the Animikie Group. Some erosional features (scour marks?) seem to be present at the bottom of the sill. No clear deformation phenomena have been recognized in the massive sedimentary rocks. Another intrusive body pinches out below the sill, it can be distinguished from the country rock because it is unstratified and by the network of leucocratic veins that characterize its extent. The sill is gently dipping to the north.

Other locations have been visited, but given the poor exposures or the scarcity of structures they will not be reviewed here. The attitudes of the sills on which it was possible to make a precise measurement have been summarized in a stereonet (fig. 38).

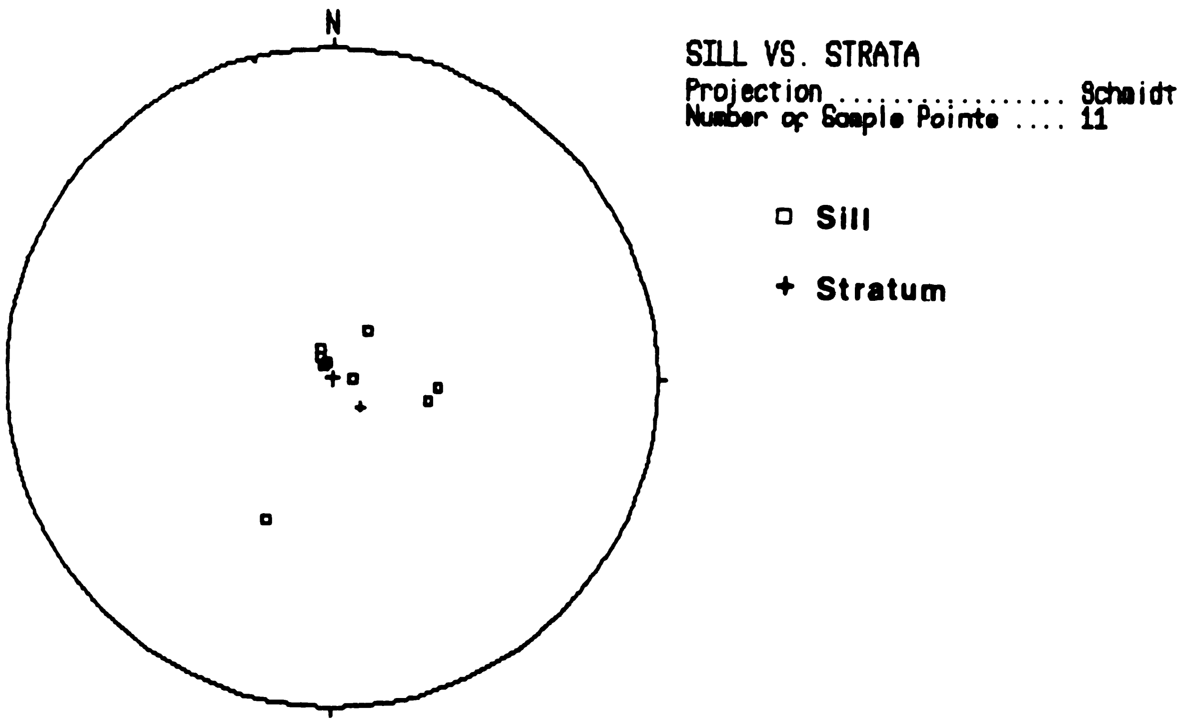


Figure 38. Poles of sills versus poles of strata at Kama hill.

Figure 39. Nipigon dike.

DATA ELABORATION AND ANALYSIS

In the following chapters the structural data collected in the field is analyzed with special reference to shear bands and to the petrographic-microfabric study performed on some selected samples. Furthermore a physical model to quantify the shear stress at the wall of a sill intrusion is developed.

The aim of the analysis presented is to test some assumptions that are critical for the theoretical interpretation of the geological structures observed in the field and demonstrate kinematic indicators in the microfabric of the rocks.

Shear band analysis

Some tests were carried out on the data collected from the small normal faults present in the Rossport formation of the Sibley group. This was done to determine whether they are the result of synsedimentary soft sediment deformation (as currently believed; see Franklin 1982) or of shear strain connected to the opening of the Keweenawan rift system.

First of all in the case of a synsedimentary deformation there should be a pattern something similar to the geometry of a growth fault, particularly the thickening of strata across the hanging wall in the direction of the footwall. The massive nature of the quartz arenite and the dolostone prevented this investigation. A different approach was thus used. The data were compared with features typical of a shear zone. In particular the geometry of the shear bands, also called extensional crenulation cleavage (Platt and Vissers 1980, Platt 1984), was compared to the available data. It has also been demonstrated that joints, which acted as pre-existing weaknesses were present and reactivated at the time of shearing. These latter are crucial for the proof of a non-synsedimentary origin of the faults.

According to Platt and Vissers (1980), shear bands (extensional crenulation cleavage) occur in a shear zone and facilitate extension with the same significance as Riedel shears have in a strike slip fault zone (fig. 40). In particular two sets of extensional crenulation cleavage will develop (fig. 41), one at low angle to the shear zone (ecc1 or low angle Riedel shears) and one at high angle (ecc2 or high angle Riedel shears); usually the more active and developed set is the ecc1, while the ecc2 is most of the times subordinate.

Platt and Vissers (1980) have also showed that in the case of shear strain, where we have a non-coaxial deformation, it is expected that the ecc2 rotate quickly towards the axis of infinitesimal extension and rapidly become inactive, while the ecc1 will rotate very slowly or won't rotate at all, and can remain active indefinitely,

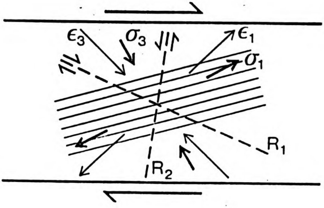


Figure 40. Orientation of deviatoric stress axes (σ_1, σ_3) generated by transpression along the shear zone, instantaneous strain axes (ϵ_1, ϵ_3), and potential secondary fracture orientations in a fault zone (from Platt and Leggett, 1986).

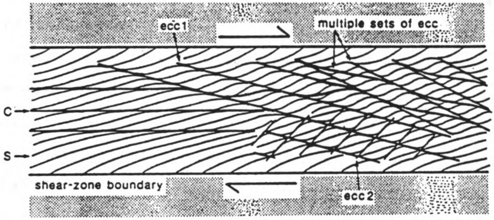


Figure 41. Diagram to illustrate the orientations and mutual relationships of foliation in shear-zones. S, shape fabric; C, shear bands. ecc1 and ecc2, conjugate sets of extensional crenulation cleavages (from Platt and Visser, 1980).

accomplishing most of the deformation. If we examine the situation on the Kama hill roadside outcrop, ecc1 shears should dip south and progressively rotate in an anticlockwise direction for a top-down-to-the-south sense of movement (fig. 42). As the shear continued and the rotation increased the displacement along the shear should also increase.

The angle of the shear band with the shear zone (alpha in fig. 43) versus the displacement was plotted; and as can be seen in fig. 43 the expectations were confirmed by the alignment of the points on a line with a negative slope showing an increase in displacement as the angle alpha decreases. An interesting cluster of points denoting an high alpha angle, with a very small displacement may represent a joint system reactivated by the shear; the high angle joints find themselves in the clockwise quadrant and are quickly rotated towards the extensional axis becoming inactive (because of their orientation we cannot refer to these as ecc2).

A similar result was obtained by plotting alpha versus beta (fig. 44), the angle between the bedding plane and the shear band. Here the situation is more compressed, but a cluster of points with high alpha and high beta is interpreted as a reactivated joint system. The other points align themselves on a line with a negative slope showing an increase in beta with a decrease in alpha. In particular the angle beta for the joints was very close to 90 degrees.

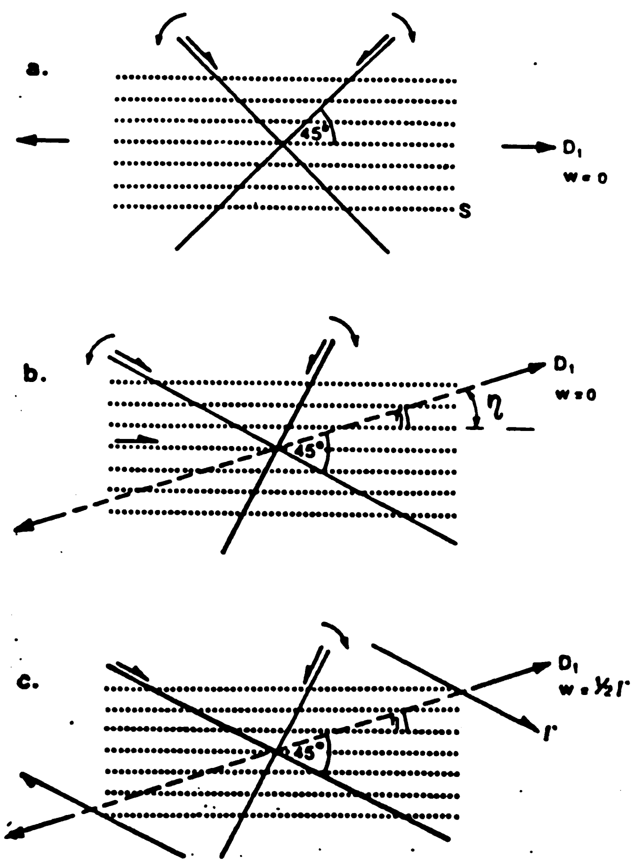


Figure 42. Orientation of shear bands. (a) Coaxial deformation. (b) Coaxial deformation with D1 at an angle η to foliation. (c) Simple shear at rate Γ at ($45-\eta$) to foliation. Shear bands with thicker lines are dominant. The sense of rotation of the shear-bands is shown by the curved arrows (from Platt and Visser, 1980).

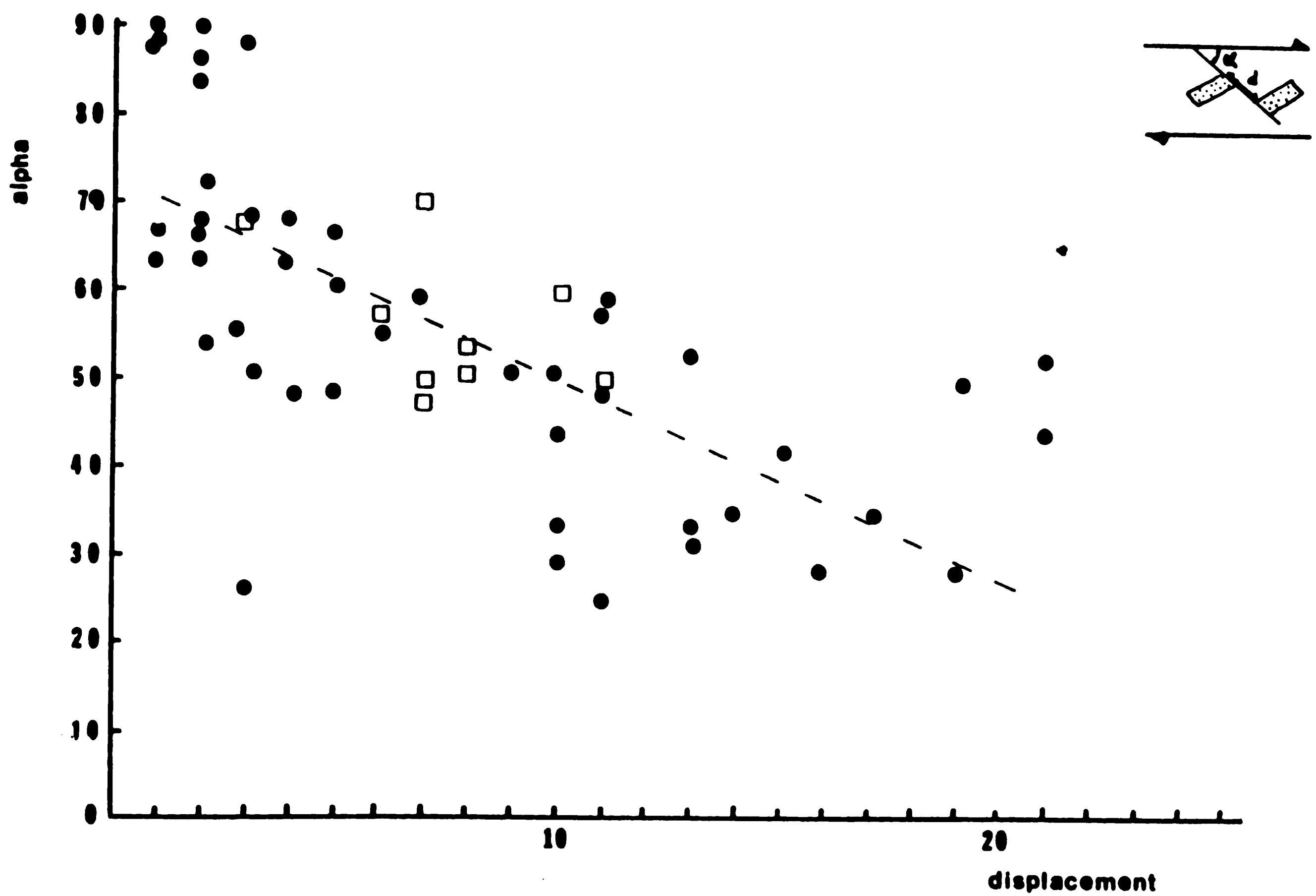


Figure 43. Plot of the angle of the shear band with the shear zone (alpha) versus the displacement. The circle and the square refer to values measured on two different levels separated by a 40-cm vertical distance (square lower level).

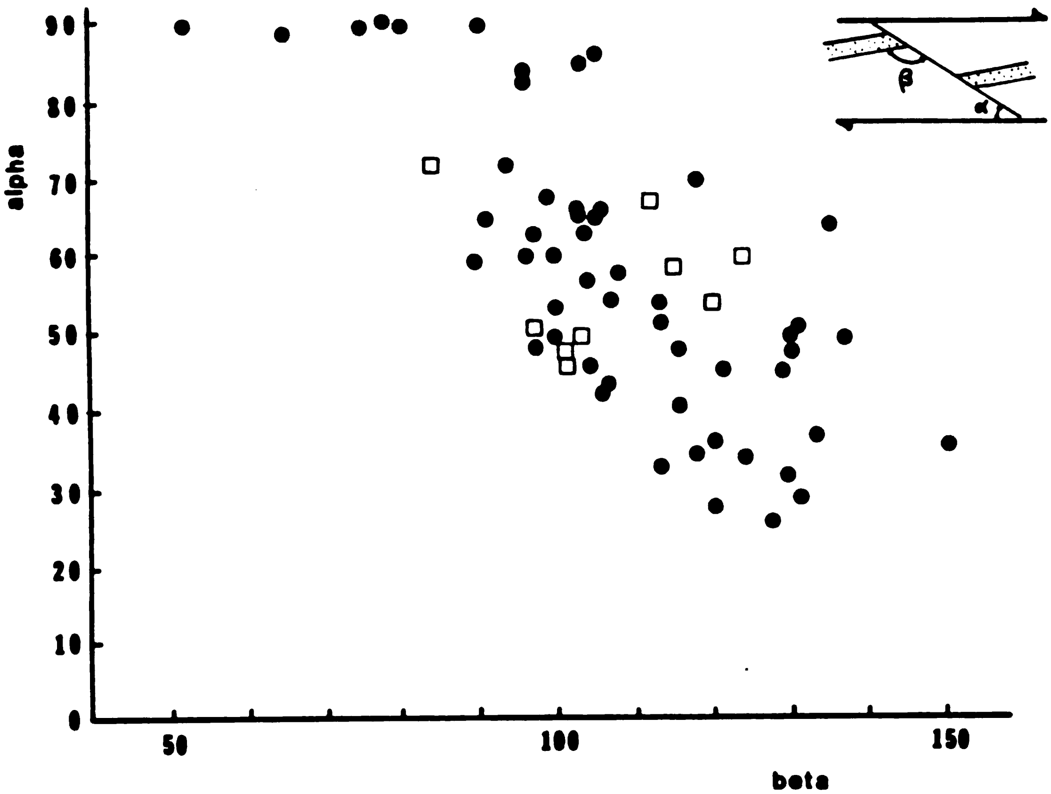


Figure 44. Plot of alpha versus beta, the angle between the bedding plane and the shear band. Same symbols as in figure 43.

High values of alfa and beta and the association high alfa-small displacement are interpreted as joints at right angle to the bedding. These joints could have been reactivated mimicking the behavior of the shear bands during the rotation of part of the layers undergoing shearing along the detachment plane. The high angle joints become locked very easily because of the following clockwise rotation towards the infinitesimal extensional axis. The existence of a reactivated joint system shows the inconsistency of a synsedimentary soft sediment deformation.

Another important observation that tends to rule out soft-sediment conditions at the time of sill emplacement is the lack of pervasive sediment deformation phenomena as described by Kokelaar (1982) and Krinauw and others (1988) in Antarctica and in Connecticut. In these conditions, wet-sediment fluidization structures develop together with the formation of granite and granophyre and disharmonic folding in large parts of the sedimentary sequence intruded. These fluidization phenomena overprint any sedimentological feature and create chaotic structures within at least three meters of the sill margin; nothing similar has been observed at Kama hill.

The presence of joints in the rocks prior to the development of the detachment introduces a problem which must be addressed. In Hanmer (1986) analysis of boudins in shear zones he points out that angular boudins should rotate sympathetically with the shear sense (class 1 boudins; top row in fig. 45) and one might expect joint blocks to do this.

However in the case Hanmer describes, the boudins are separated whereas the joint blocks are not. It seems that in the Rossport Formation shear bands developed in the rock and these subsequently controlled the type of movement in the shear zone (type 2B boudins according to Hanmer, 1986). Any reactivation of joint surfaces mimicked the shear bands because of this. If one accepts the interpretation of these discontinuities as shear bands then there is a consistent top to the south sense of movement.

Hanmer (1986) refers to boudins whose rotation is opposite to the sense of motion of the shear zone as class 2 boudins. In particular he further subdivides this class into two types: type 2A where the back-rotation of the boudins is due to their pinch-and-swell form and type 2B where the back -rotation is controlled by the shear bands. According to Hanmer (1986) type 2B asymmetrical structures are associated with pinch-and-swell extension of competent layers.

In the upper member of the Rossport Formation the clay content increases considerably, so that the competence/viscosity contrast within the layers could be that considered by Hanmer (1986) to be responsible for the formation of the Type 2A pull-apart structures (fig. 45). The presence of these structures, of the jointed boudins and

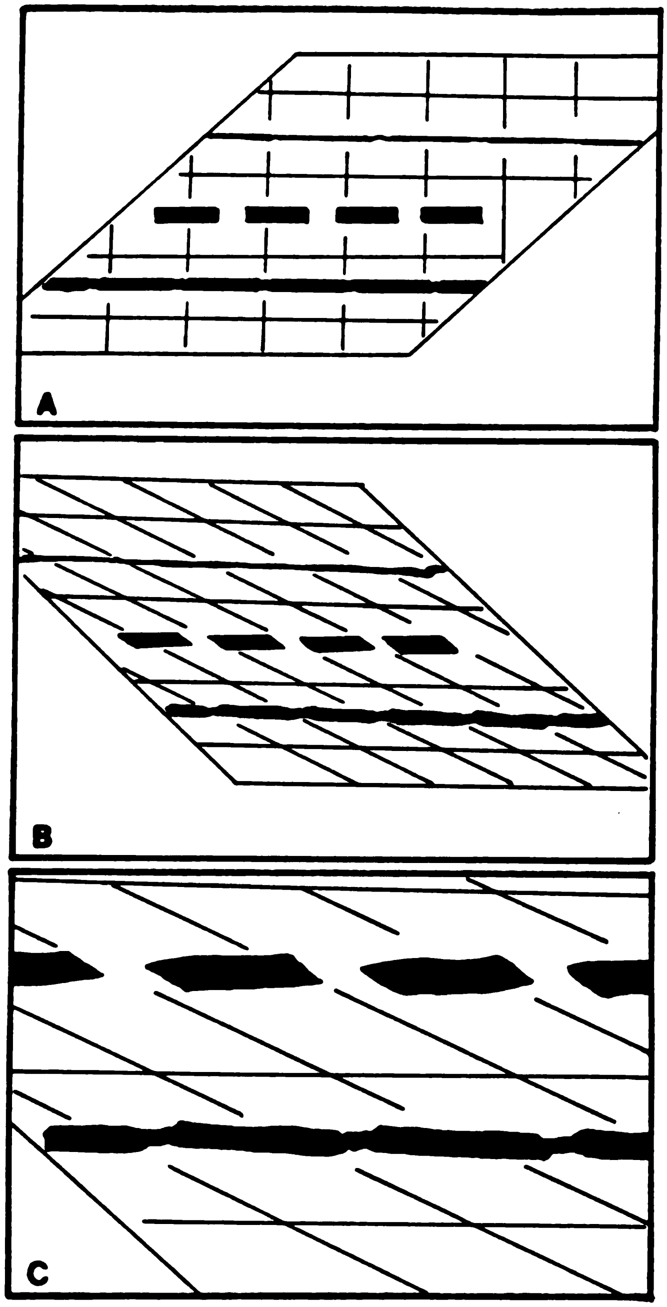


Figure 45. Plasticine-silicone putty inserts in silicone putty matrix. Model subjected to a sinistral shear strain of two gamma in these pictures. (a) Initial configuration, rectangular grid of passive marker lines. (b) After deformation. (c) A detail of (b) (modified after Hanmer, 1986).

of the shear bands at the same time cause ambiguity in the sense of shear determination; such ambiguity is apparent in the stratigraphic level in question.

We can conclude that the sedimentary sequence was consolidated at the time the shear was active in the area. The bedding-parallel shearing, constrained by the presence of boudins and shear bands, is consistent with the hypothesis that the detachment was climbing up the sedimentary sequence following discrete stratum planes.

Petrographic-microfabric study

This analysis has been made to find in the microfabric of the rocks kinematic indicators consistent with those described in the field at the mesoscopic scale. Two kinds of shear strain have been identified, one related to the top to the south movement during the regional extension and one to sill emplacement. The latter indicates a top to the north sense of shear at the lower boundary of the sill intrusions at Kama hill and follows in time the regional strain.

Thin sections of samples taken in proximity of the shear bands (sample FS 1), at the sill country-rock contact (samples 18, 18a) and in the Rossport Formation close to a sill step-up (sample 9) have been cut and analyzed with standard optical technniques. A slab of rock in which reduction spots were present has been cut along three perpendicular planes to detect any kind of strain on these markers. The reduction spots were spherical on the three sections which suggests that the shear was localized and not penetrative.

The most interesting thin sections were those taken on the boundary layer between sill intrusion and country rock. Descriptions of these thin sections, of one taken on a shear band and of one taken close to a sill step up are given in the following paragraphs.

Thin sections description

Sample 18

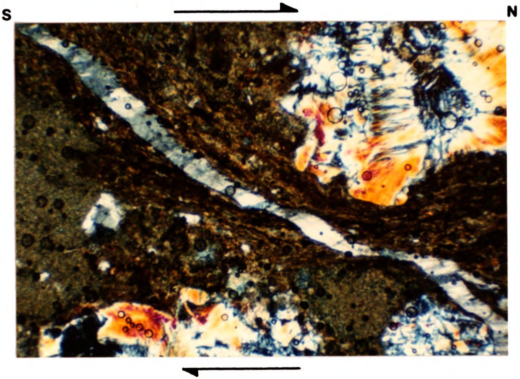
Location: Kama Hill.

Stratigraphic and structural relations: Contact between the sill and the Rossport Formation. Lower basal contact of the sill. Orientation of the thin section : North-South (longer side N-S).

Description: The basalt originally composing the sill has been completely replaced by carbonate. The rock of the Rossport Formation has been partially mixed and intruded by the basalt, now replaced by carbonate. Where the original host rock was sandstone it is now possible to identify chert aggregates with the characteristic fibrous, commonly radially spherulitic form of the crystals (chalcedony). The size of the aggregates is coarse. An asymmetric carbonate-filled vein (fig. 46 and fig. 47) provides a good kinematic indicator of shear top to the north. The calcite crystals are parallel to the shear plane showing a good example of a pull-apart structure. Minor calcite veins surrounding the major one have the same characteristics. Calcite veins sometimes pass through the chert aggregates and the platy minerals (chlorites) of the roughly horizontal shear zone determined by the fabric of the mineral grains. However, locally, some chert aggregates seem to cut through the calcite veins previously described (not the major ones, but those with a sense of shear top to the north).

Figure 46. Asymmetrical pull-apart vein. Sample 18.

Figure 47. Asymmetrical pull-apart vein and mineral fabric denoting a layer parallel shearing. Sample 18.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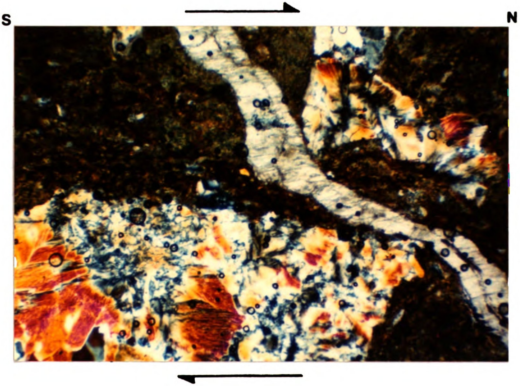
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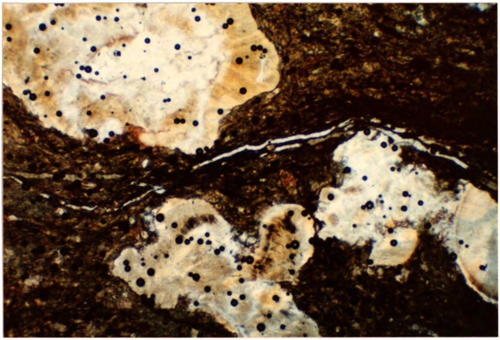
The shear direction is parallel to the north-south axis all over the thin section; however, it is affected by the large chert aggregates which seem to have grown after its formation (fig. 48). Locally the growth of the chert affected the shape of the shear zone, especially at the grain boundary.

Some blobs of basalt replaced by carbonate seem to be enclosed completely in the shear zone. A quartz vein is parallel to the sense of shear of the platy minerals along an horizontal plane (fig. 49). This quartz vein is parallel to another vein this time filled with carbonate (probably replaced by carbonate); calcite crystals don't show clearly the top-to-the-north sense of shear; this latter vein cut through the chert aggregates (hence it is of later origin). The shear zone marked by the platy minerals seems to be a preferential zone for vein formation (easy to break or pull apart). Streaks of iron oxides run parallel to the horizontal shear plane.

A small layer of quartz arenite (with a very high content of chert), completely enclosed by the carbonate, shows a boudinage-like sheared structure (fig. 50). At the apex of the sheared boudins the quartz crystals seem to align in the direction of the shear. It is hard to follow the plane of shear in the carbonate. Sometimes it seems that it follows a calcite vein, other times it seems it disappears into a calcite vein. The shear sense detectable from this small layer is top to the south contrary to that of the calcite

Figure 48. The growth of chalcedony aggregates affects the fabric of the platy minerals in the shear zone. Sample 18.

Figure 49. Asymmetrical vein and platy minerals aligned on the plane of bedding. Boudins in quartz filled veins show top-to-the-south sense of shear. Sample 18.



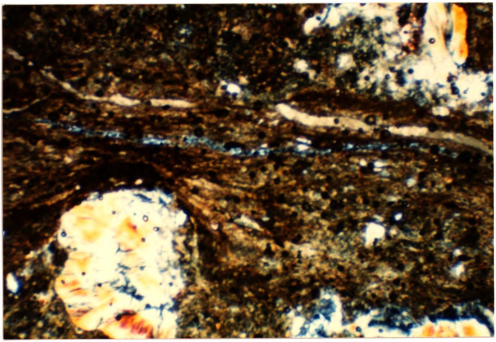

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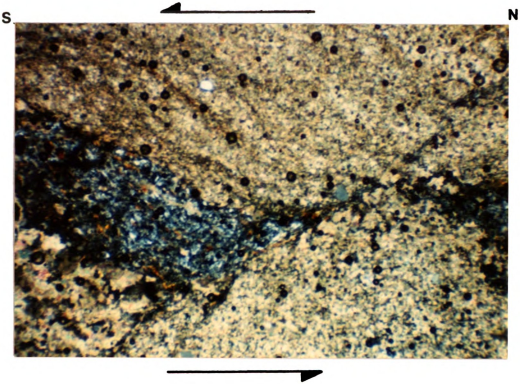
Figure 50. Corner of a sheared boudin (top-to-the-south). sample 18.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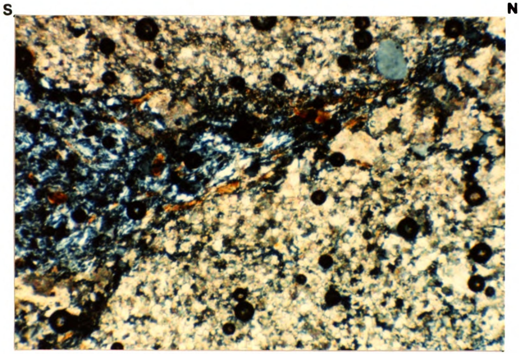
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veins; it is thought to be earlier and probably related to the regional shear connected with the North-South extension; the platy minerals aligned along the horizontal plane could be the result of the same shear as well.

Some polycrystallyne quartz grains have equant single crystals completely recrystallized with no evidence of undulose extinction (fig. 51). These large quartz grains are completely enclosed in the carbonate of the sill, and are almost exclusively present in the carbonate replacing the sill, while among the chert of the metamorphosed Rossport Formation we can recognize only fibrous aggregates (chalcedony). This observation very likely reflects the different degree of metamorphism associated with the intrusion.

Sample 18a

Location: Kama hill.

Stratigraphic and structural relations: Contact between the sill and the Rossport Formation. Lower basal contact of a 1 meter thick sill.

Orientation of the thin section: North-South (longer side N-S)

Description of the thin section: A shear zone with the characteristics described in the thin section of sample no. 18 is recognizable also here. However, here the veins are

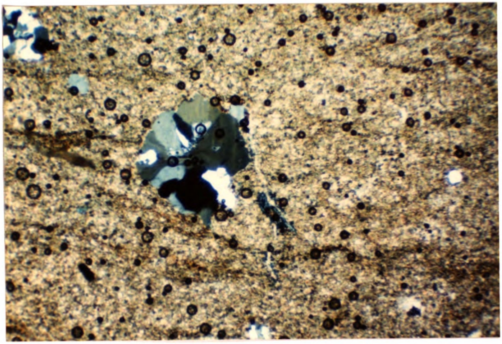


Figure 51. Polycrystalline quartz grain in the carbonate that replaced the basalt. Sample 18.

filled (very likely as a replacement) by large crystals of calcite which presumably overprinted any preexisting fabric of the crystals.

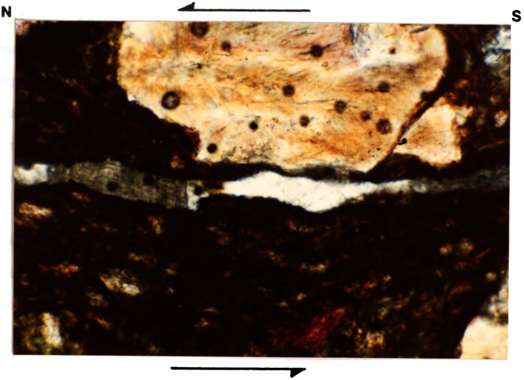
There is a north-south trending vein with boudins of calcite and quartz (fig. 52). It indicates a sense of shear top-to-the-north; in some boudins of this veinlet, especially where the crystals were not replaced by carbonate, it is possible to recognize in the cleavage of the crystals the sense of shear as it is detectable in the shear bands between the boudins (eccl). The growth of the chert aggregates seem to have affected the pattern and the path of this veinlet, hence its formation should precede the chert growth. Other smaller veins with these characteristics (although less evident) are present in proximity of the contact with the intrusion.

Large veins with a vertical trend are generally filled with coarse calcite crystals; these latter veins pass also through the chert aggregates.

Some boudins are also visible in two other veins (fig. 53), which slightly converge; also these veins show a top to the north sense of shear, the lower vein then merges in a secondary late vertical vein filled with large carbonate crystals. Both the veins are filled, probably as a replacement, by coarse calcite crystals.

Figure 52. North-South trending vein with boudins filled with quartz and calcite. Sense of shear top-to-the-north. Sample 18a.

Figure 53. Boudins in slightly converging veins. Top-to -the-north sense of shear.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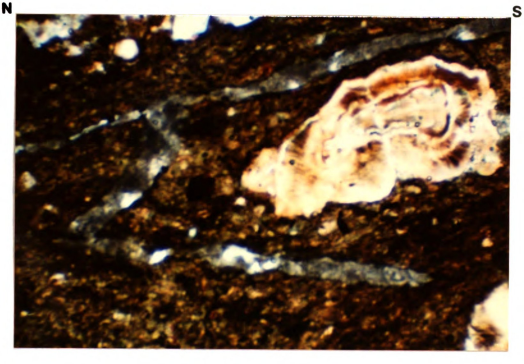
N
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Sample FS 1

Location: Kama hill, Railroad section.

Stratigraphic and structural relations: Quartz arenite sample along a shear band.

Orientation of the thin section: Perpendicular to the shear band plane.

Description of the thin section: The most abundant grains are made up of quartz (mostly monocrystalline with undulose extinction less than 5 degrees). Some quartz grains show overgrowths attacked and replaced by the calcite cement. No calcite twinning was observed.

The packing and the sorting of the grains changes locally according to the iron oxide content of the sandstone. Locally the grains are tightly packed but elsewhere they are randomly scattered. Some microcrystalline grains were observed. Most of them are highly altered and replaced by carbonate cement. Also chert grains have been observed in a consistent quantity. The suggested provenance is that from plutonic rocks of the basement, or high grade metamorphic rock with a considerable source of chert fragments probably coming from the unroofing of the area overlain by the Gunflint formation of the Animikie group.

Sample 9

Location: Kama hill, road section.

Stratigraphic and structural relations: Sample collected 40 cm below a sill step-up in the sedimentary sequence.

Orientation of the thin section: Three thin sections were cut at 90 degrees one from the other. This description summarizes the petrographic characteristics for all of them.

Description of the thin section: The quartz grains are enclosed in a matrix of carbonate and iron oxides. Some microcline grains and quartz grains with undulose extinction (greater than 5 degrees) have been observed. Locally a preferential fabric of the grains can be attributed to deposition in the original sedimentary environment.

Secondary overgrowths on the quartz grains can be recognized, some of them are replaced by the carbonate cement. Overall quartz overgrowths are not very common. The quartz overgrowths seem to have occurred after the deposition of the sandstone; in fact in most of the cases there are no overgrowths at the grain contacts. Furthermore some complex quartz overgrowths have been observed in which the quartz overgrowths of one grain in optical continuity with the grain itself is completely enclosed or surrounded by the overgrowths of another grain. It is evident how diagenetically the carbonate cement precipitated after these quartz overgrowths; most of the times replacing them.

Detrital carbonate has been recognized as a conspicuous constituent of the matrix. The presence of detrital carbonate implies that the sandstone was not buried at great depth and it didn't undergo any kind of metamorphism (Sibley personal com.); in fact the sandstone appears very poorly compacted. Any large increase in pressure and/or temperature would have caused the dissolution of the detrital carbonate, that on the contrary is still recognizable in the matrix.

From the analysis of the relations among the microstructures observed in the samples 18 and 18a it is possible to draw some conclusions.

Two kinds of structures related to shear strain have been described. The first one gives a sense of motion top to the south (quartz arenite boudins) and an overall horizontal bedding-parallel sense of shear. The second one (extensional pull-apart veins) has also a bedding parallel plane of shear but this time the sense is top to the north.

On the basis of cross-cutting relationships the first one occurred before the second one. I relate the top to the south shear to the regional extension along a detachment plane that entered the sedimentary sequence and followed the bedding discontinuities during its development. The top to the north sense of shear is probably connected to the later intrusion of the sills along the bedding-parallel detachment plane and to a local amount of shear stress induced by the magma emplacement dynamics.

The shear zones clearly seem preferential zones for sill intrusions.

Model for the shear stress at the wall of a sheet intrusion

Flowing magma applies a shear stress at the contact with the country rock; drag folds, rotations of boudins and extensional pull-apart veins at the boundary of certain sill intrusions which could be related to this agent have been observed at Kama hill.

In order to evaluate the magnitude of the shear induced by a flowing magma in a sill and to point out in which kind of intrusions this factor is more important a physical model used by engineers (Wiggert et al., 1990) which describes flow between two parallel plates has been adopted. The fundamental equation of this model is similar to that mentioned, but not developed by Pollard (1973) in his paper which was mainly concerned with slow viscous sialic magmas. To apply the model of flow between two fixed parallel plates it is necessary to make a few assumptions. The first is that the liquid is a simple Newtonian fluid. This seems reasonable given the low viscosities of a basaltic magma (Pollard 1973, Murase and McBirney 1973, Wilson and Head 1981). The second assumption is that the flow is laminar, hence with a Reynolds number (Re) less than 2000. The boundary value of the Reynolds number is that commonly used in engineering applications, and as it will become apparent in the development of the model, the range of values for the velocity and the viscosity of the magma are consistent with a $Re < 2000$.

In this simplified flow situation the velocity distribution has a parabolic profile and is given by

$$u(y) = (1 / 2 \mu) \ [d(p + \gamma h)/dx] \ (y^2 - a y)$$

where "u" is the velocity, "a" the distance between the two plates, μ the viscosity, h the elevation respect to a datum plane, γ the specific weight and y the distance from the bottom of the lower plate.
From this velocity distribution it is possible to evaluate the flow rate per unit width,

$$Q = - (a^3 / 12 \mu) \ [d(p + \gamma h)/dx]$$

and hence calculate the average velocity

$$V = Q / A = - (a^2 / 12 \mu) \ [d(p + \gamma h)/dx]$$

In terms of average velocity the pressure drop becomes

$$\Delta p = - (12 \mu V L) / a^2$$

and considering a developed flow, where dp/dx = constant we can derive the maximum velocity

$$u\ max = - (a^2 / 8 \mu) (dp/dx) \qquad \text{at } y = a / 2.$$

The shear stress by definition is

$$\tau = \mu (du/dy) = (1 / 2) (dp/dx)$$

and at the wall where y = 0

$$\Delta p = 2 \tau_0 L / a$$

or introducing the dimensionless friction factor

$$f = \tau_0 \; 8 / (\rho V^2) \qquad (1)$$

we have

$$\Delta p = f (L / 2 a) (V^2 / 2 g).$$

From this latter expression we can derive the head loss

$$hl = f (L / 2 a) (V^2 / 2 g)$$

and rearranging we find an expression for f in the case of a laminar flow

$$f = 48 / Re \qquad (2)$$

where Re is the Reynolds number a dimensionless parameter controlling the characteristics of the flow.

The Reynolds number is defined as

$$Re = \rho V a / \mu \qquad .$$

To solve for the shear stress at the wall from equations (1) and (2) it is convenient to write a short program which is listed in the Appendix.

To model the shear stress at the wall of the intrusion one must first choose a range of values for the critical parameters governing the flow.

Viscosity - The viscosity for the basalt of the sill is considered, using the data of Murase and McBirney (1977), to be in the range 10^3 - 10^6 poises. The value of $3x10^4$ poises is the one most commonly accepted for a hot flowing basaltic magma.

Velocity of the magma - This is one of the most delicate and important factors because the results vary by the square of the velocity in the equations used. The Wilson and Head (1981) model for basaltic eruption constrains magma velocity along a fissure to a range of 0.01 to 100 m/s. These velocities are required to allow eruption of basalt from depths of 0.5 - 20 Km with negligible yield strengths, viscosities greater than 10^3 poises and crack width in the range 0.2 - 0.6 m.

Furthermore Wilson and Head (1981) point out that for commonly occurring terrestrial basalt the rate of ascent of the magma must be greater than 0.5 - 1 m/s to avoid strombolian phenomena.

This model will consider magma velocities in the range indicated in Table 5 of Wilson and Head (1981) with modifications for viscosity and intrusion thickness (V = 0.2, 0.8, 1, 2, 10, 100 m/s).

A comparison between the velocity of a sill intrusion and a dike intrusion is possible for the following reasons. The magma in sill intrudes along a parting surface which considerably helps the melt in overcoming the yield strength of the rock and, as experimentally shown by Pollard (1973) drives the shape of the intrusion into an horizontal sheet; under these conditions the a sill would intrude more easily than a dike.

The second reason directly follows from the continuity relation in fluid mechanics, when considering the existence of dikes and sills in close association. The magma flowing in a fracture with constant cross sectional area should have a constant velocity at any point to provide the same rate of flow. During the process of reorientation of dikes into sills the scalar velocity of the magma is not expected to change, so the figures valid for the dikes are also valid for the sills.

Friction along the contacts does not affect the velocity of the magma (continuity relation) but reduces the magmatic pressure significantly. This latter observation is important because in order to explain sills of large areal extent we have to assume low viscosities and low friction factors (Re = 1000-2000) and consequently a Newtonian behavior for the magma. On this basis Pollard (1973) considers dike and sill intrusions as pressure gages for the magma.

Density of the magma - The value 2850 Kg/m3 is assigned for the purposes of this study (Bradley, 1965).

Thickness of the sill - Three different thicknesses are considered: 0.1, 0.8 and 100 meters which covers the range sills present in the area of study. Structures related to shear stress at the wall of the intrusion have been observed in a 0.8 m thick sill at Kama hill.

The shear stresses modeled with these assumptions are summarized in Table 1, 2, 3 and presented in compact form in the diagram of fig. 54.

The straight line relationships observed on the log-log plots indicate a shear-stress/velocity relationship typical for laminar flow. Large shear stresses are associated with high velocities, high viscosities and small thicknesses of the intrusion. From the diagram it can be seen that shear stresses associated with large intrusions (about 100 m) are always negligible for acceptable values of viscosity and velocity whereas for thin sills they are extremely high.

It is important to point out that the shear stress is

TABLE 1. Shear stress at the wall of a sheet intrusion for a 0.1 m thick sill.

Thickness of the sill = 0.1 m
Density of the magma = 2850 Kg/m^3

Viscosity = $3x10^4$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	36000	3.7
0.8	144000	14.7
1.0	180000	18.3
2.0	360000	36.7
10.0	1800000	183.5
100.0	18000000	1835.0

Viscosity = $3x10^5$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	360000	37
0.8	1440000	147
1.0	1800000	184
2.0	3600000	367
10.0	18000000	1835
100.0	180000000	18342

Viscosity = $3x10^6$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	3600000	367
0.8	14400000	1468
1.0	18000000	1835
2.0	36000000	3670
10.0	180000000	18349
100.0	1800000000	183486

TABLE 2. Shear stress at the wall of a sheet intrusion for a 0.8 m thick sill.

Thickness of the sill = 0.8 m
Density of the magma = 2850 Kg/m^3

Viscosity = $3x10^4$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	4500	00
0.8	18000	00
1.0	22500	00
2.0	45000	00
10.0	225000	23
100.0	2250000	00

Viscosity = $3x10^5$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	45000	00
0.8	180000	18
1.0	225000	23
2.0	450000	46
10.0	2250000	230
100.0	22500000	2294

Viscosity = $3x10^6$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	450000	45
0.8	1800000	184
1.0	2250000	230
2.0	4500000	459
10.0	22500000	2294
100.0	225000000	23000

TABLE 3. Shear stress at the wall of a sheet intrusion for a 100 m thick sill.

Thickness of the sill = 100 m
Density of the magma = 2850 Kg/m^3

Viscosity = $3x10^4$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	36	00
0.8	144	00
1.0	180	00
2.0	360	00
10.0	1800	00
100.0	TURBULENT FLOW !	

Viscosity = $3x10^5$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	360	00
0.8	1440	00
1.0	1800	00
2.0	3600	00
10.0	18000	2
100.0	180000	18

Viscosity = $3x10^6$ Poises

Velocity (m/s)	Shear stress (Pascals)	Water col. (m)
0.2	3600	00
0.8	14400	1
1.0	18000	2
2.0	36000	4
10.0	180000	18
100.0	1800000	184

Figure 54. Model for the shear stress at the wall of a sheet intrusion. Continuous line: 0.1 m thick sill Dashed line: 0.8 m thick sill. Dashed-dotted line: 100 m thick sill. (a) viscosity = $3x10^4$ Poises. (b) viscosity = $3x10^5$ Poises. (c) viscosity = $3x10^6$ Poises.

confined to the area immediately adjacent to the walls of the intrusion and cannot be transmitted very far into the lithified wallrock. This fact together with the effective absolute magnitudes of the shear stress possible in the model, rule out any effect of shear stress due to the flow in the lithified country rock. However contact metamorphism may partially mobilize a layer close to the contact and the shear stresses cause drag phenomena for the values of stress indicated in the diagram.

For a 0.8 m thick sill shear stresses at the wall within a range of 10^5 - 10^7 Pascal are to be expected as average values for this model. Such values seem to be a plausible explanation for the rotation of the boudins, the pull apart veins and the drag folds observed at the wall close to an intrusion at Kama hill (fig. 55).

Despite the extremely high shear stress associated with the very thin intrusions (about 0.1 m) no macroscopic shear phenomena have been observed along their margins. This apparent contradiction can be explained by the dependence of the velocity on the friction; the higher the velocity the larger the drop in pressure (head loss) so that after a short time a very thin intrusion will not be able to sustain a high enough pressure to hold the fracture open and therefore will stop. It follows that a thin sill will only continue to intrude if the emplacement velocity is low, otherwise it will close. This consideration is implicitly stated in Table 5 of Wilson and Head (1981).

Figure 55 b and c. Rotation of boudins at the boundary with the sill intrusion. Kama hill, southern outcrop.

B

C

Having demonstrated theoretically that stress can be important during intrusion the field evidence which supports this idea will now be examined. It was shown earlier that the pattern of boudinage in the Rossport Formation at Kama Hill is consistent with development in an already lithified and jointed sandstone layer. The joints represented discontinuities and, during simple shear parallel to the bedding associated with a regional detachment, they allowed the development of Hanmers' class 2 boudinage (Hanmer, 1986). These structures clearly indicate a top-down-to-the-south movement.

Using the model presented above sill intrusion would be expected to produce a second later shear strain at the sill margins. If intrusion was from the south the sense of shear on the lower margins would be the reverse of the earlier one.

A model to explain the disturbance introduced by the intrusion on the row of boudins is presented in fig. 56. The kinematics of the boudins are controlled by three displacement vectors of which the second (no.2) is directly connected to the shear stress at the wall (anticlockwise rotation) during magmatic intrusion. Displacement vector number 1 is active at the front of the sill before its emplacement, and displacement vector number 2 acts perpendicularly to the sheet intrusion once it has intruded the country rock; these two displacements have been added to

Figure 56. Model for the rotation of boudins at the boundary with a sill intrusion. (a) Boudins with shear bands indicating a sense of shear top to the south connected to the regional extension. (b) Sill causes the rotation of the boudins, furthermore the kinematic of displacement is governed by the stress field induced by the intrusion (see fig. 6c and explanation in the text at pag. 129). (c) Situation after the sill emplacement.

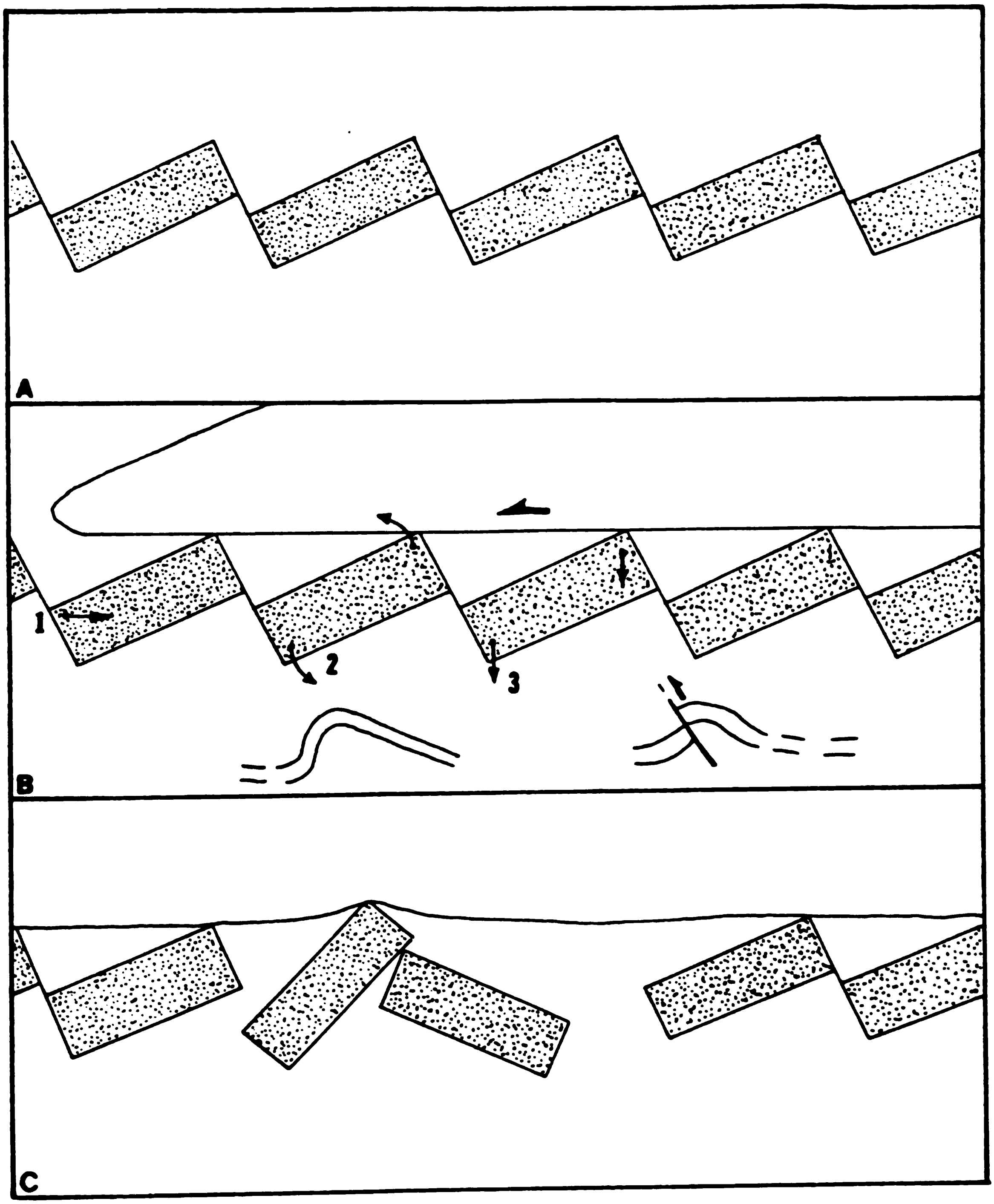


Figure 56

complete the kinematic description (reference to fig. 6). Given the results of the stress modeling, one could ask why the sill intrusion was not also responsible for the initial formation of the boudins as it worked its way through the parting surface. First of all the model presented in this work has shown that the shear stress is important only at or very close to the contact.

In addition the magnitudes of shear stress required to create shear bands in a consolidated sedimentary sequence seem to be much larger than those associated with the flow of magma in a sill. However the shear stress at the wall is important to create local structures such as those observed in the boundary layer of the sill. The field evidence is most persuasive. There are boudins in sandstone layers which are some considerable distance from a sill. The sense of movement in the disrupted boudins close to the lower margin of one of the sills shows a reverse sense of movement when compared with the earlier sense of shear strain which is consistent with intrusion from the south. As viewed in fig. 56, looking east the earlier movement indicates clockwise shear (top down to the south), the later one which disrupts the boudin patterns an anticlockwise sense of shear. In undisturbed regions away from the sill margins only the first, clockwise rotation is present. The direction of emplacement of the magma from south to north is confirmed by some drag folds detected in the same outcrop and by some kinematic indicators in a thin section taken on the boundary layer. In summary the sense of shear in fig. 56 caused by the sill is left lateral (as detected from the rotation of some boudins and from the drag folds) while the sense of shear as detected from the shear bands is right lateral. From field evidence it is clear that the shear bands formed first as result of top-down-to-the-south movement and then the sill intruded inducing a top-up-to-the-north sense of movement on its lower surface. Shear strain induced by the sill at the upper contact of the intrusion would have the same sense as that of the earlier detachment induced shear, which may account for why the complex disturbances are restricted to the lower margins of the sill. The observations on the rotation of the boudins and on the drag folds in the sediment close to the parting surface where the intrusion occurred are also important because they constrain the sense of emplacement of the sill, from the south to north. This fact is relevant in the development of the geological model.

DISCUSSION AND CONCLUSIONS

The most important result gained from the analysis of the mesostructures at the Kama hill outcrop is that the pattern of the kinematic indicators permits the resolution of two episodes of shear strain; one connected to sill intrusion and the other to regional extension. The latter is present also in two other outcrops and is not directly associated with intrusive sheets. Its sense of shear can be generally constrained as top-to-the-south - southwest by different kinds of mesostructures (shear bands, thrusts, asymmetric folds). However, some ambiguous kinematic indicators are present.

In particular, the compressional features in the lower part of the section at Kama hill could represent a problem in the framework of an extensional shear zone related to the opening of the Kewenawaan rift system. The problem is the coexistence of compressional and extensional structures at the same time and in close association. Platt and others (1986) suggested a valid explanation in the study of a shear zone outcropping in a subduction wedge of Pakistan. Their hypothesis can be briefly summarized in these terms: if a portion of a shear zone causes more resistance to the movement (it is an obstacle to the movement) than the surrounding ones it will create a local increase in the horizontal shear stress. This will cause an increase of the horizontal normal stress ahead of the obstacle and a decrease of the horizontal normal stress at the back. The result is that in front of the obstacle we have compression and consequent formation of thrust faults, whereas behind the obstacle we have extension and formation of normal faults. This is shown in fig. 57. The increase in friction, the "obstacle" described by Platt, could be caused by the geometry of the shear plane. This last hypothesis was recently explored by Ellis and McClay (1988) with sand box experiments. In their models a normal extensional ramp-flat fault creates reverse faults to accommodate the excess of sediment above the ramp (fig. 58). The ramp-flat fault extensional model is very similar to the extensional model which can develop in a shear zone. Here the flats are the portions of the shear zones that follow the sedimentary layers, whereas the ramps are the steps up of the shear zone in the sedimentary sequence. This geometry of the shear zone can cause locally the compressional phenomena observed. However, the geometry of the thrust faults in the Ellis and McClay model does not closely resemble that described at Kama hill; this could be interpreted either using a different kind of obstacle or implying a prevalent role of the layered sequence in driving the horizontal stress and in constraining the reverse faults to a very low angle shape.

The important observation is that compressional features can exist in the framework of an extensional shear zone and that their sense of shear is top to the west - southwest, very

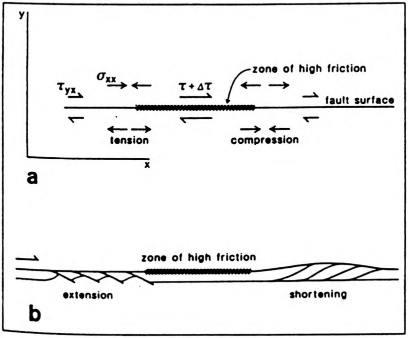


Figure 57. (a) Variation in shear stress and deviatoric effective normal stress parallel to fault plane with local zone of high frictional resistance. (b) Predicted footwall deformation (from Platt and Leggett, 1986).

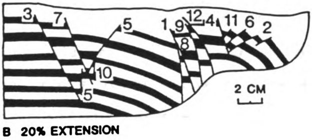


Figure 58. Reverse faults formed on a ramp-flat extensional fault in a sand-box model experiment (from Ellis and McClay, 1987).

closely approximating that of other indicators.

The other ambiguity observed in the field data is the reverse sense of shear "top to the north" indicated by the faults if interpreted as "shear bands" in a restricted vertical span along the railroad section. The problem of kinematic indicators pointing out a sense of motion opposite to that generally present in a shear zone is reported in the geological literature and Reynold and Lister (1990) have presented a resolution of this problem in a recent paper. Conjugate shear zones are the best explanation that these authors can come up with to explain the reverse sense of shear. The amount of data available for my research does not allow me to make any further speculation about this problem and I will consider it an open question. I can just add to this a few observations about these "shear bands" that perhaps are more correctly identified as normal faults in this part of the section.

Plotting alpha versus displacement (fig. 59) for these structures (Kama hill railroad section) we see that there is a not a strong correlation along a line with a negative slope as in fig. 43. Furthermore plotting alpha versus beta (fig. 60) it is evident how the beta angle is constrained within a range of 40 degrees from the vertical. The high dip angle of these structures (90-50 degrees) from the bedding planes suggests they could be normal faults developed during extension in the lower plate, some distance from a discrete shear zone. This is also confirmed by the existence of some

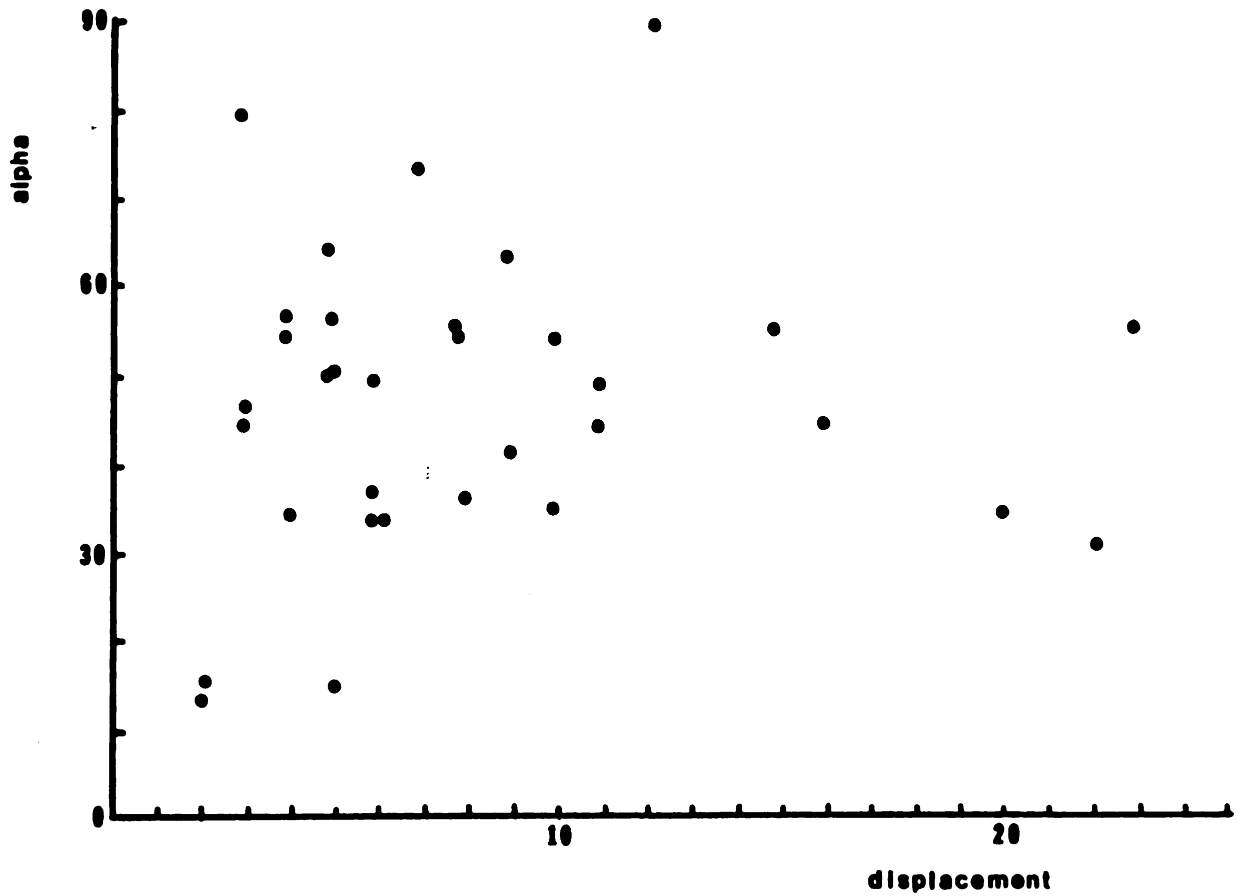


Figure 59. Kama hill, railroad section. Plot of the angle of "shear band" with the shear zone (alpha) versus the displacement.

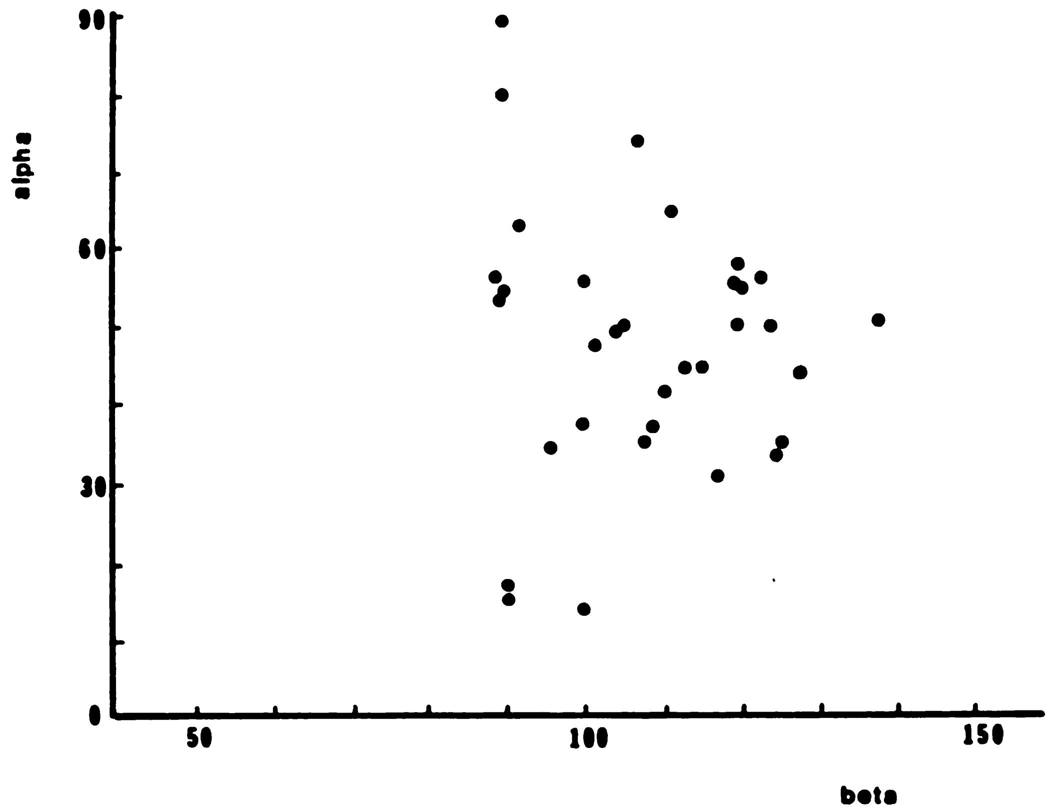


Figure 60. Kama hill, railroad section. Plot of alpha versus beta, the angle between the bedding plane and the shear band.

antithetic faults delineating some small grabens. Furthermore the values of alpha and beta indicate strong rotation of the boudins (expecially for the values with a low alpha and an high beta) as predicted by the model of the type 1 asymmetrical pull-aparts of Hanmer. This latter behavior could have here been caused by the lack of shear bands to drive the flow of the boudins' corner. The shear bands very likely didn't develop here because this location is not on one of those discrete shear zones that higher in the section have controlled the sill emplacement. In conclusion the structures present in this part of the section are not shear bands but normal faults, and minor amounts of top-to-the-south bedding parallel shear in this lower plate could be detected by the rotation of type 1 asymmetrical pull apart boudins. Other points to consider in the analysis of the kinematic indicators is the possibility that the rheological conditions of the rock could locally change, that anisotropies are not preexisting or that they do not develop so extensively as in other areas.

I have introduced these considerations to give a possible solution for this inconsistency in the sense of shear for this part of the section. The problem could only be solved by identifying the interplay between the individual shear zones and the extension in the areas in between them for the whole sedimentary sequence. The exposures in Northern Ontario do not allow this.

The shear band analysis has pointed out the fact that the sedimentary sequence was consolidated at the time the shear took place. This is an important assumption with regard to the mechanics of magma emplacement following a detachment plane that is stepping up in a layered consolidated sedimentary sequence; in fact Pollard (1973) has shown experimentally that the presence of horizontal anisotropies is determinant in driving the magma into a sheet intrusion, regardless of the tectonic environment.

The petrofabric study has confirmed most of the observations made on the mesoscopic-macroscopic structures directly on the outcrop. In particular layers with microboudinage, alignment of platy minerals and extensional pull-apart veins have been used to constrain both the sense of shear and its timing. The top-to-the-north sense of shear indicators are later with respect to the top-to-the-south one in the boundary layer of the sill intrusion and its surroundings. The younger kinematic indicators are here considered to be related to shear strain induced by the intrusion. The shear stress responsible for this deformation seems to have a magnitude consistent with the values obtained from the physical model developed in this study. This model tries to constrain the shear stress at the wall of a sheet intrusion for different values of magma viscosity, velocity and thickness using an assumption of Newtonian flow between two parallel plates.

Considering this kind of shear strain, particularly well expressed by the drag folds and the rotated boudins at the wall of the intrusions at Kama hill, it is possible to infer a northward directed sense of emplacement for the magma. Furthermore the shear zones present in the layered sedimentary sequence were a preferential path for the magma intrusion as shown by the association sill-shear zone observed in the outcrop and in the petrofabric study.

All the elements discussed lead to the proposal of a geological model for Lake Superior rifting in which the sills, during their emplacement, took advantage of the discontinuities created by the detachment plane. As the detachment plane entered the sedimentary sequence it has deflected along the bedding planes. The sills followed the detachment in the sedimentary sequence undergoing shear as they stepped up to the north in the Nipigon area as far as the magmatic pressure was able to drive their emplacement. The basalt at the time of the intrusions was readily available through what very likely was a well developed interconnected magmatic plumbing system able to transmit to the magma the hydrostatic pressure necessary for its emplacement. The large quantity of magma formed through decompression melting induced by rifting over an area with anomalous mantle temperatures (hot spot).

As can be seen from the Bouguer gravity anomaly map (fig. 15) there is not a simple unique basin but a series of offset depocenters beneath the Superior region. This suggests that also the detachment plane is not unique but changes its dip to accomplish the extension in the different sub-basins. The areas across which the shear plane changes dip could be considered accommodation zones (Rosendhal 1987, Lister and others 1987) and they can represent distribution boundaries for the sill occurrence areas if their emplacement has been affected by the detachment plane geometry and extent as it is suggested in this study.

An overall scenario is currently emerging for continental extensional rifting thanks to deep seismic imaging of the crust and especially to the BIRPS group. Hutchinson and others (1990) have confirmed that the reflectors in the lower crust of the MRS are very similar to those detected in the western United States. These are ultimately interpreted as mafic intrusions (Goodwin and Thompson 1989, McCarthy and Thompson 1988, Warner 1990) on the basis of seismic modeling. The reflectors represent an area of stress relaxation accompanied by ductile behavior (induced by heat transfer) during extension. The symmetric distribution of the intrusions at depth and the aseismicity below a certain depth in most of continental rifts suggest a ductile extension at depth in the lower crust (Goodwin and Thompson 1989). On the other hand the upper crust could have a brittle behavior with consequent formation of simple shear extension controlled by a detachment plane; here the magma can be driven by the shear zone geometry and be asymmetrically emplaced. Under these assumptions the original Wernicke model must be modified in a way such that the detachment bottoms down at a midcrustal interface and not at the Moho.

Another point extremely important for the MRS relates to the magma availability. In a region that is weakly extending, intrusive horizontal sheet emplacement is favorable if plenty of magma is available (Goodwin and Thompson 1989), whereas vertical dike emplacement is easier in an area which is quickly extending like the Basin and Range Province.

The MRS doesn't seem to have been a region subject to strain rates like those described in the western United States, so intrusions of sills in the upper crust is a likely eventuality. Furthermore the asymmetry in the distribution of the sills in the sedimentary sequence around Nipigon Lake respect to the neighbor areas is what one would expect from a simple shear extensional model. In fact the detachment plane changing its dip across the accommodation zones directly controls sill emplacement in the brittle part of the crust, giving the magmatic intrusions an asymmetric distribution pattern. On the other hand a uniform symmetric distribution of the sills in the lower crust suggest ductile extension governed by pure shear.

APPENDICES

APPENDIX A

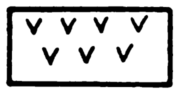
```
      PROGRAM NEWFLU
C     PROGRAMMER: MARCO ANTONELLINI, 3/17/1990
C     THIS PROGRAM IS DESIGNED TO CALCULATE THE
C     SHEAR STRESS AT THE WALL CAUSED
C     BY FLUID FLOWING BETWEEN TWO PARALLEL PLATES.
C     THE SHEAR STRESS AT THE
C     WALL IS MEASURED IN PASCAL. THE CALCULATIONS
C     ASSUME A LAMINAR FLOW.
      INTEGER N, I
      REAL VEL, VIS, SIND, DEN, RE, SHST, FRI, EQW
      N = 0
      PRINT *,'HOW MANY COMPUTATIONS DO YOU WANT?'
      READ *, I
 10   IF (N .LT. I) THEN
          PRINT *, 'ENTER VELOCITY OF MAGMA IN M/S'

          READ *, VEL
           PRINT *, 'ENTER VISCOSITY OF MAGMA IN
       POISES'
          READ *, VIS
          VIS = VIS * 0.1
           PRINT *, 'ENTER THICKNESS OF THE SILL IN
          M'
          READ *, SIND
           PRINT *, 'ENTER DENSITY OF THE MAGMA IN
         KG/M3'
          READ *, DEN
          RE = ( VEL * SIND * DEN ) / VIS
          IF ( RE .GT. 2000.0 ) THEN
             PRINT *, 'TURBULENT FLOW !'
          ELSE
             FRI = 48 / RE
             SHST = FRI * DEN * (VEL**2) / 8.0
             PRINT 30, VEL, DEN, RE, SHST
 30           FORMAT ('VEL= ', F10.2, 2X, 'DEN= ',
           F12.2,                   2X,
     +      'RE= ', F10.2, 2X, 'SHST= ', F12.0 )
          EQW = SHST / 9810.0
          N = N + 1
           PRINT *, 'EQUIVALENT COLUMN OF WATER IN
         METERS= ', EQW
          END IF
```

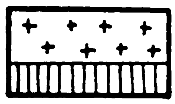
```
            GOTO 10
        ELSE
            PRINT *, 'END OF COMPUTATIONS'
            END IF
            END
```

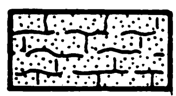
APPENDIX B

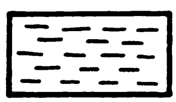
Stratigraphic section at Kama hill.

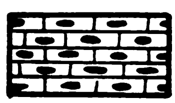
Legend:

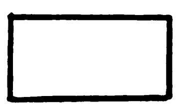

Diabase sill

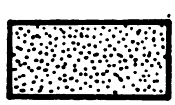

Chilled margin

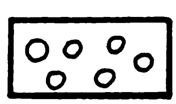

Sandy dolostone in prevalent clay + feldspar

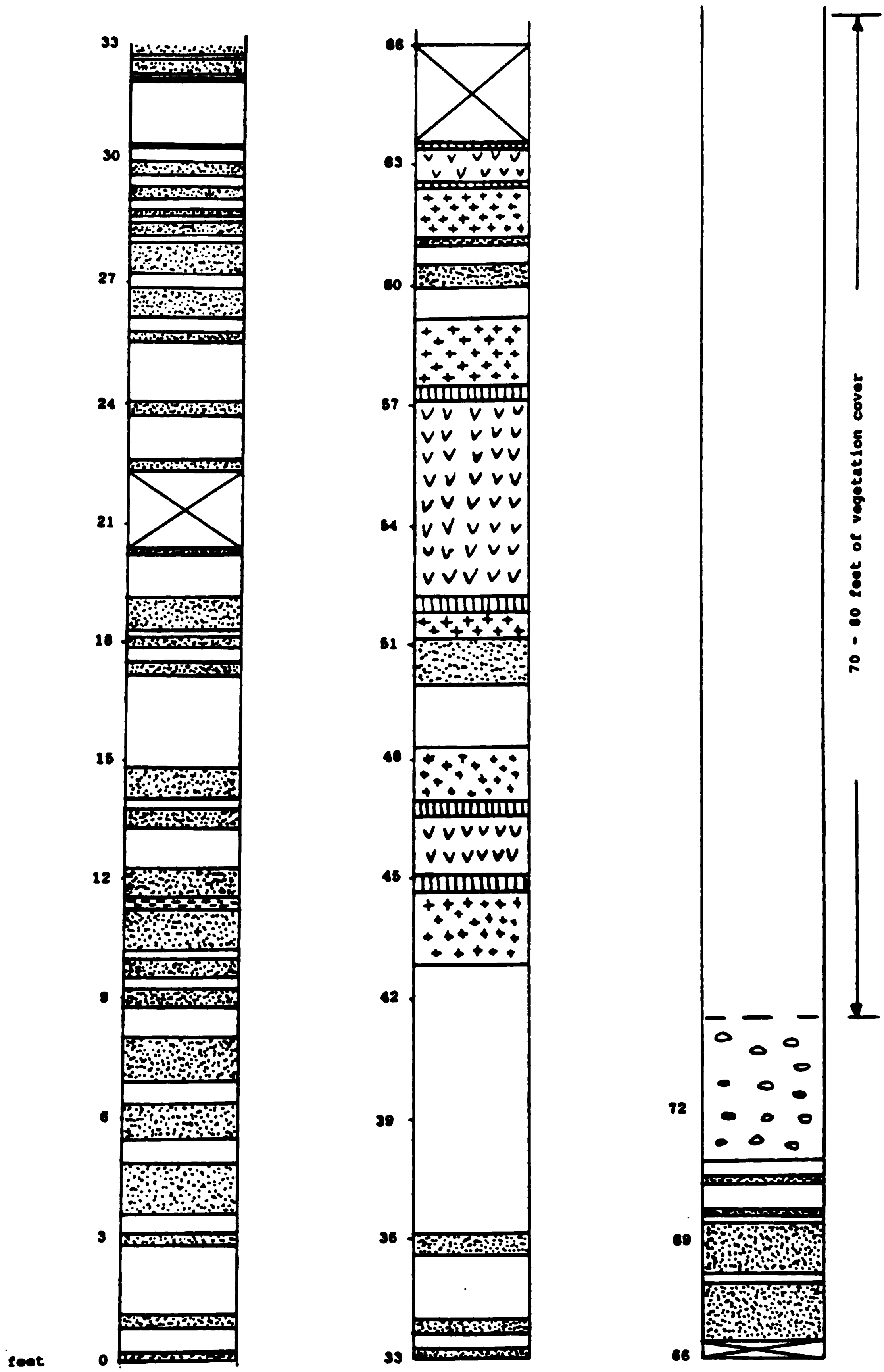

Massive dolostone richer in clay and feldspar


Central chert member

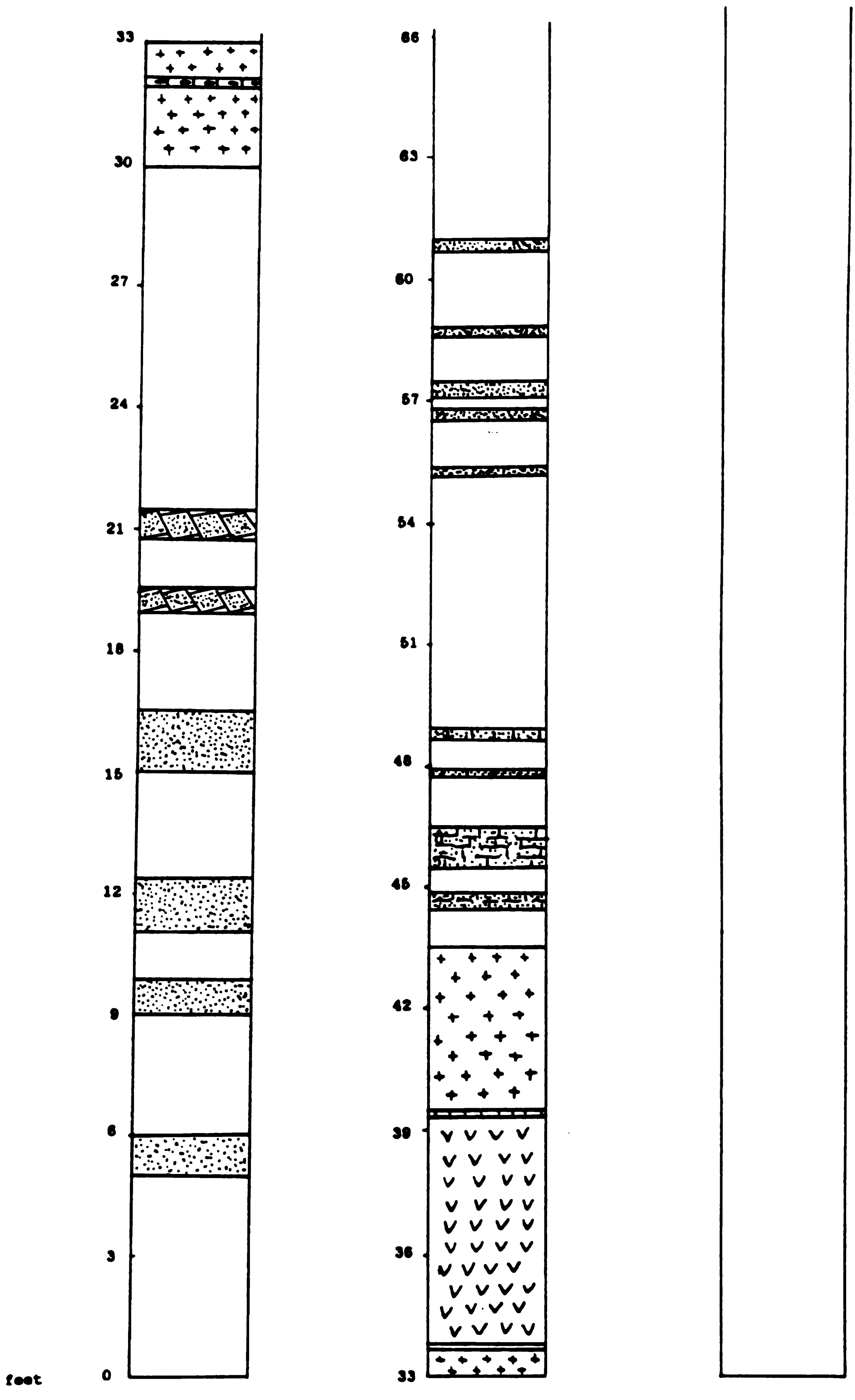

Dolostone


Quartz arenite


Reduction spots



feet
0
3
6
9
12
15
18
21
24
27
30
33
36
39
42
45
48
51
54
57
60
63
66
69
72
70 - 80 feet of vegetation cover


33
30
27
24
21
18
15
12
9
6
3
0
feet
66
63
60
57
54
51
48
45
42
39
36
33

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