lakes and ponds considered or could indicate that between 10,000 and 8,000 years B.P. in the Great Lakes region the climatic improvement occurred with a steeper temperature gradient than to the east and south.

The ^{18}O increase is paralleled by increasing $^{13}C/^{12}C$ ratios (Fig. 1). These $\delta^{13}C$ values could reflect a decreasing importance of aerobically decomposing organic matter and aquatic vegetation in the deepening Lake Erie, a relative increase of the exchange with the atmospheric CO_2 reservoir, or a more significant input of isotopically heavy CO_2 produced in a reducing environment within the bottom sediments (*4, 18*). Significant amounts of CH_4 are produced in Lake Erie sediments (*19*), and thus this latter possibility cannot be discounted a priori.

After the sharp increases in both ^{18}O and ^{13}C between 10,000 and 8,000 years B.P., the $\delta^{13}C$ values in the carbonates increased gradually and eventually attained a maximum close to ± 0 per mil in the surface sediments (Figs. 1 and 2). Changes in the ^{18}O curve, however, are probably associated with changes in the water budget of Lake Erie. The ^{18}O concentrations decreased upward in small but clearly visible steps, possibly reflecting an increased rate of flushings and a decrease in the relative evaporation intensity with the increased deepening of Lake Erie. According to the pollen data (*20*), the decrease in ^{18}O by about 2.5 per mil around 6,000 years B.P. is correlated with increasing sedimentation rates and changes in sediment type at the time of increased flow through Lake Erie when Nipissing Great Lakes drainage was diverted from the North Bay to the Sarnia outlet (*21*). The ^{18}O shift could then be explained in terms of the increasing amounts of ^{18}O-depleted water passing from Lake Huron to Lake Erie at that time.

Unfortunately, with the exception of a few shell fragments in the surface sediments, we were unable to recover sufficient fossil material for isotope analyses from the strata above the 5-m level. Therefore a detailed discussion of younger shell samples is not yet possible. However, ^{13}C analyses of the organic matter in Lake Erie sediments revealed significant differences between old and modern organic carbon deposited, reflecting the increased "terrestrial" input from man's activities since forest clearance and settlement of the drainage basin (*22*).

This change in carbon input is possibly documented in some of the modern mollusks. Published isotope data (*5*) from a number of pelecypods and gastropods collected in Lake Erie have revealed considerably lower ^{13}C contents than observed in the modern samples in this study (Fig. 2). The lower values could be due to habitat differences since the mollusks were collected from areas characterized by significant inputs of isotopically light "terrestrial" carbon (*22*), whereas the samples analyzed in the present study were collected off the relatively nonpolluted north shore of central Lake Erie.

P. FRITZ
Department of Earth Sciences, University of Waterloo, Waterloo, Ontario, Canada
T. W. ANDERSON
Geological Survey of Canada, Canada Centre for Inland Waters, Burlington, Ontario
C. F. M. LEWIS
Geological Survey of Canada, Ottawa, Ontario

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8. All carbonate shells, including ostracods, were handpicked from the core material after the sediment had been carefully washed through a series of sieves. The separated samples were first treated for 6 to 12 hours with 5 percent NaClO solution and then thoroughly washed with distilled water and treated under vacuum with 100 percent H_3PO_4 to produce CO_2 suitable for mass spectrometric analyses. All isotope data are expressed in the conventional δ units (per mil) notation and refer to the Pee Dee belemnite standard (PDB) for ^{18}O and ^{13}C contents in the carbonate shells.
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Precambrian Paleomagnetism: Magnetizations Reset by the Grenville Orogeny

Abstract. *Paleomagnetic results from iron-rich metasediments folded during the Grenville orogeny (~1000 million years ago) indicate that stable remanence was acquired after folding as the rock cooled from above 615°C through 550°C. As well as imparting a new remanence, the thermal event destroyed any preexisting magnetizations. In a proposed model, virtually all Grenville paleomagnetic poles have been reset and can provide no evidence of plate tectonic processes before 1000 million years ago. The polar sequence suggested by potassium-argon age trends within the Grenville province does, however, indicate rapid drift of North America between 1000 and 900 million years ago.*

Metamorphic rocks pose special problems for the paleomagnetist. Frequently, the natural remanent magnetization (NRM) is weak and unstable, and its direction may have been deflected by the magnetically anisotropic fabric of the rock. A more subtle problem is dating the NRM within limits that are not intolerably broad. Potassium-argon and rubidium-strontium ages can be reset during a metamorphic event if the peak temperatures are of the order of 250° to 350°C and 700° to 800°C, respectively (*1*). Furthermore, the NRM itself may be partially or totally reset. The heating accompanying a major metamorphic event may be very prolonged for deeply buried rocks. A single event may involve more than one heating episode, especially in the presence of igneous activity. Rocks will generally be exposed to high pressures as well as high temperatures, and the original magnetic minerals will be altered.

Following metamorphism, the reset NRM may comprise one or more generations of thermoremanent magnetization, viscous partial thermoremanent magnetization, or crystallization remanence. Nor is the surviving fraction of the original NRM, if any, necessarily primary. It may itself be a magnetization reset in a prior orogenic event. In view of these complexities, understanding of the nature of the NRM is all-important in deciding whether the appropriate age of magnetization is one reset at low temperature (K-Ar), one reset at high temperature (Rb-Sr), or the crystallization age (whole-rock Rb-Sr or zircon U-Pb).

A case in point is the Grenville structural province of the Canadian Precambrian Shield. Most Grenville rocks have undergone amphibolite to granulite grade metamorphism through being buried during the Grenville orogeny (~1000 million years ago) at depths to 20 km, where they were heated to 600° to 800°C. Uplift and cooling were slow, judging by the difference of nearly 150 million years between reset K-Ar and Rb-Sr mineral ages. Older Rb-Sr and U-Pb dates (*2*) and structural trends (*3*) in some areas record the effect of the Hudsonian orogeny (~1800 million years ago). Occasionally, structures resulting from the Kenoran orogeny (~2500 million years ago) are recognized (*3*) near the Grenville front (Fig. 1), the boundary between the Grenville and the neighboring Superior province. The Grenville front is in places an actual fault, and is everywhere a striking structural, metamorphic, and isotopic age boundary (*3*).

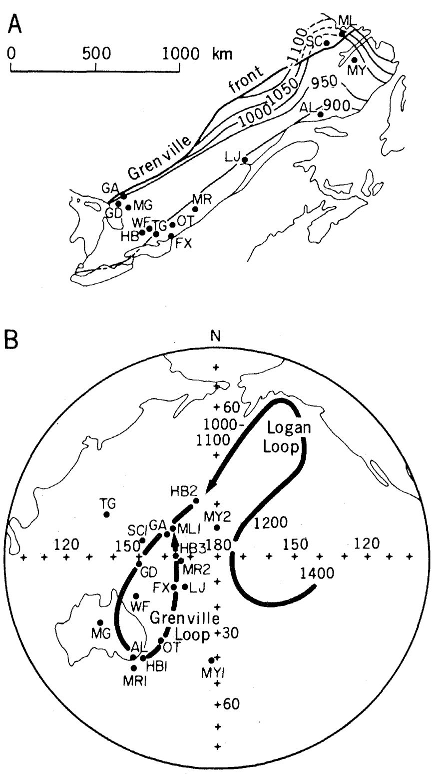

Fig. 1. (A) Thermochron map of the Grenville province, after Harper (*20*). Contour values are K-Ar ages in million years. Formations sampled for paleomagnetism: *ML*, Michael gabbro (*15*); *SC*, Seal-Croteau rocks (*23*); *MY*, Mealy Mountain anorthosite (*7*); *AL*, Allard Lake anorthosite (*21*); *LJ*, Lac St. Jean anorthosite (*7*); *MR*, Morin anorthosite (*8*); *OT*, Ottawa intrusions (*24*); *FX*, Frontenac axis dikes (*25*); *TG*, Tudor gabbro (*26*); *HB*, Haliburton intrusions (*6*); *WF*, Wilberforce pyroxenite (*26*); *MG*, Magnetawan metasediments [(*12*) and this report]; *GD*, Grenville dikes (*27*); *GA*, Grenville Front anorthosites (*26*). (B) Paleopoles for the rock units in (A). The numbers 1, 2, and 3 denote distinct NRM components from a single unit. Numbers along the curves are ages in million years. The polar wander curve from 1400 to 1050 million years is well established (*4, 5*). The younger Grenville loop is hypothetical and assumes a time sequence of poles suggested by the thermochron map.

Paleomagnetic data from rocks of the Grenville province (Fig. 1) are divided into two general groups. Paleopoles north of 15°S (group B) are possibly consistent with the late Precambrian polar wander curve [the "Logan loop" (*4*)] established for other structural provinces of the Shield (*5*). Poles south of about 25°S (group A) are clearly discordant with results from elsewhere in the Shield. (It is uncertain to which group poles TG and WF should be assigned.) Superimposed, directionally distinct components of NRM, of the sort one might anticipate in rocks with such a complex metamorphic history, have been reported in only four studies (*6–9*). In each case, one paleopole falls in group A, the other in group B. The superposition of A and B components in the same rock is evidence that groups A and B do not represent scatter about some single average pole (*10*), but record pole positions at two quite different times.

Three models could explain the discordant Grenville poles. One (*6, 7, 11*) supposes that plate tectonics operated in Precambrian times and attributes the Grenville orogeny to plate convergence within the Grenville province. The other two models propose a "Grenville polar loop" (Fig. 1), either predating (*8, 11*) or postdating (*11, 12*) the Logan loop, reflecting drift of the entire Shield but (as yet) unrecorded from rocks outside the Grenville province or its chronological and probable lithologic equivalent in the Baltic Shield (*13, 14*). The models stand or fall according to the dates assigned to the A and B poles. For example, the two-plate model makes sense only if the B poles marking the proposed polar juncture correspond in time to the Grenville orogeny, while the A poles are older and "see through" the event.

We report here the results of a fold test of the NRM of an iron-rich metasedimentary formation near Magnetawan, Parry Sound District, Ontario (45°45'N, 79°40'W), about 85 km south of the Grenville front. The iron formation (Fig. 2) occurs near the boundary between amphibolite (possibly a metamorphosed basic sill) and paragneiss in an area of upper amphibolite grade metamorphism. Geological mapping is poor, but the fold is outlined by three pronounced negative magnetic anomalies, corresponding to steeply dipping east and west limbs and a south limb offset by a postfolding fault. The age of folding is Grenvillian rather than Hudsonian, according to Wynne-Edwards [table 1 in (*3*)]

Figure 2 shows that the fold test is negative. The mean magnetization directions of the three limbs are statistically indistinguishable. Incorporated in the limb means are cleaned magnetization directions of the 83 samples (23 sites) that remained stable throughout stepwise alternating field demagnetization to 1000 oersteds. The average of the 23-site mean directions is declination 302°E, inclination –73°. The result is statistically highly significant ($k = 51$, $\alpha_{95} = 4°$) and defines a group A paleopole at 130°E, 24°S.

This result and a similar one from the nearby Whitestone anorthosite (*9*) are the first indications of A magnetizations within 100 km of the Grenville front. On any

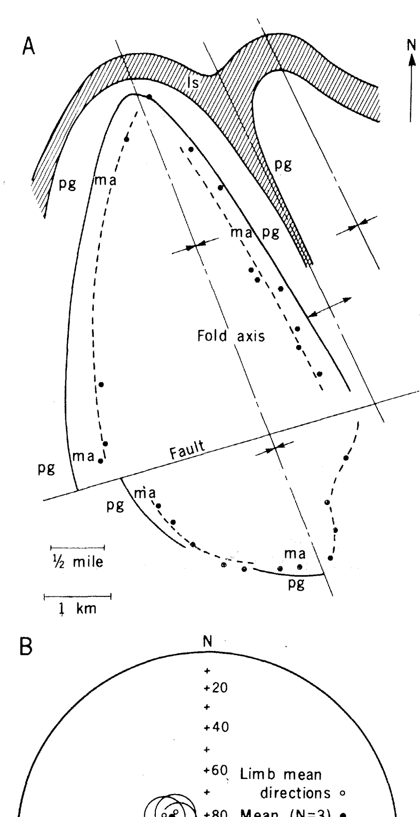

Fig. 2. (A) Map of the sampling area, with sampling sites indicated by dots and magnetic anomaly axes by dashed lines. Rock units: *ls*, crystalline limestone; *pg*, paragneiss; *ma*, amphibolite. (B) Equal-area projection of the mean magnetization directions of the three limbs of the fold, with circles of 95 percent confidence. The grand mean falls within all three circles. All inclinations are negative (upward).

two-plate model based on discordant groups of poles, the relict suture between the former Grenville plate and the remaining Canadian Shield must lie in the 100 km separating the Grenville front and the Magnetawan-Whitestone sites (Fig. 1) because (i) the front itself is not a suture (*3, 7–9, 15*), so the suture must be south of the front within the Grenville province, and (ii) group A poles are to represent magnetizations acquired by the Grenville plate before its collision with the remaining Canadian Shield, that is, while the Grenville plate was situated thousands of kilometers distant. However, there is no evidence in the admittedly poorly mapped region between Magnetawan and the front of even a cryptic suture (*16*), and elsewhere in the Grenville province highly metamorphosed equivalents of the pre-Grenvillian sediments of the Labrador Trough are recognizable more than 100 km south of the front (*17*).

The fold test itself provides the most compelling evidence against a two-plate model. The A magnetization in such models records the preorogeny situation of divergent plates, but according to the fold test the A remanence in our rocks was acquired after folding, itself of Grenvillian age (*3*). Other group A poles are presumably of similar age.

A lower limit to the age of magnetization follows from the observation that the A remanence in our rocks is carried by magnetite and hematite with blocking temperatures between 550° and 615°C. [The blocking temperature (*18, 19*) is the temperature at which thermoremanent magnetization is "frozen in" during cooling. Except in the case of very fine grains, it is only 10° to 20°C below the Curie temperature of the mineral involved.] Although the blocking temperatures in themselves are consistent with hematite (Curie temperature 670°C) being the sole carrier of remanence, the NRM intensity is too high (greater than 10^{-2} emu cm^{-3}) to be explained by the observed hematite content. Magnetite (Curie temperature 585°C) accounts for 80 to 90 percent of the saturation magnetization of these rocks and evidently carries a substantial fraction of the NRM as well.

In the absence of a plausible process that would generate magnetite and hematite simultaneously at low temperatures, crystallization remanence can be discounted. The A remanence was almost certainly acquired thermally, as either a thermoremanent or a viscous partial thermoremanent magnetization. The heating event in question can hardly have been more recent than the regional K-Ar age of 950 million years, for the peak temperatures required to reactivate blocking temperatures of 550° to 615°C, even for a very prolonged heating, must have been high enough to reset the K-Ar ages. The A remanence probably dates from uplift and slow cooling following the Grenville orogeny. To judge by the high grade of metamorphism, the temperatures experienced during burial exceeded the Curie temperatures of both magnetite and hematite and thus totally erased any previous magnetic memory.

Irving *et al.* (*11*) have observed that where A and B components are superimposed, the A component is generally carried by hematite and the B component by magnetite. Provided the cooling history was simple, the relative blocking temperatures of these minerals imply that A components (and A poles in general) are older than B components and their corresponding poles. Our results indicate that this criterion is too simple-minded, for both minerals carry an A remanence in the Magnetawan rocks. The detailed blocking-temperature spectra, reflecting the processes of magnetization as well as mineralogy, must be compared to establish the relative ages of the components.

We now suggest an alternative criterion that may make it possible to establish not simply the relative ages of two general groups of poles, but the relative ages of individual poles within these groups (including some poles from rocks north of the front in areas probably affected by the Grenville heating event). The proposed method is based on Harper's (*20*) thermochron map (contour map of K-Ar ages or of the activation temperature for Ar diffusion) for the Grenville province, shown in Fig. 1. Although K-Ar ages are undoubtedly younger than magnetization ages, the relative ages of the various poles should follow the same pattern as the thermochrons, provided the magnetizations were all acquired over broadly similar temperature intervals.

Although this condition is at best a rough approximation, applying the criterion to published Grenville poles (Fig. 1) leads to a reasonably consistent Grenville loop in the North American polar wander path. Poles SC1 and ML1 (both from north of the front), GF, and GD define the older end of the loop. The Magnetawan pole (MG) of this report falls on the descending branch. The Haliburton intrusions (*6*), Morin anorthosite (*8*), and Allard Lake anorthosite (*21*), spanning about 1600 km from west to east but lying near the same thermochron, yield poles HB1, MR1, and AL at the bottom of the loop. Finally, poles LJ and FX define the ascending branch of the loop.

No attempt has been made to fit poles of superimposed B components into this scheme because they are known (*6, 8*) to have been acquired at relatively low temperatures. With this exclusion in mind, only poles TG and MY1 definitely do not fit the pattern. Recent K-Ar isochron studies of the Tudor gabbro (*22*) lend support to the hypothesis that pole TG represents a much younger magnetization and thus would not be expected to follow the proposed sequence. Pole TG also does not fit any of the proposed models in the literature.

While the thermochron technique is an approximation, in that it substitutes a regional K-Ar date for the unknown magnetization age, it is nevertheless useful in the absence of more specific radiometric ages for most of the formations studied paleomagnetically. It is also unexpectedly successful, in that the polar sequence thus established, with two exceptions, defines a self-consistent Grenville loop (Fig. 1).

At this early stage of our knowledge, the existence of a Grenville loop in the North America polar wander curve appears to be the favored hypothesis to explain discordant poles from the Grenville province. To judge by the age of the Magnetawan A remanence, this loop must correspond closely in time to the uplift that followed the Grenville orogeny. The Grenville loop would therefore be younger than the Logan loop (*4*): it would span the approximate time interval 1050 to 900 million years ago. In tectonic terms, it would reflect rapid motion ($\geq$ 8 cm/year) of the entire North American craton, possibly but not necessarily related to the Grenville orogeny, and not relative motion of parts of the craton. A definitive test of this model must await improved radiometric dating of the Grenville province or the discovery of group A poles from rocks of another structural province.

M. O. McWilliams*
D. J. Dunlop

Geophysics Laboratory,
University of Toronto,
Toronto, Ontario, Canada

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* Present address: Research School of Earth Sciences, Australian National University, Canberra.

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^{18}O Changes in Foraminifera Carbonates During the Last 10^5 Years in the Mediterranean Sea

Abstract. *The Mediterranean response to major climatic events during the Upper Pleistocene could be seen as an integration of the principal phenomena particular to oceans and of regional phenomena peculiar to the Mediterranean Sea. The magnitude of oxygen isotope changes in foraminifera tests suggests that the temperature variations between stadial and interstadial periods could not exceed 11°C and that the correction factor for isotopic changes of waters should be about twice the value used for the oceans, or 2.7 per mil.*

The first data on oxygen isotope analysis in the Mediterranean deep-sea sediments were published by Emiliani in 1955 (*1*) for the core 189 from the eastern basin. In the present work some of the conclusions of an isotopic study made on foraminifera carbonates from four deep-sea cores from the Mediterranean Sea are presented (*2–6*). The locations of these cores are shown in Fig. 1: core 68 from the Alboran Sea, core KS05 from the Balearic basin, and core C_3 from the Ligurian Sea in the western part of the Mediterranean and core $V_{10\text{-}67}$ from the Ionian basin in the eastern part.

Species having relatively higher abundance throughout the cores were chosen for the analysis: *Globigerinoides ruber* in all four cores, *Globigerina bulloides* in core C_3, and right-coiling *Globigerina pachyderma* in core 68. Today the habitat of the first two species in the Mediterranean is restricted to shallower depths. The third species can migrate in the water column; however, this migration does not indicate that *G. pachyderma* has varying temperature requirements, for this species appears in greatest abundance in winter, when the water column in the Mediterranean is homeothermal (around 13°C, which is the winter minimum). Analyses were carried out on samples of 100 to 150 specimens; thus individual variations were averaged out. Samples with organic matter were cleaned by using a 24-hour treatment with Clorox, while samples with sediment filling were given a 30-second ultrasonic treatment. The carbon dioxide was released through the action of 100 percent phosphoric acid at 25°C. The $^{18}O/^{16}O$ ratios are reported as δ values (per mil) relative to the Pee Dee belemnite standard (PDB-1) (Table 1 and Fig. 2).

Sedimentation rates. Radiocarbon datings on core KS05 [C. Evin, in (*5*)], on core $V_{10\text{-}67}$ [Lamont-Doherty Geological Observatory and J. Thommeret, in (*2*)], and on sediments cored near core 68 (*4*) lead to the following estimations for sedimentation rates: (i) in core 68 sedimentation rates vary between 26 and 30 cm per thousand years; (ii) in core KS05 they vary between 10 cm per thousand years in the upper part and 15 cm per thousand years in the lower part (an average value of 12 cm per thousand years could be used); and (iii) in core $V_{10\text{-}67}$ the sedimentation rate ranges from 4.5 to 13 cm per thousand years. Core C_3 was dated by the uranium series method [Larmande-Renou and Martin (*7*)], giving an average sedimentation rate of 13.5 cm per thousand years. However, since no correction for detritic contamination had been done on the core, one must assume the rate is less.

In Fig. 2 a correlation represented by a dashed line is proposed for the 1.7×10^3 year "cold" isotopic peaks (peaks toward the positive values of δ) and the 5.5×10^3 year "warm" isotopic peaks (peaks toward the negative values of δ). Five major climatic stages can be distinguished in the last 10^5 years. These may correspond to the isotopic stages 1 to 5 of the generalized paleotemperature curve (Fig. 2).

$\delta^{18}O$ changes recorded by planktonic foraminifera (Table 1). Variations in measured paleotemperatures in planktonic foraminifera are caused both by changes in the ocean surface temperature and by glacially controlled changes in the isotopic composition of the ocean water. Most workers believe now that the last is the dominant factor (*8, 9*) and have even suggested that the existence of any residual temperature effect remains to be demonstrated. An isotopic change of 1.2 to 1.6 per mil is generally admitted for ocean waters between glacial and interglacial periods.

Isotopic fluctuations recorded by planktonic foraminifera from Mediterranean sediments within the period covered by isotope stages 1 to 5 amount at least to 2.8 per mil and can reach magnitudes of 5.2 per mil (Table 1). These variations seem extremely high when compared with that shown by Pleistocene foraminifera from Atlantic or Pacific cores: variations from 0.9 to 2.0 per mil for both oceans are reported by Van Donk (*10*). Isotopic changes

Table 1. Magnitude of isotopic changes for major paleotemperature peaks in cores 68, KS05, C_3, and $V_{10\text{-}67}$, reported as δ values relative to the PDB-1 standard. Abbreviations: B.P., before the present; *G. ruber*, *Globigerinoides ruber*; *G. bulloides*, *Globigerina bulloides*; and *G. pachyderma*, *Globigerina pachyderma*. Dashes indicate no results.

Species	Cores: 68		KS05		C_3		$V_{10\text{-}67}$	
	Depth of core (cm)	δ value	Depth of core (cm)	δ value	Depth of core (cm)	δ value	Depth of core (cm)	δ value
Age: 1.7×10^4 years B.P.								
G. ruber	380 to 250	≥ 3.3	180 to 10	4.3			58 to 35	4.1
G. bulloides	—	—	—	—	140 to 50	5.2	—	—
G. pachyderma	480 to 240	5.2	—	—	—	—	—	—
Age: 3×10^4 years B.P.								
G. ruber	—	—	—	—	340 to 300	2.8		
G. bulloides	—	—	—	—	360 to 300	3.9	—	—
G. pachyderma	—	—	—	—	—	—	—	—
Age: 5.5×10^4 years B.P.								
G. ruber	—	—	680 to 650	3.9	—	—	—	—
G. bulloides	—	—	—	—	720 to 660	4.3	—	—