Correcting distorted paleosecular variation in late glacial lacustrine clay

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Abstract

An undisturbed, horizontal chronostratigraphic marker horizon of laminated red glacio-lacustrine clay crops out over ∼25,000 km² in northern Ontario, Canada. The primary, clastic hematite laminae possess two stable vector components of magnetization. We sampled a 1.6 m vertical section at overlapping 2 cm intervals in cubic specimens (8 cm³, n = 106), precisely oriented in geographic coordinates which permitted measurement of inclination and declination. Alternating field demagnetization (12–17 steps per specimen) isolated a characteristic (ChRM), primary component (coercivity ≥40 mT) approximately 30° shallower than the mean geomagnetic field inclination at this latitude, with distorted paleosecular secular variation (PSV). A lower coercivity overprint (20–40 mT) is 6° shallower than the present geomagnetic field and similarly inclined to the depositional-remanence deflection when the clay was re-sedimented in the laboratory. We believe this angle to be representative of the inclination of the deflected primary remanence inclination during deposition, caused by the magnetic anisotropy of the clay. Using this as a proxy for the initial inclination, the ChRM (hard-component) inclinations were restored to their original values assuming vertical compaction, which averages to 51%. The restored inclinations are compatible with site-latitude and the restored secular-variation loops centre reasonably on the geographic North Pole. The structural correction based on homogeneous vertical shortening over-simplifies the reality of heterogeneous particulate flow in which grain rotations depend on shape, size and packing and effectively combines influences of compaction and depositional settlement. Nevertheless, this correction makes the PSV data more useful and interpretable in terms of paleopole migration. The more logical correction technique using anisotropy of anhysteretic remanence (AARM; Jackson, M.J., Banerjee, S.K., Marvin, J.A., Lu, R., Gruber, W., 1991. Detrital remanence, inclination errors, and anhysteretic remanence anisotropy: quantitative model and experimental results. Geophys. J. Int. 104: 95–103) is foiled here due to the high-coercivity of the remanence-bearing grains and their unknown orientation-distribution and shape-distribution.

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1. Goals and introduction

The most recurrent paleomagnetic problem in sediment and sedimentary rock is inclination-shallowing (Tauxe, 1993, 2005). It causes underestimates of paleolatitude, paleopole positions appear “far-sided” weakening paleogeographic reconstructions and secular-variation correlations. It may even cast doubt on the geocentric axial dipole (GAD) hypothesis for ancient rocks. In young strata and archeological contexts paleomagnetic signals are usually sufficiently detailed that the details of paleosecular variation are preserved as oscillatory variations in inclination and declination, although the latter is not always retrievable. Post-glacial lacustrine sediment is a popular source of young PSV data since they occur in stable depositional conditions that have recorded geomagnetic directions continuously over the last 30 Ka, with organic inclusions amenable to 14C dating. PSV declination and inclination is typically acceptably uniform over small regions (10,000 km²) and is cyclical with periods of 600–1000 a, usually with a clockwise trend of inclination (y) plotted against declination (x) (Bullard, 1948; Thompson and Oldfield, 1986). Its regional and temporal variation is due to heterogeneous geodynamo behaviour in the outer core (Hagge and Olson, 1989; Lund, 1989; Lund et al., 1988). PSV provides a stratigraphic correlation tool (e.g., Breckenridge et al., 2004; Creer and Tucholska, 1982; Mothersill, 1988, 1985, 1979; Zhu et al., 1998) and an archeomagnetic age-determination tech-
nique (Barendregt, 1995; Sternberg, 1997), with applications in paleoclimate studies (Benson et al., 1998; Evans and Heller, 2003).

Most lake-studies use sub-aqueous coring techniques that lose reference for absolute datings, generally involve some disturbance of the sediment fabric and which sample material that is too wet for certain “rock”-magnetic tests. Lake sediment has the advantage of providing continuous cores which may represent tens of thousands of years and dozens of PSV cycles that may permit inter-regional PSV correlation (interalia, Creer, 1985; Lund, 1989; Lund and Banerjee, 1979, 1983, 1985; Lund et al., 1988; Turner and Thompson, 1981). In this region, the chronological value of laminated glacio-lacustrine clay has long been acknowledged (Antevs, 1951; Banerjee et al., 1979; Rittenhouse, 1934).

Instead, we sampled a dry land section through 1.6 m of a paleo-lake red clay stratigraphic marker horizon. Thus, both declination and inclination are known for 2 cm$^3$ specimens overlapping at 2 cm intervals. The stiff and stable clay specimens were just as suitable as hard-rock specimens for repeated demagnetization and other rock-magnetic tests.

The laminated red clay unit we studied is an unusually well-defined stratigraphic marker horizon varying in thickness from 10 to >150 cm, over an area of ≥25,000 km$^2$ (Fig. 1). It is a chronostratigraphic marker of uniform geochemical signature (Lemoine and Teller, 1995; Leverington and Teller, 2003; Teller and Thorleifson, 1983; Thorleifson and Kristjansson, 1993; Zoltai, 1963). It is a primary feature since the red pigment is finely laminated and bedding-parallel and regionally extensive. The clay accumulated in eastern paleo-lake Agassiz, west and south of the Hartman moraine and Dog lake moraine as paleo-lake Kamanistikwia drained into paleo-lake Agassiz. The only two available $^{14}$C ages (~9900 BP) from the clay may be difficult to reconcile with what appears to result from the much younger drainage of paleo-lake Kamanistikwia (Fig. 1). Moreover, it is difficult to understand how such an ancient site was uniquely calibrated without a corroborative technique since only two dates are reported from the oscillatory sequence of carbon activity ratios. The ranges of thicknesses of the laminated clay reported in the literature (5–61 cm) appear to be minima from our re-investigations; our site crops out at an elevation of ~147 m, near Strawberry Creek, 30 km WNW of Thunder Bay where >200 cm are present (Fig. 1). We were able to sample a 160 cm vertical section from a hillside with sufficient care to avoid disturbance of the fabrics.

2. Magnetization of lake sediment

The permanent magnetization of lake sediment is achieved by complex processes known by the umbrella terms depositional remanent magnetization (DRM) and post-depositional remanent magnetization (PDRM). Some authors appear to include vertical compaction under PDRM although these usually involve different micro-processes. It may be wiser to separate their mechanical principles despite their chronological overlap. Of all the remanence-acquisition processes in nature, DRM–PDRM are most difficult to document. Commonly, indeterminable hydrodynamic, mechanical-electrostatic, magnetic and diagentic mechanisms conspire to produce a remanence vector-sum that obscures analysis of its components and their origins, even after incremental demagnetization. Usually, as in this study, thermal demagnetization is not possible due to the impermeability and moisture content which prevents successful impregnation with a bonding agent.

Successful studies of controlled laboratory sedimentation have provided invaluable empirical data on the preservation of the laboratory magnetic field (e.g., Carter-Stiglitz et al., 2006; King, 1955; Jackson et al., 1991; Tan et al., 2002; Tan and Kodama, 2002, 2003; Tauxe and Kent, 1984). However, they cannot predict the processes and critical sediment-characteristics in natural sedimentation and the authenticity with which an unknown paleofield direction is retained.

As a broad and oversimplified generalization, merely for the benefit of introduction, DRM–PDRM and compaction processes reduce inclinations, scatter declinations and mostly, but not exclusively, reduce magnetic intensities with respect to those values expected or present at the moment of deposition. Understandably, we have all simplified the response-mechanisms to homogeneous and continuum-mechanics processes that may be formulated mathematically.

DRM processes are almost certainly most complex and variable in marine, fluvial and subaerial environments although some work has been attempted (Tauxe and Kent, 1984; Tan and Kodama, 2002, 2003; Rosler and Appel, 1998). In contrast, lacustrine sediments are more productive for paleomagnetism since depositional rates are more stable, current variations more subdued than in other environments and post-depositional disturbances are fewer. Moreover, seasonal environmental variations may be beneficial chronostratigraphically. The thinly stratified unit of geochemically uniform, red clay laminations over a widespread area (Fig. 1) suggest a microclimatic influence.

Good introductions to DRM–PDRM are found in Butler (1993), Tarling and Turner (1999), Tarling (1983) and Verosub (1977). Specialized treatments, relevant to the relatively simple lacustrine environment, repeatedly focus on the restoration of inclinations that are too shallow for the causative geomagnetic field. The least equivocal studies focus on Holocene lake or deep-marine sediment rather than on more complicated sedimentary environments or on sedimentary rocks for which tilt-corrections, finite (tectonic) strain-corrections or paleolatitude-corrections may have been required. Most inclination-shallowing studies offer some functional relationship like

$$\tan(I) = f(\tan(I_0))$$  \hspace{1cm} (1)

Here $I$ is the observed remanence-inclination and $f<1$. $I_0$ is the “original” remanence inclination, or the inclination of the magnetizing field depending on the model of DRM or of compaction. The common presence of tangent ratios in DRM, PDRM and compaction models is not surprising since they share a geometry in which the vertical fabric-direction is reduced with no extensions parallel to bedding. It is more surprising that some models offer a linear, or otherwise simple monotonic function.
(f) Their success (as with similar approaches in structural geology strain studies) is probably because such simple functions suffice to approximate conflicting corrections at the required precision-level and for the limited range of values under consideration. Such simple functions cannot provide mathematical description of any of the constituent processes responsible for inclination-shallowing during and after grain-settlement. From structural geology, we know that the heterogeneous processes by which the remanence-inclination in grains is scattered and shallowed requires us to consider; variable amounts and rates of rigid-rotation of grains, grain-impingement, grain-size and shape, particulate flow and grain-interactions after settlement. In particular, particulate flow whether in sediment or metamorphic rock defies detection since grains switch neighbours usually without leaving evidence (Borradaile, 1981). The effects of hydrodynamic factors prior to settlement are more unpredictable.

Compaction in the sense of structural geology is simpler. Convention treats macroscopically compacted material as a homogeneous continuum with maximum shortening (Z) per-
pendicular to bedding, and zero strain parallel to bedding (i.e., $X = Y = 1.0$ where initially $XYZ = 1$). It is also assumed these effects are homogeneous at some convenient scale, in our case several cubic millimeters. Inclination-shallowing is then predicted by $\tan(I) = (Z/X)[\tan(I_O)]$. Of course, at some unsuitable scale, every petrographic examination in structural geology reveals the falsehood of homogeneous continuum-behaviour. The strain of any grain-assemblage requires particulate flow; the heterogeneous re-arrangement and neighbour-swapping of grains (Borradaile, 1981). However, for a suitably large specimen and the ratio $Z/X = V'/V_O$, where $V'$ and $V_O$ are the final and original volumes. Volume change may relate to porosity $\phi$ with a form $V_O - V' = a\phi^{bd}$ where $a$ and $b$ are lithology-sensitive constants and $d$ is the overburden depth (sediment + water/ice). Severe inclination-shallowing is observed in deep-marine sediment (e.g., Arason and Levi, 1990a,b) for which compaction ratios may be estimated from volume changes via the downcore porosity variation. Whereas the correction equation may be of the general form of Eq. (1), it depends on lithology, grain-size, grain-size variation and requires re-calibration in each study.

King’s (1955) pioneering DRM–PDRM studies re-deposited natural sediment in a laboratory setting, permitting comparison of the acquired remanence inclination with that of the laboratory field. Such studies offer empirically determined linear functions for Eq. (1), needing re-calibration for each new sediment type and each subtly different depositional process. Focused studies tackle more complex depositional environments (Anson and Kodama, 1987; Carter-Stiglitz et al., 2006; King and Rees, 1966; Tan et al., 2002; Tauxe and Kent, 1984). Our simple laboratory re-deposition experiments (Fig. 6) show that our disaggregated clay acquires a remanence $4^\circ$–$8^\circ$ shallower than the magnetic field. This is not attributed to a depositional phenomenon but rather to field refraction by the clay sediment’s anisotropy, perhaps revealed by elliptical Bingham confidence regions (Bingham, 1964).

Ideally, with perfect DRM-control, the sediment’s remanence is an expression of the magnetic grains’ fabric, an alignment due to the geomagnetic field. Thus, Jackson et al. (1991) showed that anisotropy of anhysteretic remanence (AARM) could be used to infer the strength of the DRM alignment process. Their correction for AARM anisotropy also steepened shallow vectors to their primary inclination. Jackson et al. showed that the bedding-parallel/bedding-normal AARM ratio may be used to approximately back-rotate the remanence-inclination to the paleofield inclination. Whereas this approach is successful in some sediment (Kodama and Sun, 1990, 1992; Sun and Kodama, 1992) it strictly requires that inclination-shallowing is due only to pre-compaction processes and it is preferably verified by re-deposition experiments. Of course, the AARM procedure is not technically feasible if the remanence-bearing grains have a wide range of coercivity, or if they include hematite. Unfortunately, “in many lake sediment studies, it is standard practice to blanket clean cores at an AF level determined by the response of a few pilot specimens. Under these conditions, it is doubtful if the primary NRM is isolated at all” (Dunlop and Özdemir, 1997, p. 440).

3. Sampling

3.1. Magnetic fabrics and sedimentary textures

Primary evidence of sedimentary fabrics may be relatively easily deduced from the anisotropy of low field magnetic susceptibility (AMS) (Hrouda, 1982; Tarling and Hrouda, 1993). Both silt-rich specimens and clay-rich specimens show minimum susceptibility normal to bedding (Fig. 1b) with maxima and intermediate axes in the bedding plane. Ninety-five percent confidence limits around the mean-tensor axes validate a preferred NW–SE axis for maximum susceptibility ($\sim$–mineral alignment) (Fig. 1c), parallel to paleocurrent flow that would be consistent with the paleogeography (Fig. 1a). That flow-axis persisted through >80 (annual?) rhythmites and throughout the overlying massive clay that may have accumulated over millennia (see below). The orientation of the mean-tensor has been calculated directly from the data and from the original data standardized to specimen-susceptibility (Fig. 1c). The latter suppresses the effects of outliers of unusually high mean susceptibility that may skew the orientation-distribution. The mean-tensor axes for non-normalized specimen values include the effects of subfabrics (or even individual grains) with susceptibility much greater than the matrix. That is of value because it indicates just how heterogeneous is the AMS and the importance or unimportance of large multidomain magnetic grains (Borradaile, 2001). When the specimen anisotropies are normalized by their mean value, each specimen’s AMS is reduced to a unit-volume ellipsoid and we see mainly the effects of the most numerous and moderately susceptible grains (i.e., clay). Both versions of the mean-tensor verify a NW–SE flow axis that corresponds to the orientation of the paleogeographically defined spillway connecting the former lakes Kamanistikwia and Agassiz (Fig. 1a).

Since AMS ellipsoid shapes blend contributions from different minerals present in different proportions, with different bulk susceptibilities, and different AMS anisotropies, they are generally inscrutable. For example, clay and hematite have high anisotropy but relatively low bulk susceptibility (respective $k < 1000$ and $< 10,000 \mu$SI). In contrast, magnetite has the highest bulk susceptibility of minerals ($k \sim 2.3$ SI) but most grains have low anisotropy. The balance of bulk susceptibility, anisotropy, relative abundance and differing orientations complicate interpretation of AMS ellipsoid shapes and even of orientations (Borradaile and Jackson, 2004). However, clay-rich specimens show the most oblate and most eccentric fabrics (Fig. 1d). This supports their strong and differential compaction that must affect the paleomagnetic vectors. Ising’s (1942a,b) pioneering work noted that glacial clays had magnetic fabrics as intense as slate. Recent research in metamorphic and structural fabrics explains the paradox that some sedimentary fabrics have higher degree of anisotropy ($P$) and even better alignment than some metamorphic fabrics (Hrouda, 1982; Borradaile and Jackson, 2004).

Bedding fabric controls AMS orientation but anisotropy of anhysteretic remanence (AARM; Jackson et al., 1991) better assesses the potential for paleofield-deflection since it isolates the response of remanence-bearing grains. ARM-susceptibility
varies with direction if remanence-bearing grains have a preferred alignment. Thus, we measured ARM intensities (in mA/m) caused by a weak dc bias field (in our case, 0.1 mT) during the decay of an initially large alternating field that triggers remagnetization. In our case, a peak alternating field of 160 mT sufficed to remagnetize every sample. The dc field was applied over a selected window of AF values so that the ARM contribution of a specific class of ARM-bearing grains could be isolated, for example multidomain versus single-domain magnetite. For each specimen, an ARM was applied along seven different axes to determine AARM, using the same specimen orientations we use in the measurement of low field AMS (Borradaile and Stupavsky, 1995). These experiments are not possible with hematite-rich material since its high-coercive component defies the AF demagnetization required to apply a laboratory ARM. However, hematite-rich specimens show an AAR fabric which is oblate and parallel to bedding.

3.2. Magnetic mineralogy

3.2.1. Lowrie three-axis test

Three orthogonal components of IRM were applied in distinct coercivity windows using three applications of IRM along three orthogonal axes with appropriate intervening three-axis AF treatments (modified version of test by Lowrie, 1990), using a Sapphire Instruments pulse magnetizer and AF demagnetizer. Several different ranges were used for each component (x, y, z); one example of treatment steps to give three orthogonal synthetic vectors is shown in Table 1.

We experimented with various coercivity intervals but those tabled isolated contributions from single domain (SD), pseudo-single-domain (PSD) and multidomain (MD) magnetite. This under-utilized test may be the most direct and reliable means of understanding the distribution of remanence between different domain structures of magnetite and different minerals; the actual coercivity boundaries should be adjusted to suit the samples in question and we achieved that in trials. Typical results show the importance of the 0–30 mT window for silts (PSD magnetite) and sub-equal importance of <30, 30–60 and >60 mT fractions in clay (Fig. 2). The very small contribution of higher coercivity hematite was isolated using a simplified two-axis version of Lowrie’s test and with <100 and ≥10 mT windows.

3.2.2. Microhysteresis studies

Using our Princeton Measurements MicroMag 2900 instrument, mineral samples (<0.01 g) were measured over a dc field range from +1.0 to −1.0 T. The determined hysteresis loops for clays and for silts were each similar so for the purposes of illustration we have stacked all the measurements from 35 silts and from 220 clay specimens in two separate composite graphs (Fig. 3f and g). Silts were more difficult to sample due to their friable nature and larger grain-size. (Silt was less abundant in the analyzed section, and because mineral samples for hysteresis experiments were collected at regular intervals in the outcrop this led to an under-representation of silt to clay samples in the stacked hysteresis loops.) A paramagnetic slope correction was required for all loops to yield a meaningful hysteresis loop with horizontal high-field plateaux. Hysteresis statistics for the silt loops and for the clay loops are presented in Table 2.

Remanence intensities are 10 times greater for silts, $H_{CR}$ is greater for clay and $H_C$ is lower (all conclusions significant at 95% level). These statistics are consistent with the hypothesis that clays carry significantly more remanence in finer magnetite grains (SD) and in hematite. Dunlop (2002a,b) showed that the relative importance of SD, PSD and MD magnetite is detected readily from plots of $M_{RS}/M_S$ against $H_{CR}/H_C$ (Day et al., 1977, here Fig. 3a–c). PSD behaviour dominates; note the consistent responses from the three trials from sub-sampling the same specimens. Squareness of the hysteresis loops, from plots of $M_{RS}/M_S$ versus $H_C$ also discriminate the responses of clay and silt (Fig. 3d). The slope of the silt data compares well with published magnetite data (Wang and Van der Voo, 2004) whereas the different slope of the clays, together with their less well-defined trend may be due to a mixture of magnetite and hematite. Even the simplest $H_{CR}$ versus $H_C$ plot can distinguish and silt responses (Fig. 3e); power law relationships give their strongest correlation. The clays have greater ranges in both $H_C$ and $H_{CR}$, despite their larger sample-size, indicating intrinsically greater variance in hysteresis properties. In contrast, the variance of silt hysteresis properties indicates a narrow distribution of hysteresis properties and thus a more restricted range of minerals and grain-sizes.

4. Natural remanent magnetization (NRM) and relevant properties down-profile

The natural remanent magnetization (NRM) of all specimens ($n = 78$) was measured using a Molspin magnetometer, which provided more than adequate sensitivity for all specimens which were contained in plastic specimen cubes (2 cm × 2 cm × 2 cm). Molspin precision is enhanced with cubical specimens, as

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<td>An example of one Lowrie three-component scheme</td>
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<td>Treatment steps</td>
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<td>Microhysteresis data</td>
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<td>Mean ± S.E.</td>
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<td>Silt ($n = 35$)</td>
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<td>Clay ($n = 220$)</td>
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Fig. 2. Lowrie’s (1990) experiment determines the potential contributions to NRM, partitioned according to their coercivity-distribution. Preliminary demagnetization of NRM and IRM are required to determine the coercivity spectrum. Synthetic IRMs are applied along three orthogonal axes, with suitable AF demagnetization between each IRM application so that a tri-component IRM is created with three orthogonal components with different, non-overlapping coercivity intervals. Subsequent demagnetization then verifies the respective potential of the three different coercivity fractions to carry remanence. (a and b) show typical three-component tests.

opposed to the routine cylinder shapes used in hard-rock paleomagnetism (Borradaile et al., 2006). NRM intensity (Fig. 4a) is compared with bulk low field susceptibility \((k)\) and with ARM-susceptibility (Fig. 4b and c). However, magnetite dominates throughout and ARM-susceptibility is high where fine-grained magnetite \((<1–0.05 \mu m)\) dominates the magnetite-distribution (e.g., Evans and Heller, 2003). Anisotropy of ARM (AARM) (Fig. 4d), increases somewhat erratically though the clay-rich part of the section. In the silty portion near the base, the ratio shows some lower values. Overall, the mean AARM bedding anisotropy has \(AARM_y/AARM_z = 1.52 \pm (0.13 \text{ S.D.})\) for 72 specimens. For those measurements the peak AF was 100 mT and the dc field of 0.1 mT was applied over an AF window from 60 mT down to 0.1 mT.

5. Isolation of significant components

The different vector components of NRM must be isolated by incrementally demagnetizing the NRM; in the case of our fragile sediment, the only practicable technique is alternating field (AF) demagnetization. The relative significance and relative age of differently oriented vectors requires interpretation of magnetic mineralogy and rock–magnetic properties and is less objective. This terrestrial exposure provided well-compacted specimens of lake sediment that could be cut and sub-sampled with a knife. Nevertheless, they are unsuitable for thermal demagnetization or low temperature demagnetization that disintegrates the material. Thus, we used static, three-axis AF demagnetization, with a six-orientation permutation of demagnetizations at each demagnetization level to minimize the possibility of gyroremanent artefacts (Stephenson, 1983) for which our software provides a warning. Our demagnetizer includes circuitry to detect and suppress spurious ARM.

Every cubical specimen \((n = 78)\) was exposed to at least 11 steps of AF demagnetization and re-measurement, from a peak alternating field of 2.5 up to 180 mT, in increments chosen to suit the progress of demagnetization in each sample. The Sapphire Instruments demagnetizer includes circuitry for the cancellation of spurious ARM. Typical intensity-decay curves for the magnetization are shown for silt and clay specimens, respectively, Fig. 5a and b, with their vector plots inset. For ease of identifying vector components, vector plots were
inspected visually during the demagnetization experiments as three-dimensional plots, rotatable on the computer screen for the purpose of selecting critical steps to choose principal components for which directions were calculated within the Molspin software (SPIN05.exe). Clay-dominated specimens generally show two or three vector components of NRM which fall into soft, intermediate or hard coercivity intervals (0–10, 10–40 and >40 mT). Occasionally, overlapping coercivity spectra may smear the transition between the directions of the intermediate and hard remanence vectors (see clay in Fig. 5b) but sufficient demagnetizations steps were always available to isolate the vector-orientations.

6. Re-deposition experiments and laboratory acquired DRM

We estimated the DRM potential of the clay by re-depositing disaggregated clay in still water. Outcrop clay was disaggregated to a slurry in an ultrasonic bath and plastic sample cubes in daily increments, filling the cubes progressively over 14 days. The cubes were fixed in orientation to the laboratory magnetic field. Drainage holes in the base of the cubes permitted the samples to stiffen and permit handling. All specimens magnetized close to the laboratory declination but with inclinations approximately 4–8° shallower than the laboratory field (+74°)
Fig. 4. Stratigraphic profiles: (a) NRM ($n = 106$). (b) $k$, low field bulk susceptibility ($n = 106$), from mean of seven measurements per specimen. (c) Anhysteretic remanent susceptibility (ARM) ($n = 78$). (d) Anisotropy of ARM, parallel/normal to bedding ($n = 78$).

Fig. 5. Typical alternating field (AF) incremental demagnetization of natural remanent magnetizations (NRM) for (a) silt and (b) clay. Clay and silt invariably shows at least two vector components although the lowest coercivity component in silt may be unstable. Most NRM intensity is lost in a similar coercivity range for both silt and clay but the intensity decay of silt is more characteristic of magnetite.

Fig. 6. The clay was disaggregated and added incrementally to water in sample cubes over a period of 10 days. Exposure to the laboratory field resulted in a DRM inclination reduced by clay alignment. (a) For large-specimen cubes, measured in Molspin’s “BigSpin” magnetometer. (b) For conventional small plastic specimen cubes.

Fig. 7b). The experiment used four small standard (8 cm$^3$) cubes and four large standard (125 cm$^3$) cubes; the latter requiring use of the new “BigSpin” large-sample Molspin magnetometer. This replicates the simplest aspect of settlement; in nature other processes might complicate this DRM orientation but it is valuable as limiting value for the de-compaction restorations, below.

7. Characteristic remanent magnetizations and secular variation (SV)

The hard-component ($\geq 40$ mT) is present in 97% of the samples and the intermediate component (20–40 mT) is present in 53%. Their angular-mean inclinations are 36.5$^\circ$ and 62$^\circ$, respectively, compared with the GAD inclination of 66$^\circ$ at this latitude which is expected to be the long-term average paleofield inclination. At this site, the hard-component mean inclination (36.5$^\circ$) is too low for the present GAD inclination (66$^\circ$ Fig. 7b), and even for any reasonable Quaternary pale-
ofield inclination (\(\sim 77–50^\circ\)). Its declination variation (320–40°) is similar to that of NRM-declinations of modern lake sediment cores in periglacial regions (Creer and Tucholka, 1982; Breckenridge et al., 2004; Zhu et al., 1998), including this region (Lund and Banerjee, 1985; Mothersill, 1985, 1988). Under the assumptions of continuum mechanics of a homogeneous medium, used in simple structural models, vertical compaction should only deflect the inclination of any maker, leaving its declination unchanged. However, any real aggregate is subject to the heterogeneous effects of particulate flow (Borradaile, 1981), in which grains jostle around one another in unpredictable paths. Thus, whereas there may be no bulk finite strain in the horizontal (bedding plane), at the grain-scale, irregular motions of elongate minerals ensure that they will change declination as well as inclination. Thus, bulk homogenous compaction cannot occur without grain-scale heterogeneous strain which will cause the remanence-inclination to be deflected rather systematically whereas the declination will be less predictably scattered.

The intermediate component (20–40 mT) orientations have a different temporal pattern from the hard-component and show \(\sim 4^\circ\) inclination-shallowing which we attribute to a younger viscous overprint, refracted by the anisotropy of the compacted clay.

Although the stratigraphic plots (Fig. 7a and b) suggest PSV structure in the hard coercivity components such time-series conceal the structure of the orientation-distribution. The dimensionless Quantile–Quantile (Q–Q) plot (Fisher, 1993; Fisher et al., 1987) addresses and isolates orientation-distribution variation, independent of the time attribute. This is an important and different way of evaluating secular-variation data. Samples of orientations measured at different instants in time are treated as follows. For each sample, successive orientations \(\theta\) are rearranged into magnitude-order from value 1 (minimum) through
value \( n \) (maximum) as a new array \( \theta^* \). The re-arranged sample is then plotted as \([x, y]\) pairs as follows:

from \[ \left[ \frac{1}{n+1}, \theta^*_1 \right] \text{ to } \left[ \frac{n}{n+1}, \theta^*_n \right] \]  

(2)

The \( x \)-axis \([i/(n+1)]\) is the Uniform Quantile and the \( y \)-axis is the Sample Quantile \( (\theta^*) \). Symmetrical preferred orientations track across the 45° line through the data-centroid, and preferred orientations are readily detected and compared. The Q–Q plot verifies the sensible preferred orientation of the hard-components of NRM (Fig. 7c). In contrast, intermediate components are clearly unrelated to the primary magnetization (hard-component), and poorly concentrated (Fig. 7d). The preferred-orientation distribution of the hard-component is thus well defined and contains convincing evidence of an underlying PSV oscillation.

8. AARM correction for DRM alignment

Jackson et al. (1991) argued that since paleofield magnetic forces align the remanence-carrying particles, the anisotropy of remanence in the sediment must quantify that effect. AARM is a better estimate of the orientation-distribution of the “ferro”-magnetic grains (Jackson, 1991) although IRM anisotropy has been used cautiously as a DRM-correction tool where it was necessary using high-coercivity samples (Tan and Kodama, 2002, 2003). Jackson et al. supported their work with laboratory calibration, offering an attractive and intuitively fundamental correction that is widely applicable in sediments and some sedimentary rocks. Since actual grain-shape anisotropy and orientation-distribution is usually indeterminable, this may be simplified to

\[
\tan(I) = \frac{Z}{\tan(I_0)}
\]  

(3)

This relationship is actually only precise for the rotation of a passive linear marker (i.e., a strain marker with identical physical properties to a completely homogenous matrix). However, Eq. (3) deviates little from the relationship describing the rotation of a rigid marker with high aspect ratio (Borradaile, 1992). Consequently, versions of this relationship succeed in many studies of the rotation of remanence-bearing accessory minerals in sediment (Anson and Kodama, 1987; Deamer and Kodama, 1990; Jackson et al., 1991; Kodama and Davi, 1995; Kodama, 1997; Kodama and Sun, 1990, 1992).

\( Z \) and \( X \) describe axes perpendicular and parallel to bedding; \( \text{ARM}_X \) and \( \text{ARM}_Z \) are the appropriate magnitudes. (The structural geology convention for finite strain axes is \( X \geq Y \geq Z \).) However, this correction procedure is only valid for perfect passive behaviour of grains; this correction is invalid for the rotation of tectonically aligned grains which align by much more complex processes involving grain-reshaping by steady-state diffusive mineral transformations. This is a specimen-level correction and from the AARM determinations, the \( \text{ARM}_Y/\text{ARM}_X \) ratios (Fig. 4d) have been applied to the hard-component inclinations (Fig. 9c). The corrected data is far noisier than the original measurements, reflecting the inappropriateness of a DRM-correction in our case, mainly due to the fact that our samples contain high-coercivity grains which cannot be successfully demagnetized between magnetizations steps in the determination of AARM. The technique is very successful in sediments in which remanence is carried by low coercivity magnetite, for which AARM is easily determined.

9. Correction for continuum-compaction

Sine Jackson et al.’s (1991) technique is thwarted by the difficulties of determining AARM, we are forced to correct for inclination-shallowing directly by correcting for compaction. The initial inclination \( (I_0) \) is represented by the inclination acquired in the laboratory re-sedimentation experiments. The homogenous continuum strain model predicts the reduction of inclinations of passive lines from \( I_0 \) to \( I \) according to

\[
\tan(I) = \frac{Z}{\tan(I_0)}
\]  

(4)

Here vertical compaction is 100\( Z \)% due to volume reduction without lateral strain \( (X = Y = 1 > Z, \text{ initially } XYZ = 1) \). Of course, the assumption that the passive line model is applicable to the minerals grains, much less the magnetic moments that they carry, is an optimistic approximation (q.v., Borradaile, 1981, 1993a,b, 1997) but this does not limit the usefulness of the model to provide boundaries to data-evaluation. The assumptions usually lead to an underestimate of compaction.

To better appreciate the possible effects of compaction, we review the orientation-distribution of the data, independent of the time-domain (Fig. 8). All orientation-distributions (declinations or inclinations) present the particular problems of a closed range and presence of an antimode. Von Mises’ distribution (Fig. 8a), a two-dimensional equivalent of Fisher’s (1953) unimodal-concentration on the sphere, illustrates this simply. Clearly variance/dispersion is a concept of limited value (Fig. 8b) that is best replaced by an inverse variable, concentration \( (\kappa, \text{ comparable to Fisher’s } k; \text{ Fisher, 1953}) \). Simple inspection of circular plots and \( \kappa \) are not very helpful; a rarely used improvement is the cumulative summation frequency distribution (CUSUM plot; Fisher, 1993), Fig. 8c. CUSUM plots simply compare angular distributions, regardless of other attributes, such as time. For simple circular distributions these are symmetrical with vector mean parallel to mode (Fig. 8c).

CUSUM plots reveal some features not otherwise readily apparent in our PSV data (Fig. 8d and e) show strong concentrations and other features. Focusing on the hard-component, the dominant part of the NRM, declinations clearly smear westward but the distribution still has a mean close to North. Inclinations show a nearly symmetrical distribution about a mean of 36°, well below the GAD inclination (67°). Clearly the intermediate component inclinations seem unaffected or less affected by compaction; they are not so symetrically distributed but have a mean about 8° lower than the GAD for this latitude. It is logical to conclude that the hard-component is original and inclination-shallowed by compaction; and that the inter-
mediate component is a viscous overprint, deflected 6–8° by the clay’s magnetic anisotropy. The laboratory re-sedimentation experiments corroborate that the clay’s magnetic anisotropy deflects an acquired remanence with an inclination reduced by 6° (Fig. 6).

Thus, the intermediate component provides proxy values of the inclination of the original magnetization ($I_0 = 56°$), including some anisotropy refraction. The hard-component inclinations represent the compacted primary remanences now with mean inclination, $I = 36°$. These values are substituted in Eq. (4) to determine a de-compaction function. Of course, this assumes uniform homogeneous compaction behaviour for all samples; consequently outliers of declination or inclination persist after we apply this de-compaction correction (Fig. 9a and b). This is clearly a more systematic and sensible correction than the AARM correction method (Fig. 9c), which fails in this sequence due to the presence of hematite.

Fig. 8. Orientation-distribution of declinations and inclinations. (a) Von Mises’ distribution commonly models preferred orientations on a plane, comparable to the Fisher (1953) distribution on the sphere. (b) Quite strong concentrations ($\kappa$) have broad dispersions and some directions antiparallel to the alignment are always present. (c) Comparison of orientation-distributions is facilitated by the cumulative summation plot (CUSUM) (Fisher, 1993), here shown for differently concentrated Von Mises’ distributions. (d) CUSUM declinations show asymmetry with a broad tail to the West, with vector means close to North. (e) CUSUM inclinations of the intermediate component are close to the GAD inclination. However, the hard-component is shallowed and skewed by compaction; if its initial inclination was similar to the intermediate component, compaction was 51%.
10. Conclusions

The de-compaction correction, based on an average vertical compaction of 51\% restores the two secular-variation loops to an almost geocentric orientation (Fig. 9f). The paleopoles for Fig. 9 were based on moving averages of 10 specimens; the 95\% confidence regions assume a Fisher distribution for each of the 10 specimens in each case. However, the paleopoles for the uncorrected hard-component are still somewhat far-sided (Fig. 9e), indicating that the inclinations-shallowing is still not completely corrected by this de-compaction approach. This unique marker horizon of laminated red clay was deposited in a westward-draining channel linking the former lakes Kamanistikwia and Agassiz. Anisotropy of magnetic susceptibility (AMS) verifies the flow axis was constant through 1.6 m of sediment, which corresponds to one to two millennia according to the number of PSV loops (Fig. 9f).

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