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PALEOMAGNETISM OF THE WRANGELLIA AND ALEXANDER TERRANES

AND

THE TECTONIC HISTORY OF SOUTHERN ALASKA

A

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PALEOMAGNETISM OF THE WRANGELLIA AND ALEXANDER TERRANES

AND

THE TECTONIC HISTORY OF SOUTHERN ALASKA

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Abstract

Wrangellia was the first Alaskan tectonostratigraphic terrane to be widely accepted as allochthonous with respect to North America. There is, however, considerable disagreement as to the age of emplacement of this terrane as well as the hemisphere in which it originated. Some 800 paleomagnetic samples were collected from 24 localities in southern Alaska to elucidate the paleolatitude translation history of Wrangellia and other associated terranes.

Data of known polarity from the Skolai Group (Pennsylvanian/Permian) strongly suggest that Wrangellia originated at 10-15 degrees North latitude. The Permian Pybus Dolomite yields a 9 degree S paleolatitude and suggests that the Alexander terrane moved southward in late Paleozoic and Triassic time. Evaluation of geologic data indicates that the Wrangellia and Alexander terranes amalgamated in an oceanic setting in mid to Late Jurassic time. Paleomagnetism of the Brothers Volcanics (Alexander terrane) and MacColl Ridge Formation (Wrangellia) documents a low latitude for both terranes during the Cretaceous, thereby precluding a pre-Tertiary age of emplacement for the amalgamated superterrane. Speculative apparent polar wander paths for Wrangellia and the Alexander terranes, in addition to geologic and biogeographic constraints, allow development of the following hypothetical tectonic model:

Both the Alexander and Wrangellia terranes originated in the northern hemisphere adjacent to western North America in mid-Paleozoic and late Paleozoic times, respectively. The Alexander terrane

moved into the southern hemisphere during the Paleozoic and Wrangellia began moving southward in the Late Triassic or Early Jurassic. These two terranes amalgamated in mid to low southern paleolatitudes in later Jurassic time and formed part of a composite terrane, here termed the Southern Alaska superterrane. This superterrane began northward translation in Late Jurassic time, accreting to North America in Tertiary time.

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Introduction

Although "continental drift" and plate tectonics have been generally accepted for some time, the concept that large portions of western North America were not always part of the continent has only recently gained support in the geologic community. The allochthonous nature of a segment of western North America was first suggested by Monger and Ross (1971). This inference was based on paleontological evidence for two belts of distinctive North American fusilinid occurrences which are separated by a belt of Tethyan fusilinids. Their favored explanation for this biogeographic distribution was strike-slip displacement of a once continuous North American belt against the seaward side of the Tethyan belt (Monger and Ross, 1971). This was the first of several discoveries which set the stage for the later development of the tectonostatigraphic terrane theory.

The first attempt to quantify displacement of allochthonous areas of Alaska was advanced by Packer and Stone (1972). They determined a virtual geomagnetic pole (VGP) for Jurassic rocks occurring on the Alaska Peninsula and in the Nutzotin Mountains. This pole is discordant with respect to the North American Jurassic pole. To bring these poles into coincidence requires approximately 18 degrees of northward motion and 52 degrees counterclockwise rotation of the sampled localities with respect to the craton.

Jones and others (1972) extended the work of Monger and Ross (1971) to Alaska and the western United States. Based on faunal data and Silurian sedimentary facies, Jones and others (1972) ruled out a simple rift model to explain the three fusilinid belts in favor of a transcurrent faulting model. Jones and others (1972) went on to speculate that the western fusilinid belt, now in southeast Alaska, originated near present day California.

The first part of Alaska to gain general acceptance as a far-traveled block is Wrangellia. Jones and others (1977) described the Wrangellia terrane as a distinctive Triassic stratigraphic sequence of thick tholeiitic basalt flows overlain by platform carbonate rocks and coined the name "Wrangellia". This sequence is identified in the Wrangell Mountains of southcentral Alaska (the "type" area), southeast Alaska, the Queen Charlotte Islands and Vancouver Island in British Columbia and in eastern Oregon (Jones and others, 1977). In a companion study, Hillhouse (1977a) presented paleomagnetic data from the Triassic Nikolai Greenstone in the Wrangell Mountains which document a low equatorial paleolatitude of about 15 degrees.

Jones and others (1983) define tectonostratigraphic terranes as fault-bounded geologic entities having histories which are significantly different from the histories of neighboring, often coeval terranes. Within a terrane, a uniform stratigraphy can be traced over distances of a few kilometers to many hundreds of kilometers. The sedimentary facies changes observed within a terrane follow a geologically reasonable progression of rock types, which can be related to a plausible paleo-environmental facies mosaic. Adjacent terranes show a markedly different geological history and/or sedimentary facies. The changes in sedimentary facies are abrupt and cannot be explained as simple shifts in paleo-environments. For example, the southern Wrangell Mountains contain Late Cretaceous shallow marine and possibly non-marine sedimentary rocks (Jones and MacKevett,1969), whereas the rocks just a few miles to the south probably represent a deep marine trench environment (Decker, 1980). The boundaries between such terranes are either known or suspected major faults.

Tectonostratigraphic terranes are useful units for paleogeographic analysis (Beck and others, 1980) because they comprise rock units forming laterally and stratigraphically continuous entities, which may have been transported relative to one another. It must be pointed out, however, that subdivision into terranes does not necessarily imply large scale tectonic transport of one unit with respect to the craton or other units. For example, Coney and others (1980) employ the term "suspect terrane", which is used as a descriptive term and leaves open the question of relative displacement. Subsequent to the initial discoveries of allochthonous terranes in Alaska, the entire state and indeed much of western North America has been subdivided into tectonostratigraphic terranes.

Although the allochthonous nature of Wrangellia is generally accepted (Irving, 1979, Coney and others, 1980, McWilliams, 1983), its translational history is still widely debated. One of the more significant problems in deciphering the translational history is constraining the hemisphere which Wrangellia occupied at a given time. This is difficult to do because it is not usually possible to determine from a few isolated studies whether an observed magnetization was acquired during a normal or reversed geomagnetic field. The polarity of the geomagnetic field at the time of magnetization controls the sign of the inclination of magnetization, which is in turn used to determine the hemisphere (northern or southern) in which the magnetization was acquired. Because of this polarity uncertainty, the hemisphere in which Wrangellia resided during the Triassic is much debated (Panuska and Stone, 1981, Stone and others, 1982, Hillhouse and Gromme, 1984). Since there is no a priori reason for preferring the northern or southern hemisphere Triassic paleomagnetic pole as the equivalent of the north geographic pole, the polarity is ambiguous. Thus, it cannot be determined from the Triassic paleomagnetic data whether Wrangellia occupied a position 15 degrees north or south of the equator. In addition, the time of accretion of Wrangellia to North America is not well known.

Estimates of the accretion age of the southern Alaska portion of Wrangellia range over a timespan of 100 m.y. Jones and others (1982) suggest a pre-Albian, post-Barremian (Early Cretaceous) accretion, based on the probable age of deformation in the Wrangell Mountains. Another estimate of accretion timing based on deformation is post-Cenomanian (Late Cretaceous) and pre-Paleocene (Jones and others, 1982b) in the central Alaska Range. Stone and others (1982) suggested a late Tertiary accretion for Wrangellia based on paleomagnetic evidence from outboard and adjacent terranes. Moore and others (1983) proposed a Paleocene accretion of Wrangellia and associated terranes based on geologic and paleomagnetic evidence from outboard terranes; however, they recognize the possibility that the outboard terranes may have moved independently of Wrangellia. Such drastically different hypotheses for the emplacement age of the southern Alaska terranes require testing before satisfactory paleogeographic models can be developed.

This study was undertaken to define the movement history of Wrangellia and related terranes. Although terrane motions may be detected by recognizing offsets in geological formations and by the reconstruction of paleobiogeographic provinces, paleomagnetism provides the most quantitative and most direct evidence for large scale motions. Paleomagnetic results can provide constraints for tectonic and paleogeographic models for rocks now found in southern Alaska. To do this, it is necessary to know the magnetic polarity of the field recorded by the rocks used; otherwise, each paleolatitude determination would be consistent with either a northern or southern hemisphere interpretation. The lack of constraints on the hemisphere of origin results in too many variables, which leads to an unmanageably large number of possible models. The polarity ambiguity may be resolved for these localities and extrapolated to other localities which have equivocal polarities, by sampling rocks formed during recognizable geomagnetic polarity intervals. If the polarity of a sequence can be established, it should then be possible to construct apparent polar wander (APW) paths and to test latitude versus time trajectories for individual areas. The early Mesozoic southward motion followed by late Mesozoic/Cenozoic northward motion suggested for southern Alaska by Stone and others (1982) is testable by constructing complete APW paths. Similarly, sampling in the Alexander terrane of southeast Alaska, which is geologically linked to Wrangellia (Berg and others,1972), will allow a more complete paleomagnetic history of the whole of Southern Alaska to emerge.

Since paleomagnetic data alone can only determine relative paleolongitude under a very limited set of conditions, involving well-defined polar wander paths and limited relative motion, geological and biostratigraphic data from the literature are used to establish relative paleolongitude control. These data are supplemented by a reconnaissance paleocurrent study to help determine critical ages of terrane linkages.

Regional geology, viewed from the tectonostratigraphic terrane concept, is reviewed in the next chapter in order to establish a framework in which to interpret the paleomagnetic studies (the details of local geology are discussed by formation in the data sections). This is followed by a discussion of the analytical methods and conventions. Subsequent chapters document the paleomagnetic data and relevant geology with little or no interpretation. After the presentation of the paleomagnetic data, the geologic evidence for terrane histories is described and interpreted and incorporated into a paleogeographic model.

Regional Geology

Nature and Extent of Wrangellia

The Wrangellia terrane was defined by Jones and others (1977) on the basis of a unique Triassic stratigraphy, basement rocks, fauna and geologic history. In the Wrangell Mountains of Alaska, the distinctive Triassic sequence was deposited upon upper Paleozoic andesitic rocks (Skolai Group) and consists of thin and areally restricted Daonella-bearing chert, siltstone and shale (Ladinian, Middle Triassic) overlain by subaerial tholeiitic basalt over 3000 m thick (see figure 1 for a generalized stratigraphic column for each of the southern Alaska terranes). This is in turn overlain by the Chitistone and Nizina Limestones, platform carbonate rocks of Carnian and Norian age (Late Triassic). The carbonate rocks contain diversified faunas, including the bivalve genera Monotis and Halobia, and reflect a gradual deepening of the basin from supratidal, sabkha and intertidal carbonates to shallow marine limestone and finally to deep-water deposition of shale and chert (Armstrong and others, 1970). A stratigraphic study by MacKevett (1971) shows that quiet water, starved-basin-type conditions, represented by the McCarthy and Lubbe Creek Formations, persisted until the onset of clastic deposition in Middle Jurassic time (see figure 1).

Stratigraphies similar to the Wrangell Mountains section have been identified in southeast Alaska, Queen Charlotte Islands, B.C., Vancouver Island, B.C., and in eastern Oregon and western Idaho (Jones and others, 1977). The same sequence of thick Middle to Upper

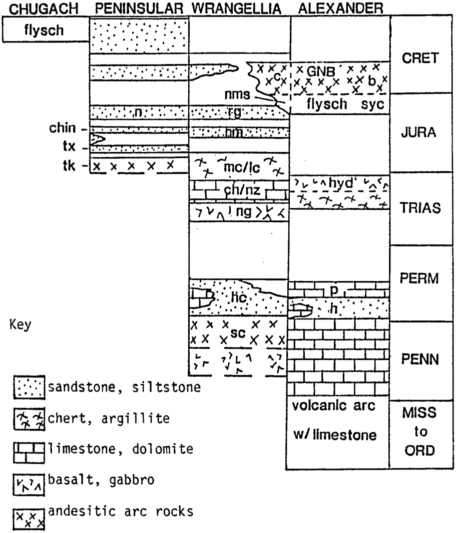

Figure 1 Generalized stratigraphy of the major southern Alaska terranes. Formation/Group names are as follows: Peninsular: n-Naknek, chin-Chinitna, tx-Tuxedni Group, tk-Talkeetna. Wrangellia: rg-Root Glacier, nm-Nizina Mountain, mc/lc-McCarthy and Lubbe Creek, ch/cz-Chitistone and Nizina, ng-Nikolai Greenstone, hc-Hasen Creek, sc-Station Creek. Gravina-Nutzotin Belt (GNB): c-Chisana, nms-Nutzotin Mountain Sequence, b-Brothers, syc-Seymour Canal. Alexander: hyd-Hyd Group, p-Pybus, h-Halleck. Wrangellia and Peninsular terrane Cretaceus formations are too numerous to show. Gravina-Nutzotin Belt is a superjacent terrane in depositional contact with both Wrangellia and the Alexander terrane. Modified from Jones and Silberling (1979) with data from Churkin and Eberlein (1977), Loney (1964) and Berg and others (1978).

Triassic basaltic rocks overlain by Upper Triassic platform carbonates (with similar faunas) deposited in a gradually deepening marine environment were observed in each area. The assignment of the Hells Canyon section of eastern Oregon, western Idaho (Seven Devils terrane) to the Wrangellia terrane is the most problematic correlation.

Newton (1983) compared Norian (Late Triassic) bivalve faunas from the Wrangell Mountains and the Hells Canyon section (Seven Devils terrane) and found very similar species compositions, which tend to support the terrane correlation. In addition, Hillhouse and others (1982) demonstrated a paleomagnetic paleolatitude similar to the latitudes observed in the Vancouver and Wrangell Mountains portions of Wrangellia. However, Sarewitz (1983) argues that the Seven Devils terrane and Wrangellia formed in different tectonic settings and therefore should not be correlated. Sarewitz (1983) studied the geology and geochemistry of volcanic rocks from the Seven Devils terrane and concluded that these rocks represent a convergent margin volcanic arc of probable calc-alkaline affinity. In contrast, the Triassic volcanic rocks from elsewhere in Wrangellia (Nikolai Greenstone, Wrangell Mountains; Goondip Greenstone, southeast Alaska; and the Karmutsen Basalt, Queen Charlotte Islands and Vancouver Island) appear to be tholeiites unrelated to a convergent margin setting (Sarewitz, 1983). Even if further work can substantiate different tectonic environments for the Wrangellia and Seven Devils terranes, a relatively close paleogeographic association is not precluded. For example, an island arc and its associated back arc basin have different petrologic affinities yet share the same general geographic setting. Paleogeographic reconstructions are further complicated by the fact that the original Wrangellia (whatever its boundaries may be) may have been dismembered in post-Triassic time and the individual fragments may have had significantly different accretion ages. Thus, regardless of the original extent of Wrangellia, a uniform chronology of events leading to the accretion of the various parts to North America is not required. Indeed, considering the present geographic distribution of Wrangellia rocks over perhaps 2000 km, a uniform accretion history for the entire length of Wrangellia is not to be expected. Since the accretion history of one segment of Wrangellia cannot be used as a tectonic constraint for the entire terrane, the scope of this study is restricted to the paleogeography and tectonics of the Wrangell Mountains portion of Wrangellia and presently juxtaposed terranes (figure 2).

Neighboring Terranes

Jones and others (1977) excluded rocks from Admiralty, Kuiu and Kupreanof Islands, southeast Alaska (part of the Alexander terrane) in their original delineation of the Wrangellia terrane. Although this area does contain Upper Triassic basalt (part of the Hyd Group), these volcanic units are slightly younger than the Nikolai Greenstone (Wrangellia) and are generally only a few hundred meters thick. Moreover, the Triassic rocks of the Alexander terrane do not rest on rocks that can be equated with the Pennsylvanian-Permian andesitic volcanic basement of the Wrangellian Triassic (figure 1). The age correlative rocks of the Alexander terrane are largely carbonates and clastic

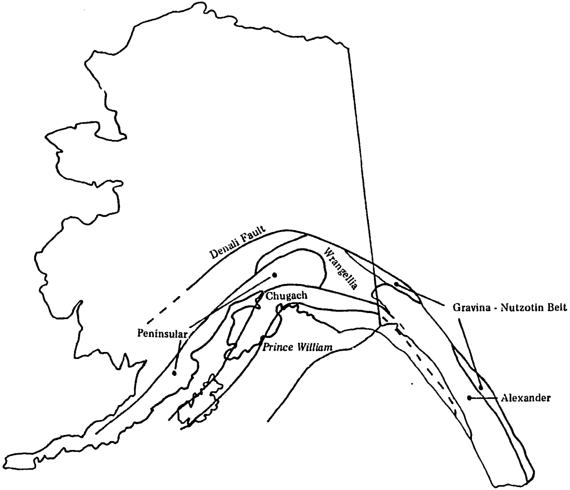

Figure 2 Map showing generalized locations of the major tectonostratigraphic terranes occurring south of the Denali fault. Terrane boundaries are from Jones and others, (1981). Terrane boundaries are dashed where approximated.

rocks. Additionally, the base of the Alexander terrane is composed of volcanic, volcaniclastic, carbonates and granitic rocks as old as Ordovician (Berg and others, 1972, Jones and others, 1977, Churkin and Eberlein, 1977), considerably older than the oldest Wrangellia rocks.

Based on the markedly different Paleozoic and Triassic stratigraphies, Jones and others (1977) suggested that the Alexander and Wrangellia terranes joined in post-Triassic time. These two terranes are in fault contact (MacKevett and Jones, 1975) and had to be linked by Late Jurassic because the superjacent Gravina-Nutzotin flysch-volcanic belt of Late Jurassic age depositionally overlies both terranes (Berg and others, 1972) (figure 1). Coney and others (1980) interpret this linkage as an amalgamation prior to accretion to the North American continent, since marine conditions inboard of this superterrane persisted at least until mid-Cretaceous time.

Although the boundary between the Wrangellia and Alexander terranes and their stratigraphic distinctions are fairly well known, the relationship between Wrangellia and the Peninsular terrane to the west is more equivocal. Jones and Silberling (1979) suggested a separation of these terranes which ended by Late Jurassic time with the deposition of the Root Glacier Formation (Wrangellia) and the Naknek Formation (Peninsular terrane), which they consider correlative units (figure 1). The Peninsular terrane was originally distinguished from Wrangellia on the basis that the Triassic volcanism in the Peninsular terrane seems to be of Norian age rather than Carnian (the age of the Nikolai Greenstone) and the apparent lack of any Triassic rocks older than late

Carnian (Jones and others, 1977). It must be pointed out, however, that the negative evidence for terrane correlation could be due to the exceedingly rare occurence of Triassic exposures anywhere in the Peninsular terrane (Jones and others, 1977).

Early Jurassic geographic separation of the terranes is suggested by marked differences in the Lower Jurassic stratigraphy (Jones and Silberling, 1979) (figure 1). The Lower Jurassic section of the Peninsular terrane is dominated by the widespread Talkeetna Formation, a thick unit of andesitic volcanic and volcaniclastic rocks. The time equivalent rocks in Wrangellia are entirely fine-grained, deep-water rocks (McCarthy and Lubbe Creek Formations). The lack of coarse-grained detritus and the lack of a volcanogenic signature in the Lower Jurassic Wrangell Mountains rocks prevent the establishment of a common Early Jurassic history for Wrangellia and the Peninsular terrane (Jones and Silberling, 1979). However, since the closest exposures of Lower Jurassic deposits on the two terranes are separated by some 90 km of late Cenozoic fill (Copper River Basin), geological ties between Wrangellia and the Peninsular terrane cannot be precluded. The pre-Late Jurassic paleogeographic relationship of Wrangellia and the Peninsular terrane remains enigmatic, due to the paucity of Peninsular terrane Triassic rocks, uncertain relationships and lack of any obvious orogenic event that could be related to terrane amalgamation.

The Peninsular and Wrangellia terranes appear to share a common history from Late Jurassic time through the remainder of the Mesozoic (Jones and Silberling, 1977). Major orogeny, uplift and erosion occurred near the close of the Jurassic. In Early Cretaceous time, shallow marine and non-marine conditions were re-established and remained throughout much of the Cretaceous. These rocks form a continuous belt of Cretaceous deposits along much of the length of the combined Wrangellia and Peninsular terranes.

Outboard of these terranes, there are two major terranes made up of deep-water flysch-like sediments and volcanic rocks of Late Cretaceous and early Tertiary age (Plafker and others, 1977) (figure 2). These two terranes, the Chugach and the Prince William, form part of a logical paleogeographic association of terranes. The Peninsular, Wrangellia and Alexander terranes could form the backbone of a mid-to-late Mesozoic island arc and the Gravina-Nutzotin, Chugach and Prince William Terranes could form the deep-water basins expected to flank the island arc. As yet, there are too many gaps in the stratigraphic record and too many poorly known geological relationships (especially paleogeographic links) for this simplistic association of terranes to be very satisfying. It is, however, important to note these relationships because several tectonic and paleogeographic models rely heavily on the association of outboard deep-water terranes and the inboard arc.

The problem of geologic evidence for terrane amalgamations will be discussed in a later chapter. Specific details of the geology are deferred to the paleomagnetic data sections, where local geologic relationships are of critical importance. However, before the data are described, a discussion of paleomagnetic techniques and sampling plan is in order.

Paleomagnetic Data Acquisition

The object of paleomagnetic research applied toward tectonic problems is to obtain measurements of the ancient geomagnetic field recorded in rocks and use these data to interpret past movements of the rocks. Several readings of the instantaneous geomagnetic field averaged over periods greater than 100,000 years yields an axial geocentric dipolar field which is roughly parallel to the earth's geographic axis (McElhinny, 1973). Thus, with a knowledge of the time-averaged direction of magnetization of a given rock unit, it is possible to calculate a virtual geomagnetic pole or VGP (that is, the geographic position of one pole of a dipole field that could have produced the observed magnetization) and equivalent paleolatitude of the site. In this way, it is possible to reconstruct the plate tectonic motion of a stratigraphic succession of rocks.

Sampling Plan

In most cases, paleomagnetic samples were collected as cores drilled from an outcrop using a hand-held, gasoline-powered drill with a diamond tip drill core barrel. Samples were oriented using magnetic readings, visual sightings to distinct geographic landmarks and sun compass measurements. In the rare instances where cores could not be drilled, oriented hand samples were collected. Cores were then drilled from the hand samples in the laboratory. Hand sample collection was avoided, where possible, as the intermediate collecting step introduces an additional source of error. Unless otherwise stated, all samples were drilled directly from the outcrop.

Various conventions exist to describe the paleomagnetic sample hierarchy. McElhinny (1973) employs the terms "collecting site" or "locality" to indicate the geographic location of an outcrop, "site" to indicate a unit representing an instant in geologic time (e.g. a sedimentary bed or lava flow), "sample" to indicate an oriented core or hand sample and "specimen" to indicate a replicate of the sample which is subjected to various magnetic cleaning and measurement procedures. For this dissertation, the following sample nomenclature is used. "Locality" designates an outcrop area which exposes a nearly continuous section of undisrupted rock. Localities are on the order of 50 to 200 meters in longest dimension. Due to the relatively small size of a locality, a more or less uniform post-depositional history may be inferred for all samples derived from it. The uniform post-depositional history would entail factors such as depth of burial and diagenetic and thermal conditions but necssarily excludes small scale effects such as thermal overprinting due to dike injection or localized chemical alteration along veins. The term "bed" or "flow" refers to a unit of rock that can be thought of as having been deposited very quickly. Examples include massive turbidite sandstones, pyroclastic units and lava flows. Any remanent magnetization observed within this unit would then give an instantaneous record of the earth's magnetic field. Any lithologies not conforming to such a rapid deposition model must be evaluated on a case by case basis. The term "sample" is simply an abbreviation for "core sample". In this study all core samples are standard one inch (2.54 cm) diameter cores. A "specimen" is an individual piece of rock, 2 cm in axial length, cut from a core.

To recapitulate, a locality comprises several flows or beds. A bed may have several core samples drilled from it; a core sample may have one or more specimens cut from it.

An oriented hand sample is a special case of the preceding scheme. The hand sample is a portion of a bed or flow. Several cores are drilled from the hand sample and the cores are sliced into specimens.

In the event that a pluton is sampled, the bed or flow sample level is meaningless. Since plutons generally require a relatively long period of time to cool, samples of a pluton automatically contain a time-averaged thermoremanent magnetization (TRM) and the bed/flow sample level of the sampling hierarchy is thus eliminated.

The sampling plan at any given locality took one of two general forms. Reconnaissance sampling consisted of drilling one core per bed over as large a stratigraphic distance as possible. Sampling over a large stratigraphic thickness is an attempt to sample sufficient time to average out secular variation of the magnetic field. However, the lack of replicate samples from each bed precludes the possibility of assessing within-bed variation, except where a sufficient number of specimens from individual cores could be obtained. The alternative high density sampling technique entailed drilling multiple core samples in each bed, again over as large a stratigraphic interval as possible. The number of duplicate samples in each bed varied between two and eight. The most common sampling density was three or four samples per bed. Occasionally six samples were taken as an experiment to determine whether additional samples could substantially reduce the error limits. Eight samples were sometimes taken to perform particular experiments such as the baked contact test (McElhinney, 1973). Two samples per bed at a "high density" locality usually reflects poor drilling conditions or the lack of a nearby water supply (for drilling). The reconnaissance sampling scheme was employed when the amount of time that could be spent at a locality was limited by logistic considerations. In addition to the details of the outcrop scale sampling plan, the regional sampling scale also deserves discussion.

The goal of this paleomagnetic research is to develop a latitude versus time curve and an apparent polar wander (APW) path for Wrangellia and associated terranes. The correct polarity of a latitude determination can be inferred from such a path. Ideally, this objective would best be satisfied by sampling a continuous stratigraphic sequence from the base of the terrane to the most recent beds in one structural block from one geographic location. In addition, this location should have suitable structures for the application of fold tests. Such a scheme would yield an APW path from the present north pole back through time, giving latitudes of known polarity, and would obviate the need to correlate poles (perhaps incorrectly) from separate structural blocks that could have undergone differential rotation. Unfortunately, such conditions are infrequently met in nature. Specifically, in the Wrangell Mountains, the geographic distribution of the rock systems and the absence of some critical systems and series (especially the early and mid Tertiary) prevent the acquisition of data from a continuous section. Moreover, on the pragmatic level, a geographically restriced paleomagnetic sampling pattern is not desirable, because remagnetization of this one locality would render the entire data set worthless.

A regional sampling plan was used in an effort to limit the number of data sets lost to remagnetization problems and to acquire data from the appropriate time-stratigraphic units. The first requirement of this plan is to sample units representing the necessary time intervals. Pennsylvanian/Permian, Jurassic and Cretaceous units are deemed high priority sampling targets. Since the Triassic paleolatitude of Wrangellia is relatively well known (Hillhouse, 1977), these rock systems show the most promise for providing paleomagnetically useful information, especially the Permian which can resolve the polarity ambiguity. (See figure 3 for a time-stratigraphic representation of sampling distribution.)

The second most important sampling criterion is the occurrence of field relationships that lend themselves to testing the stability of magnetization. This requires that sample localities be selected which have been relatively well-mapped and have good geochronological control. (Existing geological maps and geochronological control is discussed on a formation by formation basis in the data sections.) The two most powerful stablility tests are the reversal test and the fold test. The reversal test consists of finding statistically antiparallel reversals of magnetization direction, recording reversals of the geomagnetic field, within a stratigraphic sequence. The fold test entails determining a statistical alignment of magnetization directions measured from different limbs of a fold after applying a tectonic tilt correcton. A positive fold test proves that magnetization was acquired before tectonic deformation. Where well-mapped folds are absent in

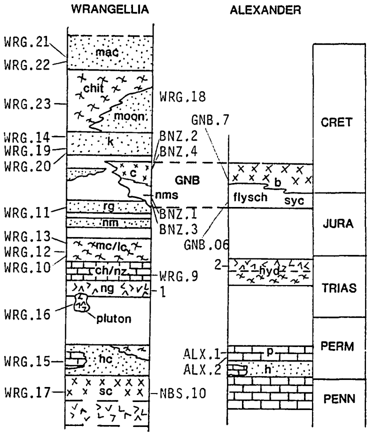

Figure 3 Generalized stratigraphic columns of Wrangellia and Alexander terranes depicting stratigraphic distribution of sampling localities. Locality 1 is from Hillhouse (1977) and locality 2 is from Hillouse and Gromme (1980). Symbols are the same as figure 1. Formation names: mac-MacColl Ridge, chit-Chititu, moon-Moonshine Creek, k-Kennicott, c-Chisana, nms-Nutzotin Mountains Sequence, rg-Root Glacier, nm-Nizana Mountain, mc/lc-McCarthy and Lubbe Creek, ng-Nikolai Greenstone, hc-Hasen Creek, sc-Station Creek, b-Brothers, syc-Seymour Canal, hyd-Hyd Group (includes Hound Island Volcanics), p-Pybus and h-Halleck. Sample localities are listed in Table of Contents by formation names with the exception of BNZ localities (Bonanza Creek) and WRG.15, 16, 17(Skolai Group). GNB is Gravina-Nutzotin Belt.

rocks of appropriate age, a regional fold test may be applied by comparing mean outcrop magnetization directions before and after tilt correction. The regional fold test is, however, an inherently less powerful test than a mapped fold-fold test, due to more equivocal structural relationships.

With these sampling criteria and logistic considerations as a guide, a regional sampling program was undertaken. The major sampling effort was concentrated in late Paleozoic, Jurassic and Cretaceous rocks in the southcentral Wrangell Mountains, within the Wrangellia terrane (the major theme of this study). In addition to the main sampling area, an outcrop of volcanic rocks of Pennsylvanian age was collected near Nabesna in the northen Wrangell Mountains. A sequence of Upper Jurassic and Lower Cretaceous flysch and volcanic rocks was sampled in the Nutzotin Mountains near Chisana. This locality was selected because it contains a thick, well-dated section (necessary for a reversal test) of the only non-Triassic, Mesozoic volcanic rocks in the Wrangell-Nutzotin region. The section is of Early Cretaceous age, an age which is not represented elsewhere in the southcentral Wrangell Mountains study area. The final sampling area of this project is located within the Alexander terrane bordering Frederick Sound, southeast Alaska. Samples were collected in this region to look for paleolatitude similarities and differences between the Alexander terrane and Wrangellia. The general locality was chosen on the basis of logistic feasibility, occurrence of rocks of suitable ages and the relatively mild degree of metamorphism.

In any paleomagnetic investigation, the question of age of magnetization is of critical importance. Sampling sites were thus restricted to areas where previous workers have mapped and determined the ages of the units. The age of a rock unit may be determined by radiometric means, by fossil evidence and with the aid of magnetostratigraphic data. Discrepancies often arise between the radiometric and fossil ages of a particular unit due to inadequate calibration of the biostratigraphic time scale with absolute radiometric ages. Numerous geologic time scales exist because of the continuing effort to determine radiometric ages of biostratigraphic boundaries. For the purpose of this study the Decade of North American Geology geologic time scale of Palmer (1983) is used to correlate radiometric and paleontological ages.

Measurement

Measurements of the directions of remanent magnetizations were made at the Woodward-Clyde and Associates paleomagnetic laboratory in Oakland, California (now Z-Axis Exploration of Walnut Creek, California). The laboratory is housed in a magnetically shielded room, with internal fields of less than 500 gammas. Instrumentation consists of a three-axis, 6.4 cm diameter access, Superconducting Technologies magnetometer with a computer interactive system and instantaneous graphic data display screen. The principles and capabilities of this type of magnetometer are discussed by Goree and Fuller (1976). The magnetometer has an optimum sensitivity of approximately 2 x 10^{-8} emu/cc with less than 2 percent error.

Specimens are lowered into the measurement chamber in a plastic holder. The holder characteristically has an NRM (natural remanent magnetization) of about 2 x 10^{-8} emu/cc. However, the holder's NRM may be subtracted from the measurements to give a background noise of approximately 3 x 10^{-9} emu/cc in random directions. This intensity is 2 to 3 orders of magnitude weaker than the average specimen in this study.

All specimens were subjected to alternating field (AF) or thermal cleaning techniques or a combination of the two, in order to remove any secondary components of magnetization (see McElhinny, 1973). Blanket cleaning treatment of the specimens was not done, since the magnetic properties of specimens vary and blanket treatment may overshoot or undershoot the optimum cleaning step. Rather, all specimens were cleaned in a stepwise fashion, with demagnetization steps being determined on an individual basis by a specimen's preceding response to the cleaning.

Thermal cleaning was performed in a Schonstedt thermal demagnetizer. Specimens were heated for a minimum of 15 minutes at low temperatures (less than 300 degrees C) and 25 minutes at high temperatures (above 450 degrees C). The specimens were then cooled in a shielded low field environment for lengths of time equivalent to the heating time. Specimens that broke up and could not be unequivocally reassembled to their original configurations before a characteristic direction could be obtained were eliminated from further analysis.

AF demagnetizations were carried out in an S.C.T. shielded, 400 Hertz, three-axis tumbling demagnetizer. The decay time of the field was increased with increasing peak field intensity. Peak fields of 50 oersteds and less were decreased to zero over no less than 30 seconds. In general, the cut-off maximum peak field for various decay times was 200 oersteds at 45 seconds, 350 oersteds at one minute, 600 oersteds at 1.5 minutes. Peak fields exceeding 700 oersteds were decreased to zero over 2.25 minutes.

Demagnetization data were evaluated using standard graphical plots: percent normalized demagnetization curves, Zijderveld vector diagrams and stereographic projections of vector directions. In this study, a modified Zijderveld diagram was used. The standard Zijderveld diagrams are double orthogonal vector projections into a horizontal plane and a vertical plane coincident with either a north-south or east-west direction. This plotting convention has the disadvantage that a specimen with two greatly differing components, or a poor choice of orientation of the vertical plane, shows only an apparent dip of the vector inclination. This type of vector diagram was modified so that the inclined component was plotted in the plane containing it, thus plotting the angle of inclination without regard to the direction of the horizontal component. This has the affect of "rotating" the vertical plane for each point so that the vector is always contained in the vertical plane, thus accurately representing inclination angle and intensity.

Stereographic vector plots were made in both the geographic (uncorrected for bedding tilt) and stratigraphic (corrected for bedding tilt) frames of reference. The AF cleaned data were plotted as geographic directions because this frame of reference facilitates the identification of rotational remanent magnetization (RRM).

RRM was first positively identified by Wilson and Lomax (1972). They found that specimens demagnetized while being rotated in an alternating field could acquire a magnetization antiparallel to the innermost tumbling axis of the specimen. This effect was observed at even very low rates of rotation and was found to be independent of any pre-existing remanent magnetization. This phenomenon was further studied by Stephenson (1980a, 1980b), who ascribed this behavior to gyroscopic properties of electron spin and proposed the term gyroremanent magnetization (GRM). Regardless of its origin, this spurious, acquired magnetization can be related to the inner rotation axis of a specimen tumbled during AF cleaning. Thus, the descriptive term RRM will be used.

In order to routinely check for unwanted RRM, specimens were loaded into the AF demagnetizer in a consistent orientation such that any RRM acquired would be acquired along the +Y coordinate of the specimen. If RRM was suspected at a given demagnetization level, the specimen was loaded in the inverted position and demagnetized again at the same intensity. If RRM was acquired, it would then be observed along the -Y coordinate of the specimen, in the opposite sense of the previous demagnetization.

To aid in the identification of RRM, geographic frame of reference stereographic plots were used. In these plots, the specimen's Y-coordinate corresponds to the direction 90 degrees to the azimuth of the core (the bearing of the axis of the core sample as it was drilled in outcrop). Thus, in these plots, RRM would show up as a deflection towards the +Y direction, followed by deflection towards the -Y direction.

Where only a mild RRM component was observed, the final, stable vector direction was taken as the mean of the measured vectors with positive and negative RRM (Hillhouse, 1977b). Some specimens showed a significant RRM before a stable direction could be isolated. In these cases, where RRM was much stronger than the stable vector direction, the specimens were disregarded in further analysis.

Magnetic Recording Properties of Rocks

Various workers have measured magnetization directions from volcanic, plutonic, metamorphic and sedimentary rocks in order to estimate the direction of the ambient geomagnetic field at some location during the geologic past. The plane of ancient horizontal for the locality must be known for magnetic vector directions to be useful in paleogeographic investigations. In most cases, metamorphic rocks are of no value because the ancient horizontal datum has been lost. Similarly, plutonic rocks have limited applicability as paleogeographic indicators. In some instances, investigators have made paleolatitude interpretations based on data from plutonic rocks by inferring original horizontal from structural arguments and field relationships. Such inferences are rarely entirely satisfactory.

Volcanic rocks are widely used in tectonic studies due to their good rock magnetic recording properties. As a lava flow cools, it acquires a thermoremanent magnetization (TRM) which accurately records the ambient field direction. However, erroneous tectonic tilt corrections can be applied to the data if flow contacts, which may have high initial dips, are taken as faithful indicators of ancient horizontal. In this respect, sedimentary rocks provide the best representation of ancient horizontal. Unfortunately, some sedimentary rocks contain a systematic inclination error. Sedimentary rocks usually record the declination of the magnetic field accurately but may record an inaccurate inclination. Whenever an inclination error is observed, it is always systematically lower than the true inclination.

Sedimentary rocks can record the direction of the geomagnetic field by either chemically producing magnetic minerals after deposition (chemical remanent magnetization, CRM) or by the mechanical alignment of magnetized detrital grains along the magnetic field (detrital remanent magnetization, DRM). A depositional DRM inclination error can be produced by several processes such as rolling of grains into depressions immediately after deposition or flattening of elongate magnetic grains (preferentially magnetized parallel to the long dimension) as they settle on a flat surface (see Verosub, 1977, for a review). Such inclination errors have been observed in natural sediments and sediments deposited under laboratory conditions. Tarling (1983) considers the inclination error to be related to high rates of deposition and low initial porosity, which would inhibit the post-depositional realignment of magnetic grains. Post-depositional realignment of the magnetic grains is considered to occur as a result of thermal agitation. This can only occur if the magnetic grains are small compared to the size of the interstitial spaces and if the water content is high enough to permit grain mobility. Experimental evidence supporting post-depositional realignment is reported by Payne and Verosub (1982). Payne and Verosub (1982) found that sediments with high silt or clay contents resisted magnetic realignment but sediment with greater than 60 percent sand content realigned with a new magnetic field direction, provided that the sediment contained sufficient water (40-60 percent). Thus, it is likely that sandstones generally possess a post-depositional DRM largely free of inclination errors. Provided that the sandstone dewatered within a geologically short period of time following deposition, the magnetization can be considered to date back to the age of the rock. Although more experimental work is needed, the present evidence suggests that at least sandstones are reliable recorders of the ancient geomagnetic field. Additionally the inclination error in some fine-grained sediments has been found to be relatively small (Tarling, 1983).

Carbonates form a special class of sediments where the remanent magnetization is thought to be carried by single domain and pseudo-single domain biogenic magnetite grains. Once the biogenic magnetite grains are incorporated in the carbonate sediment, post-depositional realignment processes similar to those acting on clastic sediments should operate. However, little experimental work has been done on carbonate rocks.

Analysis of Data

Characteristic components of magnetization were determined for each specimen using a method based largely on the method of Zijderveld

(1967). This technique relies on the principle that magnetic vectors measured from a given specimen subjected to stepwise cleaning procedures will display a minimal change in direction and a systematic decrease in intensity over successive demagnetization intervals when a stable component of magnetization has been isolated. In general, the component with the highest coercivity or highest unblocking temperature is considered the characteristic component of magnetization. Final acceptance of the characteristic magnetic directions as primary or original directions is usually deferred until a suite or suites of specimens has passed various field stability tests.

There is considerable disagreement as to how to calculate a mean direction for a suite of specimens. When a good cluster of directions is observed, the calculation is a straightforward unit vector addition to obtain the resultant vector, as outlined by McElhinny (1973), with confidence limits determined using Fisher (1953) statistics. In the instances where the data points have an unusually large dispersion, Watson's (1956) test can be applied to determine whether the observed distribution could be the result of sampling a random population. Data sets that are statistically random are dismissed as magnetically unstable. The disagreement arises when a relatively tight cluster of directions with a few highly disparate directions is observed.

Several investigators (for example, Doell and Cox, 1972, Stone and Packer, 1979) have applied the criterion that a direction lying more than 40 or 50 degrees from the mean is probably not representing a sampling of the axial geocentric dipole field and should therefore be rejected from the final mean. This practice has drawn harsh criticism.

Although the precise value of the circle of rejection is debatable (Doell and Cox, 1972, Watkins, 1973, McElhinny and Merrill, 1975), there is justification for its use. McElhinny and Merrill (1975) have compiled studies of secular variaton over the past 5 m.y. and found that the VGP angular dispersion never exceeded 25 degrees and rarely exceeded 20 degrees. Bingham and Stone (1972) reported secular variation studies from Aleutian lava flows in which all flow mean directions were less than 30-35 degrees from the mean. More recently, Verosub (1979) reported data from Pleistocene glacial lake sediments in New England where individual directional data were less than 30 degrees from the mean. Values of magnetic direction that greatly exceed these values for secular variation can be explained as non-dipole field behavior, failure of the rock to record the true field direction or errors in measuring and recording sample orientation data. Thus, excluding data greater than 40 degrees and certainly 50 degrees is justified as it is probably not a representative of the target population (the geocentric dipole field).

An alternative to the 40 degree or 50 degree circle of rejection criterion is to employ the Theta statisic of McFadden (1980). The Theta statistic is a small sample solution for the angle which will be exceeded with a given probability and is the analog of "Student's" test (McFadden, 1980). As used in this study, theta-95 is the angle about the mean which will be excceeded with a 5 percent probability:

that is, a datum lying beyond this sample circle of confidence has only a 5 percent chance of belonging to the sampled population. Data points lying beyond this limit may be eliminated from calculation of the final mean on the basis that the observations were not drawn from the target population (the axial geocentric dipole field). Ideally, any datum rejected on the basis of the theta-95 criterion should be reexamined in order to find evidence supporting the rejection, such as an error in the orientation data, poor paleohorizontal control, evidence for a lightning strike, etc. In all cases of failure of the theta-95 statistic, the laboratory notes, field notes and paleomagnetic specimens were examined to look for a cause of the aberrant datum. (Two out of three specimens rejected by the theta-95 criterion had plausible reasons for aberration based on field relationships.) Reproducibility checks requiring resampling or reexamination of the original outcrop are not usually feasible in Alaskan field work. Since the critical evidence might be related to field observations, it is not reasonable to insist on final confirmation for rejection. The alternative may be to accept erroneous data which could seriously compromise the interpretability of the mean direction. As the theta statistic offers a probabilistic assessment of a datum's validity, it also acts as a safeguard against rejecting good data.

The next several chapters report the data sets acquired during the course of this study. The general format is a discussion of previous work, important local field relationships, age determination and correlation with other units, followed by a discussion of the paleomagnetic data. This approach was taken because the field data are vitally important to the analysis and interpretation of the paleomagnetic data.

Skolai Creek Data

Three localities within the Skolai Group (WRG.15, 16, 17) were sampled along the northern bank of Skolai Creek in the McCarthy C-4 quadrangle, southern Wrangell Mountains (figures 4, 5, 6). The C-4 quadrangle was mapped by MacKevett (1970). MacKevett's mapping shows bedding attitudes that strongly suggest that the late Paleozoic Skolai Creek rocks are part of the same structural block that contains Hillhouse's (1977a) Golden Horn paleomagnetic sample locality in the Triassic Nikolai Greenstone. Because they are in the same structural block, the Skolai Group poles may be compared with the Nikolai Greenstone poles without correcting for declination anomalies caused by dirrerential tectonic rotation. The data from Skolai Creek reported in this chapter were reported by Panuska and Stone (1981) in summary form. The data are presented again in order to give more complete details of the geology, analytical methods and interpretation.

Sample Localities

The first Skolai Creek locality (WRG.15) was confined to the upper part of the Hasen Creek Formation. The Hasen Creek Formation is abundantly fossiliferous and is Permian, probably Early Permian in age (Smith and MacKevett, 1970). Twenty-three paleomagnetic samples were collected over a 45-meter thick section of fine to coarse grained marine sandstones deposited in shallow water. Sandstones of the Hasen Creek Formation are arkosic or lithic in nature (Smith and MacKevett, 1970). The sandstones are intruded by a few thin felsic sills. These sills are similar to the late Tertiary felsic hypabyssal rocks

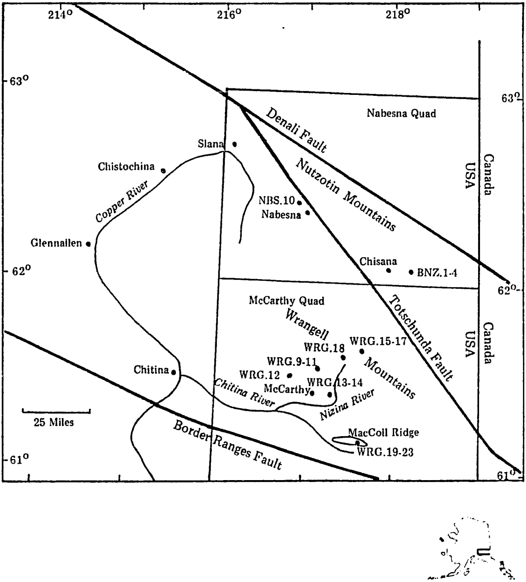

Figure 4 Map showing distribution of sample localities in the Wrangell and Nutzotin Mountains, southcentral Alaska.

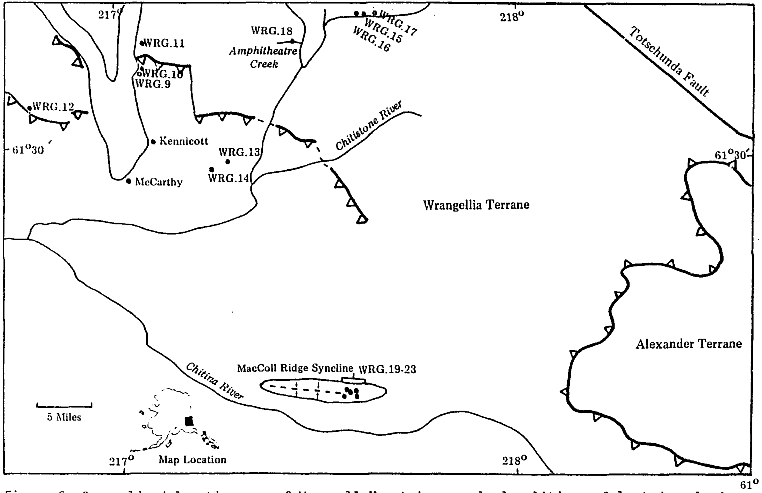

Figure 5 Generalized location map of Wrangell Mountains sample localities. Selected geologic features adapted from MacKevett (1978).

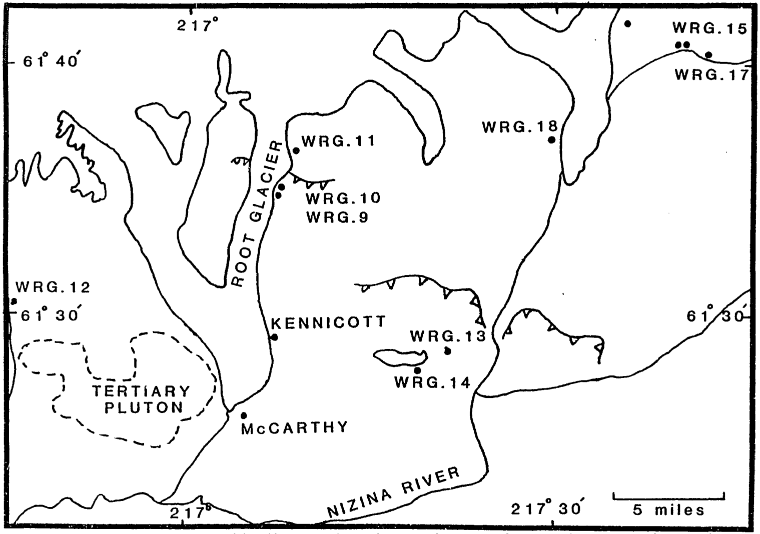

Figure 6 Map showing geographic distribution of central Wrangell Mountains sample localities.

described by MacKevett (1970). Based on lithologic similarity, the sills are probably late Tertiary in age. Four of the twenty-three samples were drilled from the sills in order to check for remagnetization of the sandstones.

Approximately 600 meters to the northwest of the Hasen Creek Formation locality, on an unnamed tributary to Skolai Creek, a small gabbro pluton was sampled (WRG.16). Six cores were drilled over a horizontal distance of about 10 meters as a check on whether the pluton remagnetized the Hasen Creek Formation. MacKevett (1970) originally assigned such gabbro plutons a Permian to Middle or Late Triassic age on the basis that they intrude the Hasen Creek Formation and gabbro clasts are included in the basal conglomerate of the Nikolai Greenstone. In a subsequent publication, MacKevett (1978) reports that the gabbro intrudes the lower portion of the Nikolai Greenstone in a few places and reinterprets the gabbro as precursor and feeder plutons of the Nikolai Greenstone. If the plutons are viewed as being coeval with the Nikolai Greenstone, a Ladinian and/or Carnian (Middle and/or Late Triassic) age may be assigned.

The last locality in the Skolai Group is located about 1.7 kilometers southeast of the Hasen Creek sandstone locality in the upper portions of the volcaniclastic member of the Station Creek Formation (WRG.17). This member of the Station Creek Formation is composed of coarse volcanic breccia, tuff-breccia, volcanic sandstone and volcanilutite (Smith and MacKevett, 1970). The locality sampled consists almost entirely of volcanic breccia with clasts ranging in size from a few millimeters to one centimeter. These rocks have been mildly metamorphosed to the prehnite-pumpellyite facies (Smith and MacKevett, 1970).

Seventeen samples were collected from the volcanic breccia over a stratigraphic interval of 21 meters. Bedding was difficult to discern at this outcrop and was recognized by subtle changes of texture and color. Indications of bedding were not always present. As a result, bedding attitudes for a few samples had to be extrapolated from neighboring samples. No fossils have been found in the Station Creek Formation. Smith and MacKevett (1970) considered the formation to be Permian(?) based on the gradational nature of the contact with the overlying Hasen Creek Formation. More recently the Station Creek Formation was assigned a Middle Pennsylvanian to Early Permian age based on correlation with the Mankomen Group and the Tetelna volcanics (MacKevett, 1978).

Paleomagnetic Data

All specimens of the Skolai Group were subjected to stepwise thermal demagnetization. Eight of the twenty-three Hasen Creek Formation samples were discarded on the basis of poor demagnetization characteristics. Specimens with stable demagnetization endpoints had blocking temperatures that ranged between 450 and 550 degrees C (table 1). Final vector directions corresponding to 400 degrees C demagnetization were selected for a few specimens as higher demagnetization levels lost too much of the original remanence and began to show unstable directions. The Hasen Creek specimens contained a secondary component of 12 degrees declination and 74 degrees inclination

Table 1

Hasen Creek Formation (WRG.15)

Longitude: 217.7
Latitude: 61.7

Sample	Demag Level*	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat
751	400	17	-13	25	-12	11	20
752	400	25	-8	29	-4	6	23
753	400	25	-43	51	-31	349	2
755	500	14	-17	31	-27	7	10
756	500	12	-19	33	-32	7	7
760	500	28	-36	56	-31	345	0
761	450	47	-31	64	-16	334	5
762	450	35	0	35	0	359	23
763	500	29	-11	41	-15	356	14
768	400	33	-12	42	-20	356	11
	Mean	26	-19	40	-19	357	12
	K=	23.7		24.1		33.0	
	α_{95}	10.1		10.0		8.0	
	N=	10		10		10	

*Thermal Demagnetization in degrees C.

(K=102.6) before bedding correction. This secondary component was removed at temperatures of 200 to 300 degrees C.

Specimens from the Tertiary(?) sills lost all but a few percent of their original intensities by 300 degreec C. The geographic vector directions of the sills are not significantly different from the secondary component observed in the sandstones and not significantly different from a recent geomagnetic field direction. This suggests that the secondary component is a late Tertiary or Quaternary magnetization. Regardless of the age of the felsic sills, the secondary components of magnetization in the sandstones which parallel the directions from the sills were easily removed by thermal cleaning. This constitutes a modified baked contact test.

The ideal baked contact test requires that systematic changes in magnetic directions be demonstrated across an igneous contact into the country rock (Irving, 1964). Where time or geologic constraints prevent the necessary serial sampling to apply the true baked contact test, samples may simply be collected from the intrustion for comparison with samples from the local suite. This sampling scheme is here termed a "modified baked contact" test.

A second modified baked contact test was applied using the data from the gabbro pluton (table 2). Two specimens had blocking temperatures between 500 and 550 degrees C. The other four specimens had blocking temperatures exceeding 550 degrees C. The remanence directions of the pluton specimens cluster very tightly and are significantly different from the observed direction of the Hasen Creek sandstones. This affords a second check against remagnetization.

Table 2

Gabbro Pluton (WRG.16)

Longitude: 217.6
Latitude: 61.7

Sample	Demag Level*	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat
~~774~~	~~550~~	~~57~~	~~-39~~	~~66~~	~~-4~~	~~330~~	~~9~~
775	550	59	-32	65	2	329	13
776	550	46	-30	54	0	340	16
777	500	62	-33	68	3	327	12
778	500	54	-29	60	3	334	15
779	500	57	-20	59	13	333	20
	Mean	56	-31	62	3	332	14
	K=	107.9		111.0		192.1	
	$\alpha95$	5.5		5.4		4.1	
	N=	6		6		6	

*Thermal Demagnetization in degrees C.

No evidence for ancient horizontal was observed within the pluton. However, a regional bedding correction was applied on the basis that the Skolai Group and Nikolai Greenstone have approximately the same attitude in this area and that the pluton was probably emplaced before the end of Nikolai time (MacKevett, 1978). The correction applied to all samples of the pluton was 275 degrees (AB), 41 degrees (DB) (following the convention: AB=azimuth of the down dip direction, DB=dip of bedding). All tilt corrections performed as a rotation around the present strike unless otherwise stated. The tilt corrected VGP is included for the sake of comparison and is not purported to be a reliable paleomagnetic pole.

The Station Creek Formation specimens tended to respond well to thermal cleaning. In general, blocking temperatures were in the 500 to 550 degree C range (table 3). Of the seventeen samples collected from the Station Creek Formation, six failed to yield a reliable vector direction. One additional datum, from specimen 784, was rejected by the theta-95 criterion, although the demagnetization characteristics were good. This was one of two samples whose bedding correction was extrapolated from adjacent samples. The rejection of specimen 784 and the higher scatter in the Station Creek data than the Hasen Creek data is probably due to the difficulty in discerning bedding, the presence of unrecognized prelithification slumping or the failure of the bedding to faithfully record ancient horizontal. There is one reversal in this suite of specimens.

There is some concern that the post-Nikolai metamorphism (prehnite-pumpellyite facies) may have reset the magnetic remanence. Sil-

Table 3

Station Creek Formation (WRG.17)

Longitude: 217.7
Latitude: 61.7

Sample	Demag Level*	Geographic Dec	Geographic Inc	Statigraphic Dec	Statigraphic Inc	VGP Long	VGP Lat
780	500	44	-20	56	-27	344	2
781	450	34	45	20	2	15	28
782	500	29	-16	47	-36	355	0
783	400	46	8	47	-11	349	14
784	450	(340	12)	(340	-31)	(57	10) θ
785	500	59	38	36	28	352	36
787	500	77	9	71	17	321	17"R"
790	450	46	-37	61	-24	339	2
791	500	21	-2	23	-6	13	23
793	500	50	34	31	22	0	35
785	500	45	17	41	-1	354	21
	Mean	45	8	43	-4	352	18
	K=	7.1		9.2		16.6	
	α_{95} =	19.5		16.8		12.2	
	N=	10		10		10	

*Thermal Demagnetization in degrees C.

θ -Rejected by θ_{95} statistic.

R-Reversed

berman and others (1980) have inferred a metamorphic temperature of approximately 200 degrees C for the Nikolai Greenstone, based on quartz/epidote oxygen isotope partitioning data from 3 veins. They considered the veins to have formed at the same temperature as the metamorphic mineralogy of the Nikolai Greenstone because: the mineral assemblages in the greenstone do not vary with proximity to the veins, the veins lack selvages (which might indicate thermal disequilibrium), the veins lack zones of metamorphic mineralogy that differ from that of the adjacent greenstone and their K-Ar metamorphic age determinations show no variation with proximity to the veins. Assuming that the vein mineralogy reflects the metamorphic conditions and not a later event, Silberman and others (1980) calculated metamorphic temperatures based on oxygen isotope partitioning in natural quartz/epidote systems with temperatures determined from independent evidence. Unfortunately, there is no experimental confirmation of quartz/epidote isotopic partitioning behavior.

The effect of heating the Skolai Group rocks to these metamorphic temperatures can be crudely estimated by using the relaxation time/temperature curves of Pullaiah and others (1975). A temperature of 200 degrees C maintained for a 10 M.Y. period could reset those magnetite grains with a blocking temperature of about 370 degrees C. The thermal demagnetization curves of the Skolai Creek specimens (figure 7) indicate maximum blocking temperatures of at least 450 degrees C, with the majority of specimens exceeding 500 degrees C. Grains with blocking temperatures of 500 degrees C subjected to a 200 degree C temperature for a duration of 10 m.y. plot well within the stable primary

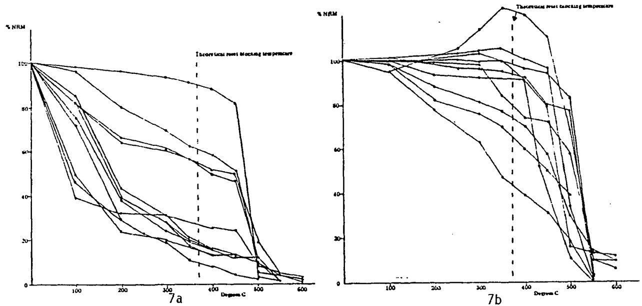

Figure 7 Thermal demagnetization curves for WRG.15 (figure 7a) and WRG.17 (figure 7b) showing high blocking temperatures. Blocking temperatures exceed the theoretical resetting temperature; therefore, the magnetic carriers probably record a primary remanence.

remanence field. Since reset temperature varies as the log of the relaxation time multiplied by a constant, changing the duration of the heating even within geologically plausible limits will have little effect on the estimate of the blocking temperature stability limit. Therefore, it is very unlikely that the inferred metamorphic temperatures could have seriously affected the stability of the primary remanence.

The calculated metamorphic temperatures are based on several assumptions. Silberman and others (1980) considered that the temperature could have been as high as 300 degrees C, based on the metamorphic mineralogy. Even this higher metamorphic temperature does not alter the conclusions of magnetic stability based on the relaxation time/temperature curves of Pullaiah and others (1975).

Despite the greater scatter in the Station Creek Formation data, the mean directions of the Station Creek Formation locality and the Hasen Creek Formation locality are virtually identical (tables 1 and 3). The remanence directions for these two formations, without bedding corrections, are plotted in figure 8 along with the uncorrected mean direction for Hillhouse's (1977a) Golden Horn locality. The circles of confidence for these directions do not intersect. They are distinctly different. Thus, there is no evidence that the eruption of the Nikolai Greenstone or the emplacement of any of the intrusives has remagnetized the Skolai Group. In addition, the presence of one reversal in the Station Creek Formation, suggests that the remanence is primary.

The Hasen Creek Formation was deposited during the Permian and

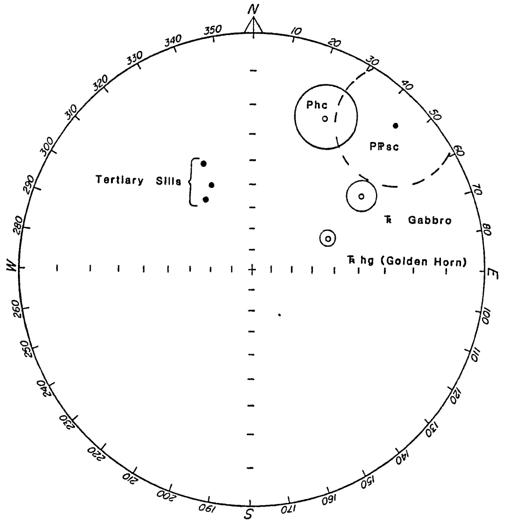

Figure 8 Geographic directions for Skolai Creek data. Phc-Hasen Creek Formation, PPsc-Station Creek Formation, TRg-gabbro intrusive. Golden Horn Locality is from Middle and/or Late Triassic Nikolai Greenstone. Since the gabbro, the greenstone and the Tertiary sills are statistically distinguishable from the Paleozoic directions, there is no evidence that the intrusives or the eruption of the flows remagnetized the Paleozoic rocks.

the Station Creek Formation was deposited in the Permian and/or Pennsylvanian time. This corresponds to the Permo-Carboniferous reversed (PCR) magnetic polarity interval established by Irving and Parry (1963). A later compilation by Irving and Pullaiah (1976) shows several short-lived intervals of normal polarity in the PCR. With the exception of one specimen from the Station Creek Formation, all data from the Skolai Group are of one polarity. The probability of sampling only normally magnetized beds over a large stratigraphic section deposited during the PCR is very small. Therefore, the estimate of the true north VGP (corresponding to the Permian north geographic pole) and the true paleolatitude is almost certainly given by the poles and directions antiparallel (reversed) to those observed and reported in tables 1 and 3.

These data for the Skolai Group were previously reported by Panuska and Stone (1981). The data were originally analyzed according to the "closest approach" procedure outlined by Stone and Packer (1979). Although the methods of analysis differed, the recalculated mean directions are not significantly different.

Tetelna Volcanics Data

The Tetelna Volcanics were sampled in the bed of an unnamed creek approximately 2.5 km northwest of Skookum Creek, in the Nabesna B-5 quadrangle (NBS.10, figure 4). These rocks were mapped as Tetelna Volcanics by Richter (1976).[1] Richter and Dutro (1975) assigned the Tetelna Volcanics a Pennsylvanian age based on Middle Pennsylvanian fossils found in the conformably overlying Mankomen Group. Richter and Schmoll (1973) found Permian fossils in the upper part of the Tetelna Volcanics in the Nabesna C-5 quadrangle. Thus, the age of the Tetelna Volcanics may be taken as Pennsylvanian to Permian (Richter, 1976). These rocks are correlated with the Station Creek Formation (Richter and Dutro, 1975, Smith and MacKevett, 1970, MacKevett, 1978). Forty-one samples were collected from nine lava flows over a distance of 70 meters, spanning about 14 meters of stratigraphic thickness. These rocks are weakly metamorphosed porphyritic basalts or andesites, many displaying crude columnar jointing. Individual flows were identified by subtle textural changes and by planar features accentuated by erosion. The slightly more weathered planar zones are up to a few centimeters thick, are brecciated and are interpreted as flow boundaries. The upper portion of some flows have a finer grained texture and smaller phenocrysts than the flow interiors or bases. The average

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[1]Since this dissertation was accepted by the committee, a more detailed map (Lowe and others, 1982) has become available, which designates these rocks as Lower Permian, Mankomen Group. This raises some doubts that these rocks are actually misinterpreted sills of Triassic age. It is possible that the brecciated layers are actually slightly more weathered pyroxene crystals which mimic volcanic rubble. Additional investigation to resolve this conflict is in progress.

strike and dip of the flows is essentially parallel to the attitude of some well-bedded sedimentary rocks a few meters below the volcanics; however, the contact between the sediments and the volcanics was not observed.

In thin section, the volcanics are coarsely porphyritic clino-pyroxene, plagioclase (labradorite-bytownite) basalts(?) with sub-ophitic to ophitic texture. The plagioclase is virtually unaltered. The groundmass and portions of some pyroxene grains are altered to chlorite. Some chlorite contains thin stringers of pumpellyite(?). Elsewhere the Tetelna Volcanics have chlorite, epidote and pumpellyite as alteration products (Richter, 1976). These rocks probably have not been metamorphosed beyond prehnite-pumpellyite facies. Fairly abundant euhedral opaque minerals, observed in reflected light, are probably magnetite. Identification of the (311), (400) and (511) X-ray reflections and a calculated unit cell edge of 8.39 Angstroms suggests end member magnetite.

Paleomagnetic Data

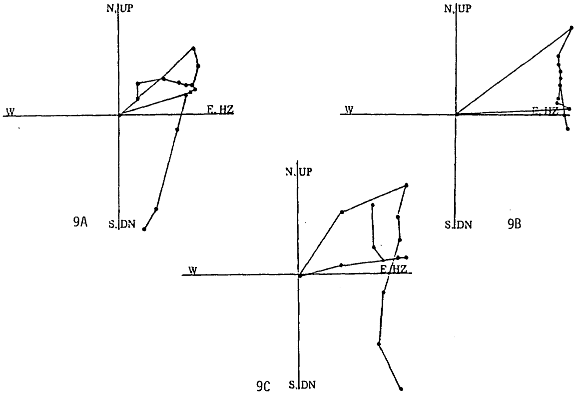
Specimens from the Tetelna Volcanics were subjected to stepwise thermal cleaning. Blocking temperatures ranged from 450 degrees to greater than 550 degrees C, with the large majority of blocking temperatures in the 500 to 550 degrees C range. The blocking temperature spectra of these specimens are unually narrow. Commonly the magnetic intensities would drop from 100 percent NRM (natural remanent magnetization - the before-cleaning intensity) at 500 degrees C to less than 10 percent NRM at 550 degrees C.

In general, the response of the Tetelna Volcanics to thermal cleaning followed one of three patterns (figure 9). The most common pattern is exemplified by specimen 817A (figure 9a). In this instance, the specimen lost virtually all of its primary remanence in one heating increment and the directions measured before and after this heating increment are substantially different. The second pattern (figure 9b) shows a similar loss of intensity in one cleaning interval but the directions before and after heating to the blocking temperature are essentially the same. The third pattern (figure 9c) displays a slightly wider blocking temperature spectrum and has an intermediate magnetic direction on the rapid intensity decay trend. Both the second and third cleaning patterns allow a fair degree of confidence in determining the characteristic direction of the specimen.

The pattern displayed by specimen 817A gives an ambiguous direction because it is not certain the the last direction measured before the loss of remanence is actually free of secondary components. This is the most common behavior in the suite of Tetelna thermally demagnetized specimens. Thus, the mean direction of 80 degrees declination,-38 degrees inclination (K=116.3), is not considered a very reliable estimate of the characteristic remanence direction.

In order to obtain a better estimate of remanence direction, the replicate specimens of the Tetalna Volcanics were cleaned by AF demagnetization. The specimens responded well to AF cleaning. In general, fields of 200 oersteds were sufficient to remove secondary magnetizations. The direction of the mean secondary component before bedding correction is 47 degrees declination, 75 degrees inclination (K=19.3),

Figure 9 Vector diagrams for selected specimens of Tetelna Volcanics, which were subjected to thermal demagnetization. Vector diagrams are double orthogonal plots of declination and inclination demagnetization data. Bold lines are declinations plotted in the north-south-east-west plane; fine lines are inclinations plotted in a vertical (up-down-horizontal) plane. a) Plot of specimen 817A showing a narrow blocking temperature spectrum. Since there are no intermediate points on the decay curve towards the origin and since the pre- and post-blocking temperature directions are different, the characteristic direction is poorly known. b) Vector diagram of specimen 819A showing good agreement of pre- and post-blocking temperature directions; thus, the characteristic direction is fairly well known. c) Demagnetization plot of specimen 829A shows a slightly wider blocking temperature spectrum than previous specimens. One intermediate point on the decay curve of the characteristic component gives moderate confidence in the value of characteristic direction. Since the behavior displayed by specimen 817A typifies the thermally demagnetized Tetelna suite, the mean direction of thermal data is not considered to be an accurate measure of the ancient geomagnetic field.


N, UP
W
E, HZ
S, DN
9A
N, UP
W
E, HZ
S, DN
9B
N, UP
W
E, HZ
S, DN
9C

which is approximately the direction of the present geomagnetic field. Several specimens began to show signs of RRM acquisition at coercive forces of 150 oersteds and above. In most of these cases, the effects of RRM could be cancelled by averaging vectors with positive and negative RRM (figure 10, specimen 820 B).

Of the AF cleaned specimens, 35 gave stable demagnetization end points and 4 failed to arrive at a stable end point. The mean magnetic direction of flows and flow means is shown in table 4. These specimens displayed several intermediate points on Zijderveld plots, showing the decay of the primary vector. Therefore, these data are considered to give a more accurate estimate of the characteristic magnetic direction than the thermal data. Recognition of incomplete isolation of the characteristic direction is important inasmuch as the mean directions of the thermal and AF data sets are significantly different at the 95 percent confidence level.

These data can only be tentatively accepted as a measure of the ancient geomagnetic field as there are no independent stability tests. A fold test was not possible and reversals are not to be expected as the flows were erupted during the Permo-Carboniferous Reversal interval (PCR). In addition, since only 9 flows were sampled, secular variation may not have been adequately averaged out, as the flows could have been extruded in a geologically short period of time. However, the mean VGP is similar to the Skolai VGPs and is thus considered as supportive evidence for the validity of the Tetelna VGPs.

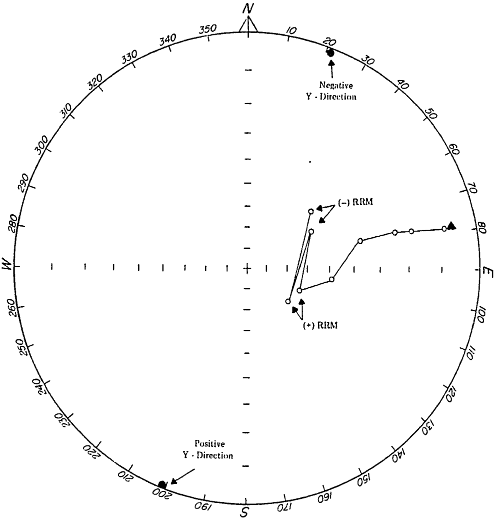


Figure 10 Stereographic plot of AF demagnetization data for specimen 820B in the geographic reference frame. The large triangle is the NRM direction with subsequent points on the curve representing successively higher demagnetizing fields. Positive RRM is observed in this specimen as a deflection of the vector towards the +Y axis of the core and negative RRM is observed as a deflection towards the -Y core axis. See methods section (page 24) for a complete description of RRM.

Table 4

Tetelna Volcanics (NBS.10)

Longitude: 216.9
Latitude: 62.4

Flow	Demag Range*	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K	VGP Long	VGP Lat	K
1	100-200	5	76	-48	43	80	-26	42	322	-9	56
2	250-300	2	79	-60	384	85	-35	366	321	-15	393
3	100-200	4	67	-51	31	75	-28	28	327	-5	31
4	150-200	4	76	-43	30	78	-25	30	323	-7	34
5	100-200	2	55	-48	11	65	-24	11	334	-1	11
6	100-200	6	56	-52	16	70	-34	16	335	-8	18
7	150-200	3	60	-59	66	75	-34	77	330	-9	64
8	200-250	4	74	-51	120	80	-29	120	323	-9	123
9	150-300	5	72	-46	43	77	-22	42	323	-4	61
	Mean		68	-51		76	-29		327	-8	
	K=		102.1			132.8			151.0		
	α_{95} =		4.6			4.0			3.8		
	N=		9			9			9		

*Alternating Field Demagnetization in Oersteds.

Nizina Limestone

The Nizina Limestone was sampled in reconnaissance fashion in the southeast corner of the McCarthy C-6 quadrangle, locality WRG.9 (figures 4, 5, 6). The McCarthy C-6 quadrangle was mapped by MacKevett (1972). The shallow marine Nizina Limestone is late Karnian to early Norian in age (Late Triassic) (Armstrong and others, 1969). At locality WRG.9, 15 samples were collected over 21 m of stratigraphic section.

The WRG.9 specimens were cleaned using AF cleaning techniques. Of the 15 specimens treated, 6 specimens displayed no stable demagnetization end point and 6 specimens showed signs of incomplete demagnetization before remanence intensities were too low to measure reliably. Five of the six incompletely demagnetized specimens grouped moderately well; however, due to a lack of intermediate points showing intensity decay with little directional change, these vectors could not be positively identified as a primary components. Two of the three specimens with good demagnetization behavior gave geographic vector directions similar to recent field directions. No secondary components were observed other than a present field direction (30 degrees declination, 73 degrees inclination, k=19.0). Due to the problems of isolating reliable vector directions, these data are not considered further.

McCarthy Formation

The lower member (informal usage) of the McCarthy Formation was sampled in a reconnaissance manner at three localities in the McCarthy B-5 and C-6 quadrangles (figures 5 and 6). The geology of these quadrangles has been mapped by MacKevett (1972, 1974). The lower McCarthy Formation consists of calcareous, carbonaceous shale, impure limestone and impure chert deposited in a relatively deep marine starved basin (MacKevett, 1971). Upper Norian (Upper Triassic) fossils occur in the lower part of the lower member and Hettangian (lowermost Jurassic) fossils have been found in the upper member (MacKevett, 1971). Thus, the lower memeber of the McCarthy Formation is Late Triassic to Early Jurassic(?) in age (MacKevett, 1971).

Each of the 3 McCarthy Formation sample localities is discussed separately.

WRG.10

The lower McCarthy Formation was first sampled along the east flank of the Root Glacier in the McCarthy C-6 quadrangle, WRG.10 (figure 6). At this locality, 16 cores were collected over approximately 19 m of section. The samples were drilled from massive impure limestone beds up to one meter thick. Limestone beds are separated by interbeds of shale less than 20 cm thick.

Specimens from WRG.10 were subjected to AF cleaning techniques. A well defined secondary component was removed from the specimens at coercive forces up to 100 oersteds. This component of magnetization (declination = 12, inclination = 82, K = 27.8 in the geographic reference frame) is not significantly different than the present field. At coercive forces exceeding 100 oersteds probable RRM was encountered and no stable endpoint could be reached. Thus, it is likely that these specimens were remagnetized along a recent field or, if a primary remanence direction exists, it cannot be observed due to the obscuring effects of RRM.

WRG.12

The second suite of samples from the McCarthy Formation was collected from the eastern valley wall along Fohlin Creek, in the southwestern corner of the McCarthy C-6 quadrangle (figure 6). Twenty-one samples were drilled along the creek bed for a distance of approximately 200 m. Accurate determination of the stratigraphic interval was precluded by the presence of two high-angle faults.

A minor fault separates samples 683 and 684. There appears to be approximately 3 m of offset, north side down. A much more prominent fault (mapped by MacKevett, 1972) separates sample 690, on the north side of the fault, from sample 691, located some 30 m to the south. The amount or sense of displacement along this fault could not be determined, but there is no doubt that both sides of the fault represent the McCarthy Formation. About 25 m of section is represented by samples collected to the north of the fault. The 8 samples from south of the fault span a 9 m stratigraphic interval.

All specimens from WRG.12 were cleaned by AF techniques. Five specimens failed to clean to stable endpoint directions. The remaining 16 specimens cleaned exceptionally well, displaying both low coercivity (50 oersteds) secondary components and higher coercivity (100-800 oersteds, averaging 500 oersteds) directions. The mean of the secondary components in the geographic reference frame is 33 degrees declination, 78 degrees inclination (k = 66.6), and probably represents a recent field magnetization. The characteristic higher coercivity components are more difficult to interpret. These directions form a markedly elongate distribution in the geographic reference frame (figure 11, table 5). The specimens collected to the south of the prominent fault cluster very tightly (with the exception of specimen 698 which failed the theta-95 criterion) yielding a 169 degree declination, -70 degree inclination (k = 238.6). The specimens collected north of the fault form an elongate pattern trending away from the mean direction of specimens south of the fault. In general, the farther a specimen was collected from the fault, the farther it plotted from the mean of the specimens from south of the fault.

The observed arrangement of remanence directions strongly suggests that remagnetization has occurred. The elongate distribution of remanence directions most likely represents a sequence of specimens with a successively stronger partial remagnetiztion. The well-grouped mean of specimens south of the fault would then be an end member in the remagnetized sequence; however, there is no a priori way of knowing whether this mean is the completely remagnetized direction of the unaffected primary direction. There is indirect evidence that the group of data from south of the fault is the remagnetized direction. The specimens from south of the fault are the closest to the Fireweed Mountain pluton, a possible cause for the remagnetization. On

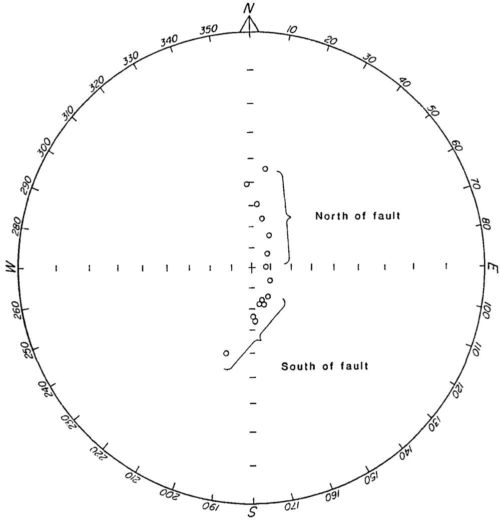


Figure 11 Geographic vector directions for WRG.12, McCarthy Formation. Elongate distribution of vector directions probably results from partial thermal remagnetization of the specimens. Circles are negative inclinations; dots (not shown on this plot) are positive inclination.

Table 5

McCarthy Formation (WRG.12)

Longitude: 216.8
Latitude: 61.5

Sample	Demag Level*	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat
677	100	8	-44	5	-16	31	20
678	200	357	-51	356	-18	41	19
679	400	29	-72	6	-47	32	0
680	400	5	-60	2	-24	34	16
683	200	12	-66	8	-43	30	3
687	200	50	-80	5	-50	33	-2
689	400	122	-79	19	-66	23	-20
690	600	86	-83	6	-56	32	-8
691	400	179	-66	230	-82	70	-68
692	600	162	-71	7	-88	36	-58
693	800	162	-72	348	-86	40	-53
694	800	168	-72	333	-78	51	-40
695	800	177	-64	292	-82	61	-53
697	800	152	-74	353	-76	40	-36
698	600	197	-48	228	-68	126	-63

No tally of average declination or inclination.

*Alternating Field Demagnetization in Oersteds.

the whole, these specimens have five times the magnetic intensity of the specimens from north of the fault, suggesting remagnetization. The increase of intensity may be due to better alignment of the magnetic carriers or to introduction of new magnetic carriers. In addition, the geographic mean direction of the southerly specimens is not significantly different from a reversed late Tertiary field, assuming no post-remagnetization tilting. The age of the Fireweed Mountain pluton is probably late Miocene (MacKevett and Smith, 1972), based on an 8.4 m.y. age of similar plutons in the area. In addition, the specimens collected farthest from the southerly specimens gave stratigraphic directions similar to the stratigraphic directions of the WRG.13 specimens (described in the next section). Since the WRG.13 locality was almost certainly not remagnetized along a late Tertiary field direction, the evidence suggests that the mean direction of the specimens south of the fault was the prevailing geomagnetic field direction during the remagnetization. Thus, the observed remanence directions are consistent with a remagnetization of the specimens south of the fault and a partial remagnetization of specimens north of the fault by the Fireweed Mountain pluton. Given the assumptions and absence of measured remanence directions from the pluton, the scenario of remagnetization by the pluton should be regarded as conjecture. However, these data are clearly not a valid measure of the time-averaged geomagnetic field.

WRG.13

The WRG.13 locality is located approximately 2.5 km ENE of the Nikolai Mine (abandoned) in the McCarthy B-5 quadrangle (figure 6). At this locality, the lower member of the McCarthy Formation was sampled over a stratigraphic interval of 65 m, from which 31 core samples (yielding 56 specimens) were collected. The samples were drilled from thickly bedded impure limestone. The limestone beds are separated by thinly bedded shale and siltstone.

Specimens from this collection were subjected to either AF or thermal demagnetization. Most specimens contained a secondary component of magnetization essentially parallel to the present field, which was removed by cleaning at 100 oersteds or 200 degress C. Of the 31 specimens cleaned by AF techniques, only 11 specimens yielded reasonably well-defined characteristic magnetization directions before RRM obsured the remanence. Eleven of the 25 thermally cleaned specimens gave characteristic magnetization directions at unblocking temperatures of 350 to 400 degrees C. The remaining specimens either failed to clean to a stable end point, became too weak to measure reliably or were incompletely cleaned because the epoxy holding the cores together decomposed at high temperatures.

The directions obtained by AF and thermal cleaning are comparable (table 6). The distribution of directions for both AF and thermal data sets is fairly dispersed (figure 12), although it is significantly different from a random distribution (by the test of Watson, 1956). The precision parameter for both data sets improves slightly when the bedding correction is applied; however, the improvement is not sufficient to constitute a positive fold test.

There is no evidence for a late Tertiary/Quaternary remagnetization. The mean geographic direction is significantly different from

Table 6

McCarthy Formation (WRG.13)

Longitude: 217.3
Latitude: 61.5

Sample	Demag Level*	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat	
699	350	33	-29	33	-44	8	-1	
703	350	33	-33	36	-47	6	-4	
704	400	43	-38	48	-52	357	-12	
707	350	38	-19	40	-34	0	4	
710	400	32	-29	37	-44	5	-2	
717	350	36	7	36	-8	359	19	
718	400	33	8	33	-7	2	20	
719	350	34	-20	35	-35	4	50	
724	400	25	11	25	-5	9	23	
726	400	23	29	23	6	11	29	
728	400	(322	36)	(328	28)	(77	39)	θ
	Mean	33	-12	34	-27	4	13	
	K=	12.0		13.7		17.7		
	α_{95} =	14.5		13.5		11.8		
	N=	10		10		10		

*Thermal Demagnetization in degrees C.

θ -Rejected by θ95 statistic.

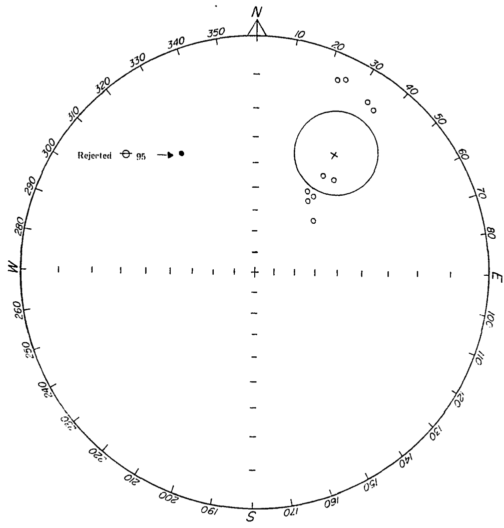


Figure 12 Stratigraphic vector directions from WRG.13, McCarthy Formation. Mean direction is shown by a "+" with its associated circle of 95 percent confidence. The datum in the northwest quadrant is rejected on the basis of the theta-95 statistic.

the present field; indeed, a present field component is removed during cleaning. Numerous Upper Miocene(?) felsic hypabyssal intrusives, with remagnetization potential, occur within the area (MacKevett, 1974). The WRG.13 locality is approximately 2 km from the nearest mapped intrusive and the observed geographic remanence direction is markedly different from any reasonably expected late Tertiary direction.

The mean stratigraphic remanence direction is similar to other Triassic directions (Hillhouse, 1977), suggesting a primary remanence. However, there are no positive stability tests - only negative evidence suggesting stability. Moreover, the lack of reversals in a 65 m thick sedimentary section is disturbing. Thus, the primary nature of the observed paleomagnetic direction remains in question and can only be tentatively accepted as valid.

Root Glacier Formation

The Root Glacier Formation was sampled at one locality along the western flank of Bonanza Ridge in the McCarthy C-5 quadrangle (WRG.11, figure 6). The geology of this area was mapped by MacKevett (1970). The Root Glacier Formation consists of shallow marine siltstone, mudstone and less abundant sandstone. Based on fossil evidence the Root Glacier Formation is late Oxfordian and Kimmeridgian (Late Jurassic) in age.

The Root Glacier Formation was sampled in reconnaissance fashion near the eastern margin of the Root Glacier. Fourteen samples were collected over an 18 m thick sedimentary section. In addition, 3 samples were collected from Tertiary (?) dikes and sills, which intrude the sedimentary rocks.

A complete set of Root Glacier specimens was cleaned by AF techniques. Eleven replicate specimens were thermally demagnetized. Due to a computer software problem, the AF data were overwritten by the thermal data file and lost. Due to disintegration problems at high temperatures only a few specimens could be measured after heating above 400 degrees C. No stable end points of demagnetization were observed at the higher unblocking temperatures. Six specimens did display fairly well-defined components of magnetization at unblocking temperatures of 300 to 450 degrees C (table 7). With the exception of one direction from a specimen (660 B) collected within an inferred soft sediment slump block, the directions clustered rather tightly. (Specimen 660 B probably represents an incompletely cleaned

Table 7

Root Glacier Formation (WRG.11)

Longitude: 217.1
Latitude: 61.6

Sample	Demag Level*	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat	
661	300	(126	-51)	(143	-52)	(273	-53)	θ
662	350	2	76	26	65	341	69	
663	350	37	87	48	73	292	67	
665	450	32	60	22	49	3	55	
666	350	13	79	3	65	30	75	
668	300	74	86	7	75	301	86	
	Mean	24	78	21	66	344	74	
	K=	48.4		45.0		24.3		
	α_{95} =	11.1		11.5		15.8		
	N=	5		5		5		

*Thermal Demagnetization in degrees C.

θ -Rejected by θ_{95} statistic, occurs within slump block.

secondary direction.) The mean direction is not significantly different from a recent field direction. The sole specimen from the intrusives for which any paleomagnetic data could be obtained yielded a direction at 500 degrees C, which is nearly antiparallel to the geographic mean of the sedimentary specimens. This suggests that the sediments were remagnetized or at least acquired a low temperature component during the episode of intrusion and that the antiparallel remanence direction of the dike simply reflects the cooling of the dike during a reversal within the remagnetization episode. Unfortunately, the cleaning temperature increments for the dike specimen were not sufficiently small to resolve a stable end point of demagnetization. Thus the 500 degree C direction for the dike could be a fortuitous aberrant magnetization. Considering the low unblocking temperatures, lack of stability tests, recent field direction and the possibiltiy of a magnetization acquired penecontemporaneously with the local intrusions, these data are not considered to be a primary paleomagnetic direction.

Moonshine Creek Formation

Paleomagnetic samples of the Moonshine Creek Formation were collected along the steep walls of Amphitheatre Creek, in the eastern McCarthy C-5 quadrangle (WRG.18, figure 6). MacKevett (1970) mapped the geology of the quadrangle. The shallow marine Moonshine Creek Formation disconformably overlies the Kennicott Formation and ranges in age from late Albian to late Cenomanian (late Early Cretaceous and early Late Cretaceous) (Jones and MacKevett, 1969).

At this locality, WRG.18, 33 cores were obtained from 13 beds over a 45 m stratigraphic thickness. The sampled section spans approximately 170 m of ground distance following the creek. Samples were exclusively of fine to very fine grained sandstones. Most samples contain evidence of burrowing by organisms.

A complete suite of specimens (less the two mislabeled cores) was demagnetized by thermal means. A replicate suite containing 11 specimens was cleaned by AF means. Approximately half of the thermally demagnetized specimens displayed the peculiar intensity spikes or intensity increases at temperatures of 300 to 550 degrees C, along with erratic directional behavior. Sixteen thermally treated specimens cleaned to stable end point directions (table 8). Some of the selected end point directions correspond to unblocking temperatures as low as 300 degrees C. It should, however, be noted that this demagnetization level reflects the direction most representative of the characteristic direction and that similar directions were observed for many specimens up to 450 to 550 degrees C.

Table 8

Moonshine Creek Formation (WRG.18)

Longitude: 217.5
Latitude: 61.6

Bed	Demag Level AF/T	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K	VGP Long	VGP Lat	K
1	/350	1	324	57	--	10	37	--	23	48	--
2	100/300	3	301	76	20	33	50	21	345	52	13
3	100/450	2	321	68	63	23	43	65	5	50	55
4	150/400	2	348	61	46	21	21	49	11	37	66
5	300/550	4	350	76	88	35	37	87	351	43	93
6	200/500	4	36	70	43	45	26	41	343	33	101
8	/400	1	344	71	--	47	31	--	339	34	--
10	/350	1	12	81	--	41	32	--	346	37	--
11	/300	1	88	82	--	53	34	--	332	33	--
13	/400	2	331	74	5737	39	37	7977	347	41	5764
	Mean		345	75		35	36		351	42	
	K=		49.2			35.5			34.9		
	α_{95} =		7.0			8.2			8.3		
	N=		10			10			10		

Note: Data from both Alternating Field and Thermal Demagnetizations.

Five specimens subjected to AF demagnetization cleaned to stable end points. The coercive forces needed to clean these specimens varied from 100 to 300 oersteds. In all cases, the RRM contribution to the final remanence was very minor. The AF data are in good agreement with their thermal counterparts.

The mean directions of individual beds show only small dispersion in either the geographic or statigraphic reference frame. As there was only about 10 degrees of bedding attitude variation at the sampling locality, a fold test is not possible for this suite of specimens. Reversals were not observed; however, none are expected, as these rocks were deposited during the Cretaceous long normal polarity interval. The mean geographic vector direction is indistinguishable from a recent field direction. This warns of the possibility of an unstable or remagnetized direction. While the paleolatitude estimate from the stratigraphic vector direction is comparable to other Cretaceous data with demonstrable stability (MacColl Ridge Formation), nearly 180 degrees of relative rotation would be required to reconcile the declination differences.

There is no compelling evidence for or against a remagnetization. Thus, the characteristic magnetic direction can only be provisionally accepted as a primary estimate of the ancient geomagnetic field direction. However, it will be shown in the Tectonic Analysis section that these data almost certainly represent remagnetized directions.

A diagenetic hypothesis can be proposed as an explanation for the observation of a recent field direction. Abundant pyrite was noticed in many paleomagnetic specimens. Pyrite is probably the result of early diagenesis of ironbearing phases in a reducing environment. Reducing conditions are inferred based on the extensive burrowing which suggests abundant organic material incorporated in the sediment. In reflected light, small grains of pyrite can be seen disseminated within the matrix and large grains of magnetite are partially replaced by pyrite. If small magnetite grains originally deposited were completely altered to pyrite, while the larger magnetite grains were only partially altered, then the paleomagnetically stable single domain and pseudo-single domain grains would be preferentially eliminated, leaving only the larger multi-domain grains. Large multi-domain grains are inherently less stable and would be expected to track the long-term changes in the geomagnetic field.

MacColl Ridge Formation

MacColl Ridge is located along the southern flank of the Wrangell Mountains, in the south central portion of the McCarthy quadrangle (figure 5). This area has been mapped most recently by MacKevett and coworkers (MacKevett, 1978, Jones and MacKevett, 1969). All paleomagnetic localities described here are confined to the McCarthy A-4 quadrangle with the exception of WRG.14, which is included with the other Kennicott Formation localities (figure 6). The McCarthy A-4 quadrangle was mapped by Miller and MacColl (1964). MacColl Ridge is a synclinal mountain composed of open folded Cretaceous sedimentary rocks, largely of shallow marine origin. Jones and MacKevett (1969) described the stratigraphy of the area and established formations.

Paleomagnetic samples were collected at five localities from three formations on the eastern end of MacColl Ridge (figure 13). This area contains one of the most complete Cretaceous sections in the Northeast Pacific region (Jones and MacKevett, 1969), with rocks ranging from Albian to Campanion or Maestrichtian age. The lower portion of the MacColl Ridge strata was deposited during the Cretaceous long normal polarity interval (Lowrie and Alvarez, 1981). Thus, the lower section has the potential to give a VGP of known polarity, while the upper section could yield reversals for stability tests. In addition, the structure lends itself to application of the fold test.

Kennicott Formation

Due to inconsistent usage and incorrect age assignments, the Kennicott Formation was redefined by Jones and MacKevett (1969). This

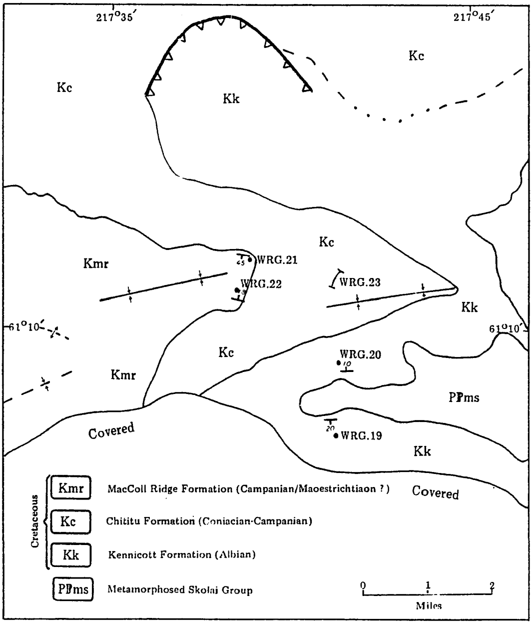


Figure 13 Generalized geological map of the MacColl Ridge area showing sample localities. Geology is modified from Miller and MacColl (1964). The V shaped formation contact pattern is due to topography; it is not due to a plunge.

unit unconformably overlies Triassic and Jurassic rocks and is the oldest Cretaceous formation in the central McCarthy quadrangle. The Kennicott Formation is entirely of Albian age (Lower Cretaceous).

Two localities of Kennicott Formation were collected in an unnamed creek valley in the central portion of the McCarthy A-4 quadrangle (figure 13). At locality WRG.19, 31 samples were collected from 12 beds of fine to very fine grained sandstone. Sampling spanned approximately 40 m of section over a distance of about 250 m. In general, the beds dip in a southerly direction at moderate to shallow angles. This area contains several high angle faults and one near vertical dike.

Locality WRG.20 lies about 2 kilometers north of WRG.19. At WRG.20, 35 cores were drilled from 12 beds of fine to very fine grained sandstone. This locality exposes an estimated 70 m of stratigraphic section over a distance exceeding 200 m. The beds dip to the north and northwest at moderately shallow dips.

Both suites of Kennicott Formation samples were subjected to thermal and AF cleaning treatment. The specimens responded poorly to thermal cleaning. The large majority of thermally cleaned specimens displayed an intensity spike in the 350 to 500 degree C range, with attendant erratic directional behavior. Most AF treated specimens failed to clean to a stable demagnetization end point. A large number of specimens did yield a well-defined secondary component magnetization (declination = 50 degrees, inclination = 81 degrees, K = 24.3 in the geographic reference frame). Only 24 of the 132 Kennicott specimens could be cleaned to a recognizable demagnetization end point.

The characteristic direction of the WRG.20 specimens (table 9) is not significantly different from the observed secondary component and is similar to the direction of the recent field. This is probably a remagnetized direction.

The WRG.19 specimens give two fairly well defined groupings. The specimens with the best demagnetization behavior gave a mean geographic direction with a declination = 349 degrees, inclination = 53 degrees, K = 14.4. This direction is virtually the same as that of the dike and suggests that remagnetization occurred at the same time as dike intrusion. The remaining specimens not remagnetized in the same direction as the dike give a fairly close cluster. The precision decreases substantially when the tilt correction is applied (table 10). This may reflect declination anomalies introduced during faulting, or alternatively the marked decrease of precisions may be interpreted simply as a failure of the "fold test". Given the uncertainties and the lack of stability tests, the Kennicott Formation localities are not considered to represent primary magnetic directions.

One additional Kennicott Formation locality (WRG.14) was sampled approximately 1 km NNE of the Nikolai Mine (abandoned) in the McCarthy B-5 quadrangle (figures 5, 6). At this locality the Kennicott Formation consists of medium to coarse grained sandstone with less abundant pebbly sandstone and fine grained sandstone. Twenty-one reconnaissance core samples were drilled from 5 to 10 m of stratigraphic section.

The WRG.14 suite of specimens was thermally demagnetized. The results were similar to WRG.19 and WRG.20 thermally cleaned specimens. Intensity increases were observed, although not as pronounced as in

Table 9

Kennicott Formation (WRG.20)

Longitude: 217.69
Latitude: 61.16

Sample	Demag Level	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat
35A	500 T	12	82	19	69	340	77
36A	450 T	259	70	298	73	151	60
55A	400 T	309	52	308	34	103	34
56A	300 T	301	67	302	49	115	41
37B	700 AF	41	71	34	58	339	60
45B	200 AF	22	59	6	50	27	60
47B	100 AF	8	54	357	44	42	54
48A	100 AF	51	73	12	68	358	78
49B	200 AF	22	61	5	53	29	62
65A	200 AF	2	59	359	42	39	53
	Mean	356	71	349	58	58	69
	K=	16.9		13.1		7.8	
	$\alpha 95$ =	12.1		13.8		18.4	
	N=	10		10		10	

T-Thermal Demagnetization in degrees C.

AF-Alternating Field Demagnetization in Oerstads.

Table 10

Kennicott Formation (WRG.19)

Longitude: 217.69
Latitude: 61.14

Remagnetized along same direction as dike.

Sample	Demag Level	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat
21A	400 T	310	27	308	27	99	30
22A	450 T	350	53	347	55	61	63
23B	450 T	8	50	6	53	26	62
21B	75 AF	345	36	343	38	62	48
23B	200 AF	355	57	352	60	55	69
22B	100 AF	4	76	357	79	209	83
25B	200 AF	17	58	53	52	323	46
15B	200 AF	(359	54)*	--	--	--	--
	Mean	349	53	351	56	52	67
	K=	14.1		10.1		7.0	
	$\alpha 95$ =	16.5		20.0		24.6	
	N=	7		7		7	

*15B is a sample of the dike and is not calculated in the mean.

Second "Cluster" of Sample Directions

Sample	Demag Level	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	VGP Long	VGP Lat
5A	350 T	145	78	162	52	233	5
6A	400 T	153	62	160	37	236	-7
7A	400 T	165	57	167	31	230	-12
6B	200 AF	191	41	186	16	211	-20
27B	100 AF	199	62	229	34	172	-1
17B	200 AF	289	47	286	46	128	32
	Mean	193	67	192	44	207	-1
	K=	8.4		4.8		3.5	
	$\alpha 95$ =	24.5		34.1		41.7	
	N=	6		6		6	

T-Thermal Demagnetization in degrees C.

AF-Alternating Fields Demagnetization in Oerstads.

the other Kennicott suites, in the 300 to 550 degree C range. No stable demagnetization end point could be isolated at high temperatures with the exception of one sample. Many of the specimens gave recent field directions at unblocking temperatures of 300 degrees C and less. Since no meaningful directions could be calculated, these data are not considered further.

Chititu Formation

The age of the Chititu Formation ranges from at least Cenomanian to late Campanian. The age of the base of the Chititu Formation varies with location and may be as young as Coniacian on MacColl Ridge (Jones and MacKevett, 1969). The age of primary magnetic remanence may therefore be taken as Coniacian to Campanian, approximately 89 m.y. to 75 m.y. The Chititu Formation was sampled over 75 meters of section along the eastern slope of peak 5470 (section 14, T8S, R17E) in the McCarthy A-4 quadrangle (figure 13). These rocks consist largely of hackly fractured mudstone and siltstone, with rare sandstone beds and clastic dikes. Due to the unsuitable drilling qualities of the fine-grained rocks and logistic considerations, only five oriented hand samples of sandstone could be collected. Each hand sample was collected from a discrete sandstone bed. Between 4 and 8 specimens were drilled from each hand sample.

The Chititu specimens were cleaned by AF or thermal/AF combination demagnetization techniques. Combination methods consisted of thermal cleaning at 100, 150 and 180 degrees C, to destroy any goethite remanence, followed by AF cleaning beginning at 25 oersteds. Specimens from samples 115 and 119 failed to show any stable magnetic direction.

Demagnetization of specimens from samples 117 and 118 yielded very well-defined directions. Coercivities of sample 117 specimens ranged from 100 to 250 oersteds and characteristic directions clustered tightly (K=24.5). Sample 118 coercivities varied from 50 to 150 oersteds. Although the direction for a given specimen was well defined for sample 118 specimens, the clustering of the various specimen directions is quite diffuse (K = 3.2). The reason for this variability of the magnetic direction recording ability of the rock was not obvious in the field; however, it could be related to a partial remobilization of the sand during clastic dike intrusion. Such a hypothesis would require the properties of samples 117 and 118 to be substantially different although their statigraphic separation is only 35 cm.

To investigate this hypothesis further, thin sections of the Chititu hand samples were made. Although both samples 117 and 118 contain largely quartz, plagioclase, chert and minor heavy minerals, there are some major differences. Sample 117 contains abundant carbonate cement (approximately 25 percent) suggesting a high initial porosity. Sample 118 contains an argillaceous matrix (5-10 percent) and a considerable amount of compaction-distorted biotite fragments (10-15 percent). The tight grouping of sample 117 remanence directions can be understood in terms of the high initial porosity allowing easy post-depositional realignment of magnetic carriers. The poor clustering of sample 118 remanence directions is most likely due to the high argillite/biotite content which would inhibit realignment of magnetic carriers by plugging up the porosity. Although Payne and Verosub (1982) found that at least 40 percent silt/clay content was necessary to prevent realignment of the magnetic carriers in their sediment studies, their conclusions were based on a limited number of samples and they did not investigate the effect of relatively malleable grains such as biotite. Thus, the relatively high matrix/mica content in sample 118 does not necessarily preclude the application of the immobilization of magnetic carriers concept.

Specimens from sample 116 gave directionally variable demagnetization curves; however, a reasonably well-defined characteristic direction could be isolated. This direction was not significantly different from the recent geomagnetic field. This suggests that sample 116 was remagnetized. However, the stratigraphic directions are generally similar to samples 117 and 118. In addition, the overlying MacColl Ridge Formation has a geographic direction similar to the recent field direction but in this case it is a demonstrably stable primary magnetization as described below. Therefore, the mean directions of samples 116, 117 and 118 are tentatively accepted as valid measures of the Late Cretaceous field.

The mean magnetic direction of the three bed means gives a westerly declination and an intermediate to shallow inclination. Since there are only three mean directions,the error limits are unusually large and any paleolatitude in the northern hemisphere would be permissible. Given the fact that the Chititu Formation has numerous sandstone dikes, it is reasonable to assume that the magnetic carriers were mobile for some time after deposition: that is, the remanence is probably a PDRM (post-depositional remanent magnetization) acquired after deposition, during dewatering and compaction. Under this assumption, it is then reasonable to suspect that remanence in a given bed may have been acquired at slightly different times, due to small scale inhomogenieties of texture and sorting. If this were the case, then the mean of the individual specimen directions would give the best estimate of the geomagnetic field direction. This mean of all specimens is virtually the same as the mean of bed means (table 11); the precision (K) is reduced slightly but the error limits are reduced dramatically (due to larger N). The mean direction is geologically reasonable when compared with the overlying MacColl Ridge Formation, although given the uncertainties and insufficient sample coverage, this mean can be used as little more than a qualitative argument. It does, however, indicate that recollecting from the Chititu Formation to obtain a statistically well-constrained remanence direction is worthwhile.

MacColl Ridge Formation

The MacColl Ridge Formation consists of coarse grained shallow marine to nonmarine sandstone and conglomerate with less abundant interbedded siltstone. The MacColl Ridge Formation overlies the Chititu Formation with apparent conformity (Jones and MacKevett, 1969). The precise age determination of the MacColl Ridge Formation is hampered by the lack of well-preserved fossils; only poorly preserved plants, Inoceramus fragments and unidentifiable ammonite fragments were found. Jones and MacKevett (1969) considered the age to be late Campanian or Maestrichtian (latest Cretaceous), on the basis that it overlies Campanian rocks (Chititu Formation) and is intruded by

Table 11

Chititu Formation (WRG.23)

Longitude: 217.68
Latitude: 61.18

Specimen	Demag Level*	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc
116A	150	18	69	282	80
116B	75	347	68	282	69
116C	75	343	53	307	59
116E	100	34	74	237	82
117A	200	272	1	275	-7
117B	150	290	32	268	28
117C	150	276	21	265	11
117D	250	289	43	258	35
117E	100	277	19	267	10
117F	150	283	35	260	26
118A	75	302	55	261	45
118B	75	8	-29	11	4
118C	50	265	-3	271	-20
118D	75	278	-4	283	-14
118E	150	286	51	256	35
118G	100	320	57	266	55
118H	150	320	44	284	47
	Mean	299	40	275	35
	K=	4.3		4.6	
	$\alpha95$ =	19.4		18.8	
	N=	17		17	

*Thermal Demagnetization (to 180° C) followed by Alternating Field Demagnetization.

Tertiary plutons. Since the age of the plutons has not been accurately determined and as no fossils have been recovered from the upper part of the MacColl Ridge Formation (Jones, 1982, personal communication), the possibility that at least part of the MacColl Ridge Formation is Tertiary in age cannot be ruled out. However, since the paleomagnetic samples were collected near the base of the MacColl Ridge Formation, a Late Cretaceous age for the primary remanence is reasonable.

Paleomagnetic samples were collected from two limbs of the MacColl Ridge syncline (figure 13). Thirty-one samples from fifteen beds were collected from a locality (WRG.21) in the northern limb of the syncline. From a complementary locality (WRG.22) in the southern limb of the fold, eighteen samples from eleven beds were collected. Both localities occupy approximately the same stratigraphic position.

At locality WRG.21, the 15 beds were sampled over approximately 13 m of section. All samples were drilled from moderately well-sorted medium and fine grained sandstone and less commonly from coarse grained sandstone. In general, the sandstone beds are 5 cm to 20 cm thick and infrequently as thick as 1 to 2 m. The sandstones tend to thicken and coarsen up section. All sandstone beds are separated by at least one and usually several siltstone or shale interbeds. Some of the sandstone beds have a distinctly lenticular shape and can be observed to pinch out over distances of about 10 m.

A total of 71 specimens from WRG.21 were subjected to thermal, AF and thermal/AF combination techniques. The 30 thermally demagnetized specimens behaved erratically and no stable remanence direction could be isolated. The demagnetization curves of most of these specimens displayed an intensity spike, occuring at 250 to 350 degrees C, which is suggestive of a magnetic mineralogic change or a high temperature VRM (viscous remanent magnetization).

Many of the AF and thermal/AF cleaned specimens displayed unstable behavior (sometimes due to RRM) and no characteristic direction could be resolved. Eleven specimens from seven beds did, however, show stable demagnetization end points (table 12), with coercivities of 100 to 200 oersteds. These directions clustered tightly and contained one reversed direction. It is interesting to note that eight additional specimens gave essentially the same direction as the eleven stable specimens. Nevertheless, due to the lack of intermediate points on Zijderveld plots showing an intensity loss with little directional change and/or the presence of strong RRM overprinting, these questionable data were not included in the calculation of the WRG.21 mean direction.

Due to the poor results of the thermally cleaned WRG.21 specimens, all WRG.22 specimens were cleaned using AF and thermal/AF techniques. Similar problems with unstable or RRM obscured directions were encountered. Eleven of the 36 specimens from WRG.22 yielded stable demagnetization end point directions. Two of the eleven characteristic directions isolated were rejected because they differed greatly in direction from replicate samples from the same bed, were similar in directions to some secondary components removed from other specimens at the locality and they failed the theta-95 criterion. The remaining 9 specimens from 7 beds gave a relatively tight cluster and had 3 reversals.

Table 12

MacColl Ridge Formation (WRG.21)

Longitude: 217.65
Latitude: 61.18

Bed	Demag Range*	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K	VGP Long	VGP Lat	K
1	100-150	2	343	76	12	241	49	11	169	14	9
4	200	3	357	84	10	216	51	10	185	12	6
5	200	1	95	-78R	--	45	-40R	--	357	-1R	--
6	100-200	2	22	68	268	221	68	243	191	27	130
7	100	1	341	73	--	239	54	--	170	18	--
9	100	1	48	87	--	203	52	--	198	6	--
12	200	1	128	82	--	201	43	--	199	-2	--
	Mean		355	82		221	52		183	11	
	K=		49.5			36.3			22.8		
	$\alpha95$ =		8.7			10.2			12.9		
	N=		7/11			7/11			7/11		

MacColl Ridge Formation (WRG.22)

Longitude: 217.64
Latitude: 61.18

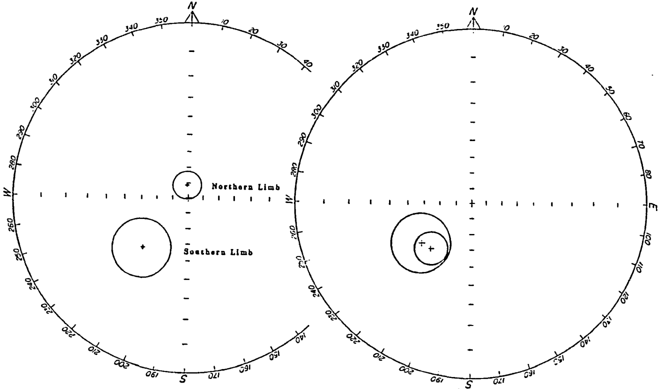
Bed	Demag Range*	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K	VGP Long	VGP Lat	K
1	200	1	75	-32R	--	80	-31R	--	324	-10R	--
4	200	1	221	43	--	230	47	--	175	8	--
6	100-200	2	220	49	6	229	53	7	177	17	6
7	200-600	2	243	68	439	259	67	375	164	37	194
8	300	1	49	-32R	--	53	-33R	--	348	0R	--
9	300	1	181	64	--	191	70	--	210	25	--
11	200	1	17	-32R	--	21	-36R	--	18	7R	--
	Mean		222	48		231	50		177	14	
	K=		13.1			13.2			10.1		
	$\alpha95$ =		17.3			17.3			20.0		
	N=		7/9			7/9			7/9		

*Alternating Field and Thermal/Alternating Field Demagnetization.
R-Reversal whose direction is inverted for calculation of the mean.

Due to difficult drilling conditions at the WRG.22 locality, some of the sampling repeated stratigraphic section. Beds 1 through 6 were sampled in descending order over approximately 10 m section along a talus slope adjoining a near-vertical cliff face. Beds 7 through 11 were sampled in ascending order in a steep couloir (about 9 m of section) repeating the previous stratigraphic interval yet no closer than 10 m to the nearest sample taken from the first 6 beds sampled. Each of the 11 beds sampled was given unit weight as an instantaneous reading of the geomagnetic field. Although some laterally equivalent beds were resampled, the unit weighting is justified by the lenticular nature of the beds and the small probability that a sample of the exact time interval of a previous sample was duplicated. Bed 1 is the lateral equivalent of bed 11. Both beds show a polarity reversal, suggesting that these rocks have stably recorded the geomagnetic field from the time of deposition.

The mean geographic vector for WRG.21 is not significantly different from the recent field direction. However, since the mean vector of WRG.21 aligns very closely with the mean direction of WRG.22 when both are corrected for tilt (passing the fold test, figure 14), the correspondence of WRG.21 with the recent field direction is almost certainly fortuitous. The two MacColl Ridge Formation localities contain reversals and pass the fold test, thus substantiating the primary nature of the remanence direction.

Since there was a high rejection rate of MacColl Ridge Formation specimens (based on observed magnetic instability), a qualitative rock magnetic study was undertaken in order to help explain the poor magnet-



14a 14b

Figure 14 Positive fold test for MacColl Ridge Formation directions. a) Geographic vector directions showing the statistical distinctness of the two limbs of the fold. b) Stratigraphic vector directions show the coincidence of the two limb directions when the tilt correction is applied. This proves that the magnetization of the rocks was aquired before folding.

ic recording qualities of many specimens. The study was conducted on several specimens that displayed the distinctive intensity "spike" during thermal demagnetization and their AF demagnetized replicate specimens. The specimens were cut into small pieces and then crushed using a steel mortar and pestle. The magnetic minerals were separated from the nonmagnetics as the rock powder was washed through a settling tube in a strong magnetic field. The magnetic separate was then ground a second time using an agate mortar and pestle and subjected to ultrasonic cleaning to remove nonmagnetic minerals that might have adhered to the surface of the magnetic grains. Due to the very low abundance of magnetic grains in these rocks, the rock powders were combined into composite AF demagnetized and thermally demagnetized samples, in order to recover manageable quantities of magnetic mineral separates.

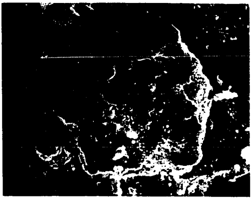
X-ray analysis of the magnetic minerals was attempted using a Rigaku Miniflex X-ray diffractometer and CuKa radiation; however, no identifiable peaks were seen on the diffractogram due to insufficient quantity of mineral specimen and possible absorption effects associated with Cu X-rays in a Fe-bearing sample.

The magnetic separates were next examined using a Jeol JSM-35U scanning electron microscope (SEM) equipped with a Kevex 7000 energy dispersive X-ray analyzer. Magnetic minerals were examined at 1000x to 20,000x magnification. Analysis of the magnetic grains was hampered by ubiquitous small particles mantling larger grains. X-ray analysis of many of these particles reveal compositions which include the elements Si, Al, K, Ca. The superficial particles are probably feldspars, quartz and possibly calcite. One region contained the elements Si, P and Ca; quartz and apatite are a possible combination.

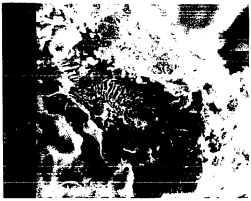
Despite the silicate particle mantle on most grains some interesting features could be identified. Photo 5336 (figure 15a) shows a large Fe-rich grain displaying striations with numerous feldspar(?) cover particles. The presence of striations and the fact that only Fe registered in the X-ray analysis suggests that the grain is magnetite or a Ti-poor titanomagnetite. Photos 5334 and 5335 (higher magnification view of photo 5334, figure 15b, 15 c) show a ribbed pattern. The valleys are largely composed of Fe, while the ridges contain minor Ti in addition to Fe. The ridges on this grain are interpreted as exsolution lamellae of a Ti-rich phase from a Ti-poor phase, probably a titanomagnetite. Other grains without the exsolution features show both Ti and Fe in the X-ray analysis - again, probably titanomagnetites. Considering the SEM evidence, the MacColl Ridge sandstones contain at least two general types of magnetic carriers, probably magnetites and titanomagnetites.

The experiments of Hartstra (1982) shed some light on the demagnetization problems encountered with the MacColl Ridge samples. Hartstra (1982) carried out a detailed magnetic mineralogy study on naturally occurring titanomagnetites ($x=0.35$, i.e. 35 percent ulvospinel component). In these experiments, magnetic intensity peaks were observed during continuous thermal demagnetization. These intensity peaks were shown to be related to descreening of low Ti-titanomagnetite inclusions in the early stages of exsolution. During successive con-continuous demagnetizations, the intensity peak was obeserved to split

15A



15B



15C

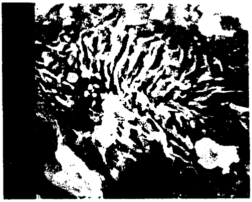


Figure 15 Scanning electron photomicrographs of magnetic minerals from the MacColl Ridge Formation. a) Probable magnetite grain. Note the striations on the grain surface. b) Probable titano-magnetite grain with well developed exsolution lamellae. c) Higher magnification view of previous photo. Light areas have higher Ti contents than the darker areas.

into two peaks. Hartstra (1982) attributed this to progressive exsolution to a low Curie temperature high Ti phase and a high Curie temperature low Ti phase. Stepwise-continuous thermal demagnetization of two artificially induced TRMs at 90 degrees to each other gave magnetic directions intermediate to the two TRM directions even at temperatures exceeding the Curie temperature of the high Ti phase. This was ascribed to a low temperature chemical alteration yielding a stable (to high temperatures) CRM parallel to the low temperature TRM (Hartstra, 1982).

The poor response of the MacColl Ridge Formation specimens to thermal demagnetization might be accounted for by processes similar to those observed by Hartstra (1982). The intensity spikes observed during thermal cleaning could be ascribed to an exsolution CRM generated during thermal cycling. Oxidation of a titanomagnetite followed by exsolution to Ti-rich and Ti-poor phases is another possible cause of the intensity anomalies.

The poor behavior of some MacColl Ridge specimens to AF treatment might also be explained by exsolved magnetic carriers. Although Hartstra (1982) was able to resolve both high and low temperature TRMs by AF cleaning (due to different coercivities of the two magnetic phases), exsolved phases with different remanence directions and overlapping coercivity spectra can produce erratic demagnetization behavior. The "well-behaved" specimens could then be due to local variations in magnetic mineral content which isolated relatively abundant analtered magnetite and single phase titanomagnetite.

Bonanza Creek Localities

The Gravina-Nutzotin belt consists of a widespread sequence of Jurassic and Cretacous flysch and andesitic volcanics (Berg and others, 1972). Since this belt depositionally overlies both the Wrangellia and Alexander terranes, paleomagnetic data from these rocks pertain to the paleogeography of the composite superterrane. A paleomagnetic investigation was undertaken on rocks collected from Bonanza Creek, in the central Nabesna A-2 quadrangle, about 43 km east of Chisana (figure 4). Bonanza Creek was selected for study because the rocks of this area form a remarkably continuous section of Gravina-Nutzotin belt rocks spanning Late Jurassic through Early Cretaceous time (Richter and Jones, 1973). As such, these rocks have the potential to fill in a marked paleomagnetic data gap. Moreover, since the units appear to lie within one structural block, a polar wander path segment could be constructed without the complicating effects of local tectonic rotations.

Sample Locations

A total of 207 cores were drilled over a 4 km length of Bonanza Creek, at 4 general localities. Locality BNZ.3, approximately 2 km ENE of the juncture of Bonanza Creek and Coarse Money Creek, is probably the oldest of the four localities. The nearest fossil localities (1.4 km to the east and 1.7 km to the west) are Tithonian (latest Jurassic) in age (Richter and Jones, 1973). The lithologies consist of subequal amounts of argillite and interbedded turbidite sandstones. Turbidite beds average 10 to 20 cm in thickness and occasionally are as much as 80 cm thick. At BNZ.3, 65 cores were drilled from 17 sandstone beds over 20 m of stratigraphic thickness. At least 3 samples were drilled from each bed and 4 beds had at least 6 samples. This hybrid sampling scheme was used in order to compare the mean directions obtained by various sampling techniques.

In addition, 4 cores from 2 dikes and a sequence of cores trending away from the intrusive contact were collected for a baked contact test. The age of these dikes is not known. If they are assumed to be feeder dikes for the Chisana Volcanics, they need be only slightly younger than the flysch. Thus, observing the same directions in the dikes and country rock does not necessarily imply area wide thermal overprinting.

Locality BNZ.1 is located at the junction of Coarse Money Creek and Bonanza Creek. The rocks at this locality represent the uppermost portion of the Nutzotin Mountain sequence. BNZ.1 corresponds to fossil localities 8, 9 and 10 of Richter and Jones (1973). A late Berriasian or Valanginian age (Early Cretaceous) is indicated.

At locality BNZ.1, 37 cores were acquired from 12 turbidite beds along a 300 m segment of Bonanza Creek. A minimum of 50 m of section is spanned by the collection. Three dikes (7 cores) and sequential samples across the intrusive contacts were sampled in order to apply the baked contact test.

The sampling at BNZ.1 extended into the lowermost part of the Chisana Formation. Bedding could not be seen at this outcrop of Chisana Formation. Since bedding had to be extrapolated from the nearest upstream indicators in the flysch, the Chisana stratigraphic directions could be in error.

The third suite of paleomagnetic samples from this area was collected about 1.6 km downstream from the previous flysch outcrop. This sampling area, BNZ.2, is located at the confluence of Little Eldorado Creek and Bonznza Creek and is stratigraphically higher than the flysch. The sampled lithologies are volcaniclastic sediments and minor flows within the lower Chisana Formation. The fossils reported from this area (Richer and Jones, 1973) are apparently not age diagnostic at the stage level. Since these rocks overlie Valanginian rocks and Barremian fossils were recovered high in the formation (Berg and others, 1972), an Early Cretaceous age for the Chisana Formation may be inferred.

Sixty cores were collected from 14 beds or flows. The sampled section covers a 30 to 50 m stratigraphic interval over nearly 200 m of ground distance. Additionally, 2 samples were drilled from one dike for a baked contact test.

The last paleomagnetic sampling locality and the stratigraphically highest is BNZ.4 in the upper Chisana Formation. This suite of cores was collected approximately 1.8 km downstream from BNZ.2 on Bonzanza Creek near the junction with Canyon Creek. Thirty-two samples were collected from 11 beds or flows. Approximately 33 m of section was represented by the samples. Although BNZ.4 is the highest sample locality in the section, the basic age constraints of BNZ.2 apply; thus it may also be assigned an Early Cretaceous age.

Paleomagnetic Data

All suites of specimens were demagnetized by thermal techniques. AF cleaning was performed on replicate suites of BNZ.1, 2, 3. The results of AF and thermal cleaning are virtually the same.

From the BNZ.1 samples, 44 specimens were thermally demagnetized. In general, these specimens responded well to thermal treatment (7 specimens did not reach a stable end point). Characteristic directions were obtained at unblocking temperatures ranging from 400 to 575 degrees C, with 500 and 550 degrees being the most common value. At temperatures above the characteristic unblocking termperatures, no additional directional groupings could be detected.

Thirty-four replicate specimens were cleaned by AF means. Remarkably stable directions were obtained by AF cleaning (figure 16). Vectors decayed in stepwise fashion towards the origin with very little directional variation. Even at 300 or 500 oersteds cleaning intensity, only 3 specimens showed signs of minor RRM. The optimum demagnetization force was 400 oersteds for most specimens and little directional change was noticed at the 600 or 700 oersted step. As with the thermally treated specimens, the AF group has no secondary magnetic component. The mean directions of the two demagnetization groups are much the same (table 13).

Two specimens from the Chisana Formation at BNZ.1 gave a significantly different magnetic direction. This could be explained by a remagnetization of either the flysch or the volcanics, a marked time discontinuity separating the flysch and volcanics or an unrecognized structural disruption between the two units.

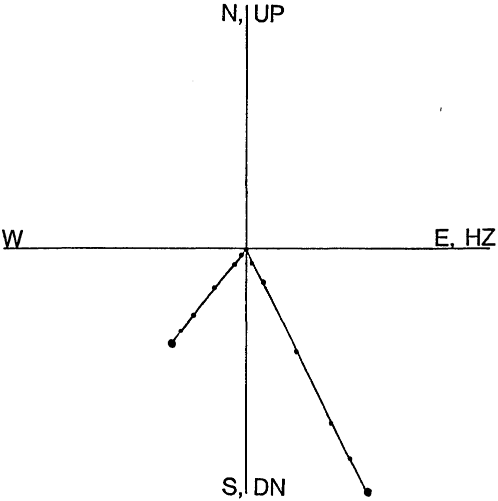


Figure 16 Vector diagram for specimen 872B. This type of behavior is typical of specimens from BNZ.1, Nutzotin Mountains Sequence. Decay of remanence intensity on a straight line trajectory towards the origin indicates that the characteristic direction has been isolated.

Table 13

Nutzotin Mountain Sequence (BNZ.1)

Thermal Data

Longitude: 218.2
Latitude: 62.1

Bed	Demag Range	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K
1	450-575	3	16	55	50	236	71	51
4	550-575	2	355	25	43	314	63	45
dike	550	2	15	54	896	--	--	--
5	500-575	6	351	63	22	230	63	22
6	500	4	8	66	53	213	65	54
7	450-550	2	46	67	2.6*	183	58	2.6*
8	400	1	13	72	--	207	56	--
9	500-550	4	12	60	45	209	80	46
10	500-550	2	4	63	71	219	76	63
11	500	3	8	67	148	208	73	142
dike	550-575	3	3	65	240	--	--	--
	Mean		5	59		228	72	
	K=		28.6			29.2		
	$\alpha95$ =		9.3			9.2		
	N=		8			8		

Alternating Field Data

Bed	Demag Range	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K
1	400-500	3	355	71	31.4	229	53	31.7
2	600	1	39	74	--	232	65	--
3	400-600	3	356	67	42.2	216	49	43.9
4	400-700	3	353	53	155.2	245	65	160.9
dike	200-400	3	23	58	137.3	--	--	--
5	400-600	6	352	69	38.2	222	59	38.6
6	200-400	4	1	67	148.8	218	63	161.3
7	600	1	16	74	--	205	54	--
8	400-700	2	11	66	312.8	212	62	428.5
9	400	2	355	64	127.6	226	72	136.1
11	400-600	3	3	66	250.0	216	72	237.6
dike	150-300	3	17	74	194.8	--	--	--
	Mean		2	68		221	62	
	K=		107.4			73.9		
	$\alpha95$ =		4.7			5.7		
	N=		10			10		

*Not included in mean, statistically random.

Sixty-two specimens from BNZ.2 were subjected to thermal demagnetization. This suite cleaned similarly to the BNZ.1 thermal specimens, with eleven failing to show a characteristic direction. Characteristic unblocking temperatures of 400 degrees C were by far the most common and only a few specimens retained the characteristic magnetic direction at temperatures of 500 degrees C. At higher temperatures, the specimens gave apparently random directions. AF cleaning treatment was applied to 15 duplicate specimens. Characteristic directions were revealed at coercive forces of 200 to 1000 oersteds. RRM was encountered in some specimens at 400 oersteds. In most cases the RRM is negligible and in a few cores it was undetectable at 800 or 1000 oersteds. The directional results of AF and thermal demagnetization are nearly identical (table 14).

Thermal cleaning was carried out on specimens from each of the 69 BNZ.3 core samples. Characteristic directions were isloated at unblocking temperatures of 400 to 500 degrees C. Some erratic behavior was seen at higher temperatures. The magnetic directions from this suite were extraordinarily consistent both within beds and between beds. The one notable exception is specimen 1001 B. This specimen yielded a well defined direction of magnetization (declination = 51, inclination = -44 in the geographic frame and declination = 44, inclination = -4 in the stratigraphic frame) at a temperature of 500 degrees C. Additional samples from the same bed are significantly different than 1001B but agree with each other and with directions of the remaining specimens in the suite. Specimen 1001B thus fails an internal consistency check and is considered an aberrant datum.

Table 14

Chisana Formation (BNZ.2)

Longitude: 218.2
Latitude: 62.1

Bed	Demag Range*	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K
1	400	4	1	65	660.1	275	67	674.1
2	400	2	6	69	60.6	263	68	60.5
3	400	3	350	69	175.5	266	62	171.7
4	400-500	3	2	63	112.3	268	69	116.2
5	400	2	20	66	1095.5	177	83	885.9
6	400	2	11	59	145.9	330	77	154.7
7	400	6	13	58	279.5	336	75	264.9
8	400-500	4	345	58	21.8	301	65	21.5
9	400-500	5	5	67	484.7	293	77	510.2
dike	400-500	2	32	56	180.2	22	78	164.1
8/9†	400	3	29	61	28.6	0	83	29.9
10	400	3	10	72	257.0	283	74	247.5
11	400-500	2	26	53	12.3	356	72	13.5
12	400-500	4	17	53	239.7	342	69	273.9
13	400-500	2	338	66	1431.9	289	64	1493.1
14	400	4	349	72	132.7	276	71	131.8
	Mean		7	64		294	75	
	K=		78.7			42.7		
	$\alpha 95$ =		4.1			5.5		
	N=		15			15		

*Thermal Demagnetization in degrees C.

†Lateral equivalent of either bed 8 or 9.

Ten replicate specimens, one specimen per bed, were selected for AF treatment. All specimens cleaned exceptionally well. The optimum demagnetization level was 400 oersteds and only minor RRM was observed at 700 oersteds. These data (table 15) cluster very tightly (k = 141.6) and are considered representative of the thermally demagnetized specimens form BNZ.3

Sixty specimens from BNZ.4, in the Chisana Volcanics, were thermally cleaned. All specimens displayed a very well defined and very stable characteristic direction. Unblocking temperatures ranged from 500 to 650 degrees C. Numerous specimens, stable at 600 to 650 degrees C, suggest the presence of hematite, since the magnetite Curie temperature is 578 degrees C. The abundance of hematite is confirmed by thin section examination and by the distinctive red drilling mud produced during core collection.

In thin section, the hematite appears to be replacing magnetite. Moreover, considerable hematite is seen within amygdules in the volcanic rocks (figure 17). This strongly suggests a secondary origin for the hematite and casts serious doubt on the primary nature of the characteristic direction (table 16).

When considering all of the Bonanza Creek localities as a whole, there is little chance that the paleomagnetic directions are primary. Magnetic evidence shows that hematite is the major remanence carrier in the BNZ.4 suite and petrographic evidence suggests a secondary origin for the hematite. The remaining localities show geographic magnetic directions similar to BNZ.4, suggesting that BNZ.1-3 are also recording a secondary diretion.

Table 15

Nutzotin Mountain Sequence (BNZ.3)

Longitude: 218.2
Latitude: 62.1

Bed	Demag Level*	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc
dike	400	10	42	315	72
2	400	355	49	287	65
3	400	351	47	289	62
5	400	355	47	279	68
6	400	21	46	279	87
8	400	5	48	289	75
9	400	11	40	324	75
10	300	2	44	300	72
11	400	4	42	311	67
12	400	7	50	291	71
17	400	1	48	312	64
	Mean	3	46	297	71
	K=	141.6		87.8	
	α_{95} =	4.1		5.2	
	N=	10		10	

*Alternating Field Demagnetization in Oersteds.

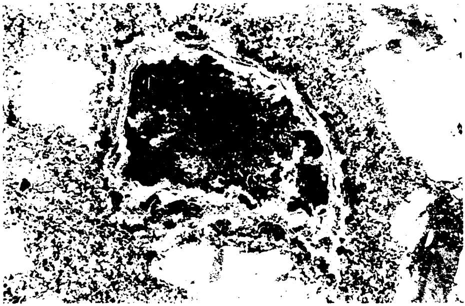


Figure 17 Photomicrograph of Chisana Volcanics. Note the hematite lining the vesicle. This hematite is thought to be secondary in origin and is the probable carrier of the remagnetized remanence directions observed in suite BNZ.4.

Table 16

Chisana Volcanics (BNZ.4)

Longitude: 218.2
Latitude: 62.1

Bed/ Flow	Demag Range*	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K
1	500	2	354	47	475.8	153	89	611.3
2	550	2	356	62	553.7	168	74	602.2
3	500-600	2	12	58	27.8	136	74	28.8
4	575-625	2	14	55	192.6	126	73	190.8
5	550	1	4	52	--	140	79	--
6	550-575	3	357	61	131.4	169	72	128.3
7	575-600	2	0	46	17.6	116	85	17.5
8	625	1	2	32	--	26	77	--
9	600-650	4	357	38	139.7	21	85	143.1
10	650	2	353	45	295.6	192	88	279.0
11	575-625	4	15	36	387.9	67	72	332.1
	Mean		2	48		124	83	
	K=		49.7			55.2		
	α95 =		6.0			5.7		
	N=		11			11		

*Thermal Demagnetization in degrees C.

The dispersion of the BNZ.3 data is remarkably low (k = 141.6). This suggests that either secular variation has not been averaged out, due to poor temporal coverage or that remagnetiztion has occurred. Since BNZ.3 was sampled from 17 turbidite sandstone beds over 20 m of stratigraphic section with abundant argillite interbeds, it is very unlikely that there is insufficient time to average secular variation. Thus, BNZ.3 is probably remagnetized.

The four Bonanza Creek localities fail the fold test in a qualitative way. The locality mean directions become more dispersed after the tilt correction is applied (figure 18); the k value for the mean of locality means reduces from 86.0 to 33.5. Unfortunately, a statistically rigorous fold test was precluded by unsuitable structural relationships.

Dikes intruding localities BNZ.1, 2, 3 have geographic vector directions similar to their locality means. However, as the age of the dikes is unknown, the baked contact test is indeterminate.

Only normal polarities were observed in the Bonanza Creek localities. These localities sampled an aggregate stratigraphic thickness of approximately 150 m, spread out over 4 km and spans perhaps 20 m.y. It is very unlikely that reversals would be missed by the sampling scheme. Therefore, remagnetization is strongly indicated.

The evidence clearly favors the acquisition of a characteristic direction that does not date back to the origin of the rocks. When this is the case, the age of the magnetizaion is no longer constrained by the age of the rocks and the ancient horizontal (at the time of magnetization) may no longer be reflected by bedding. Without knowing the

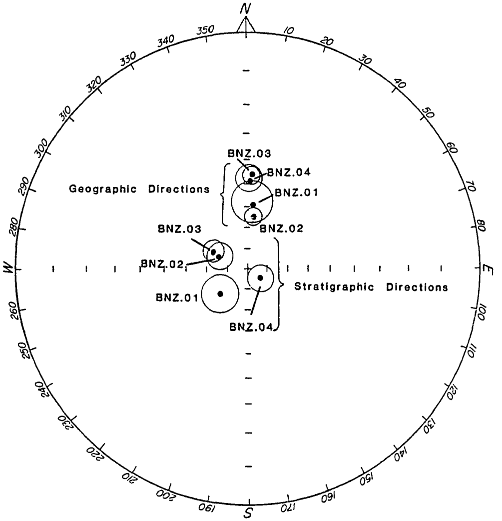


Figure 18 Mean geographic and stratigraphic vector directions for the Bonanza Creek localities. Note that the stratigraphic vector directions (corrected for tectonic tilt) are more scattered that the geographic vector directions, suggesting failure of the fold test. Insufficient variation of bedding attitudes precludes the application of a statistically rigorous fold test.

age of magnetization and the appropriate tilt correction, paleomagnetic data are of limited value in tectonic reconstructions.

Seymour Canal Formation

The Seymour Canal Formation was sampled near Grunt Point, east of Gambier Bay, on Admiralty Island. This locality (GNB.06) is situated approximately 90 km NW of Petersburg in the northern portion of the Sumdum B-6 quadrangle, southeast Alaska (figure 19). The Seymour Canal Formation, a thick unit of argillite, graywacke and conglomerate, was mapped and named by Loney (1964). This flysch-like unit is considered part of the Gravina-Nutzotin belt by Berg and others (1972). The age is Late Jurassic (Kimmeridgian to Portlandian) to Early Cretaceous (Berriasian) based on fossil evidence.

The Seymour Canal Formation was sampled within a plunging, nearly isoclinally folded anticline. Twenty-nine cores were collected over approximately 20 m of section from both limbs of the fold. The entire anticline is very well exposed for some 100 m across the axial trace, affording an excellent single outcrop scale fold test.

Thermal demagnetization was performed on 28 specimens. Six specimens failed to yield a directionally stable end point. Many of these behaved erratically and exhibited intensity jumps at 350 to 400 degrees C, with one specimen showing the spike at 500 degrees C. The remaining specimens displayed moderately to well defined demagnetization end points at unblocking temperatures ranging from 350 to 475 degrees C, with 400 to 450 degrees C being the most common (table 17).

Twenty-seven replicate specimens were cleaned by AF treatment. Good demagnetization end points were observed in 19 specimens at coercive forces of 200 to 600 oersteds. Eight specimens responded very

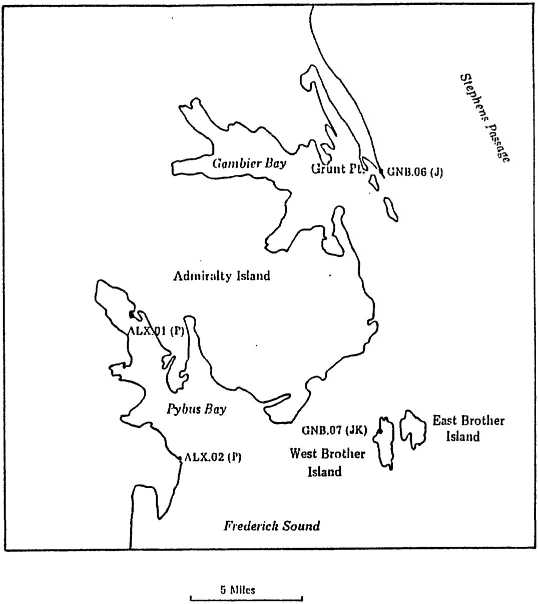


Figure 19 Map showing sample localities from southeast Admiralty Island, southeast Alaska. ALX.01 and ALX.02 - Permian, Pybus Formation. GNB.06 - Jurassic, Seymour Canal Formation. GNB.07 - Jurassic/Cretaceous, Brothers Volcanics.

Table 17

Seymour Canal Formation (GNB.6)

Longitude: 226.12
Latitude: 57.46

Sample	Demag Level	Geographic Dec	Geographic Inc	Stratigraphic Dec	Stratigraphic Inc	
128B	475 T	349	57	57	51	
129A	450 T	345	51	46	52	
135A	450 T	19	34	258	63	
137A	450 T	6	81	221	16	
138A	400 T	319	59	247	13	
139A	450 T	354	60	238	29	
140A	450 T	19	67	224	30	
141A	350 T	42	79	179	30	
142A	400 T	229	68	201	5	
143A	350 T	237	78	195	14	
144A	400 T	8	60	184	51	
145A	400 T	358	67	190	43	
146A	450 T	342	68	40	-2	"R"
147A	475 T	13	69	45	-12	"R"
148A	450 T	16	57	37	-21	"R"
149A	450 T	354	64	45	1	"R"
150A	450 T	29	69	56	-8	"R"
151A	400 T	15	72	55	-3	"R"
152A	400 T	18	55	23	-21	"R"
132B	600 AF	343	71	236	24	
134B	200 AF	110	79	208	17	
136A	400 AF	321	77	230	11	
	Mean	359	70	218	28	
	K=	20.5		5.0		
	α95 =	7.0		15.4		
	N=	22		22		

T-Thermal Demagnetization in degrees C.

AF-Alternating Fields Demagnetization in Oersteds.

R-Apparent reversal inverted for calculation of the mean.

poorly to AF demagnetization, behaving erratically in direction and frequently in intensity. The AF data is in close agreement with their thermally cleaned replicates.

The paleomagnetic directions of 22 specimens (19 thermal specimens and 3 AF specimens from cores which yielded no reliable thermal data) are plotted in both geographic and stratigraphic reference frames (figures 20 and 21 respectively). There are two somewhat diffuse but nearly antiparallel clusters in the stratigraphic reference frame. These apparent reversals seem to indicate a stable primary magnetization. However, the reduction of the k-value from 20.5 to 5.0, after correction for bedding and inversion of the reversals, suggests a negative fold test. Moreover, if the plunge is removed before correcting the bedding tilt the k-value ($k = 3.2$) is reduced still further. Thus, the data lead to the paradoxical situation of a positive reversal test yet a negative fold test.

The reversed directions occur only in one limb of the fold, while the other limb contains only normal directions. Since it is exceedingly unlikely that only one polarity would be observed in a given limb if both polarities occurred, the "reversals" may be dismissed as a consequence of fold geometry. The unfolding of a homogeneously magnetized isoclinal fold necessarily requires that the directions from one limb will be antiparallel to the direction of the other limb (figure 22). Thus, the reversed directions are not true reversals and locality GNB.06 is clearly remagnetized.

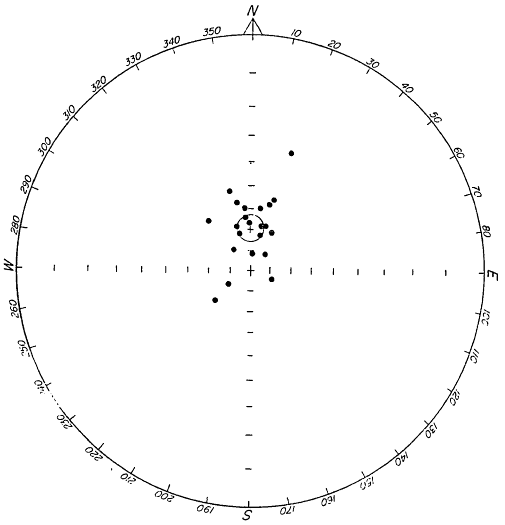


Figure 20 Geographic vector directions for GNB.06, Seymour Canal Formation.

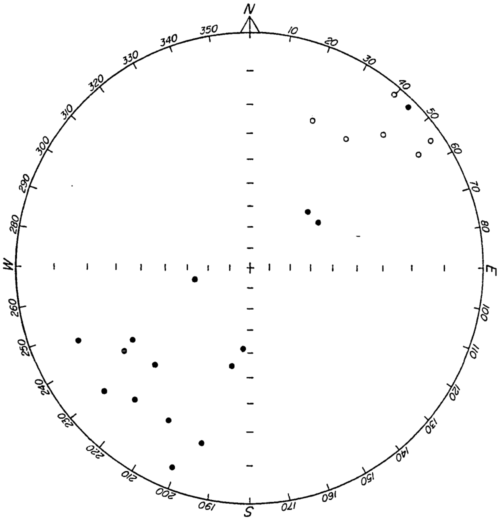
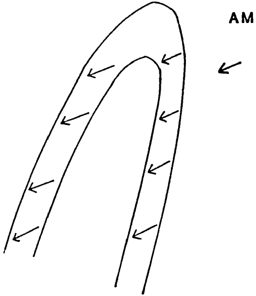


Figure 21 Stratigraphic vector directions for GNB.06, Seymour Canal Formation. Note the nearly antiparallel clusters giving apparent reversals (see text).

AMBIENT FIELD



REMAGNETIZED DIRECTIONS

IN AN ISOCLINAL FOLD
22a

APPARENT REVERSALS
AFTER UNFOLDING
22b

Figure 22 Cartoon depicting the remagnetization of the isoclinally folded Seymour Canal Formation. a) The isoclinal fold is remagnetized uniformly and parallel to the ambient field. b) After unfolding, one limb of the fold gives normal directions while the other limb records antiparallel directions or apparent reversals.

Brothers Formation

The Brothers Formation was sampled in the Sumdom B-6 quadrangle approximately 75 km NW of Petersburg, in southeast Alaska (figure 19). The Brothers Volcanics were mapped and named by Loney (1964) and is included in the Gravina-Nutzotin belt (Berg and others, 1972). This formation consists of andesitic flows, flow breccia and waterlain tuff.

Although the Brothers Volcanics are unfossiliferous and reliable radiometric age determinations are not yet available, Loney (1964) assigned a Late Jurassic and Early Cretaceous age. This age call is based on the apparent conformity between the Brothers Formation and the underlying Seymour Canal Formation (Late Jurassic) and the intensity of deformation which contrasts markedly with the gently deformed Tertiary rocks.

A sample of hornblende andesite was collected at the paleomagnetic sample locality for K-Ar age determination. Analysis was performed on a hornblende-pyroxene separate (a pure hornblende separate could not be obtained). The calculated age of 76.3 $\pm$ 2.6 m.y. (Late Cretaceous, table 18), will be reported as a minimum age due to slight alteration of the hornblende (Blum, 1984, oral communication). Although the argon retention qualities of pyroxene are not well known, it is unlikely that the pyroxene incorporated excess radiogenic argon at the time of crystallization (Dalrymple and Lanphere, 1969). Thus the minimum age estimate is justified. A Late Jurassic to latest Early Cretaceous (Albian) age is used, since the Brothers Volcanics overlie the Late Jurassic Seymour Canal Formation (Loney, 1964) and no units correlative

Table 18

K-Ar Radiometric Age Data
Brothers Volcanics

Field No. (Lab.No.)	Rock Type	Mineral dated	K_2O(wt.%)	Sample Weight (g)	$^{40}Ar_{rad}$ (mol/g) x 10^{-11}	$^{40}Ar_{rad}/^{40}K$ x 10^{-3}	$^{40}Ar_{rad}/^{40}Ar_{total}$	Age $\pm$ 1 σ (m.y.)
82JD110G (83090)	Lithic Tuff	Amphibole and Pyroxene	0.200 0.200 0.210 0.213 $\bar{x}$ = 0.206	0.6168	2.31	4.53	.325	76.3 $\pm$ 2.6* minimum age

Notes: rad = radiogenic; σ = standard deviation; $\bar{x}$ = mean; $\lambda_e + \lambda_e = 0.581 \times 10^{-10}\ yr^{-1}$;

$\lambda_B = 4.962 \times 10^{-10}\ yr^{-1}$; $^{40}K/K_{total} = 1.167 \times 10^{-4}$ mol/mol

*Minimum Age (sample does not meet petrographic criteria for a reliable age)

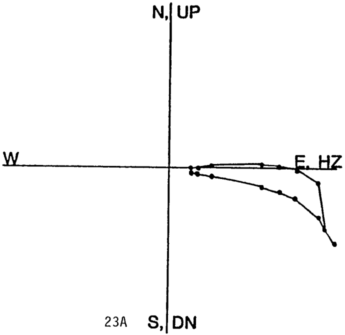
This age date was provided by the Geophysical Institute, University of Alaska--Alaska Division of Geological and Geophysical Surveys Cooperative Geochronology Laboratory.

to the Brothers Volcanics are younger than Albian.

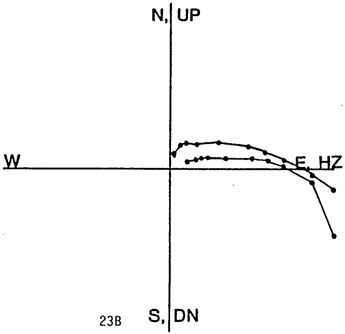
Paleomagnetic samples of the Brothers Volcanics were collected along the northeast shore of West Brother Island. Twenty-two samples were collected from 7 beds or flows over 11 m of section. In addition, 4 cores (samples 225-228) were taken from a flow (bed 7) approximately 300 m to the south.

A complete suite of specimens was subjected to stepwise AF cleaning. Most specimens responded very well to the treatment and cleaned to well-defined demagnetization end points (for example, figure 23a) at coercivities of 150-700 oersteds. The 4 AF specimens from bed 7 were directionally unstable. In all cases, characteristic directions were retained at cleaning intensities of at least 250 oersteds and several specimens preserved the characteristic direction at 1000 oersteds. A poorly defined secondary component of magnetization (declination = 201, inclination = 45, k= 5.1 in the geographic reference frame) was removed from 16 specimens by cleaning at 25 oersteds. There is little evidence for RRM in any of the specimens.

With the exception of specimens of 209A and the specimens from bed 4, all AF cleaned stable end point directions showed excellent within bed and between bed agreement (table 19). The one aberrant datum from bed 2 (209A) may be rejected on the basis of the theta-95 criterion. A replicate specimen from sample 209 was cleaned by thermal means in order to test whether good agreement could be produced by cleaning with a different technique. A steep secondary component was removed from specimen 209B by 300 degrees C and the characteristic direction, similar to other samples from bed 2, was stable



23A



23B

Figure 23 a) Typical AF demagnetization vector diagram from the Brothers Volcanics, GNB.07. b) Vector diagram from a clast within the Brothers Volcanics. The curved demagnetization path suggests the presence of two components of magnetization with overlapping coercivity spectra.

Table 19

Brothers Volcanics (GNB.7)

Longitude: 226.14
Latitude: 57.31

Bed/ Flow	Demag Range*	N	Geographic Dec	Inc	K	Stratigraphic Dec	Inc	K	VGP Long	Lat	K
1	200-300	3	78	58	186	81	14	173	319	11	383
2	250-400	6	87	51	162	87	8	186	316	5	238
3	500	3	76	69	1356	84	28	1285	313	16	2078
5	250-300	3	96	59	178	92	24	158	307	9	154
6	250-700	2	96	55	54	94	11	55	310	3	60
7	150-200	2	90	53	343	91	9	387	313	3	657
8	555T	2	84	49	248	78	14	327	323	12	549
	Mean		87	57		87	16		314	9	

	Geographic	Stratigraphic	VGP
K=	107.3	71.5	123.7
$\alpha 95$ =	5.9	7.2	5.4
N=	7	7	7

Bed 4 -- "Conglomerate Test"

Sample	Demag Level	Geographic Dec	Inc	Stratigraphic Dec	Inc
215A	500	198	27	175	32
216A	250	282	49	358	82
217A	300	110	30	108	-8

Bed 4 data are statistically random and constitute a "conglomerate test" (see text)

*Alternating Field Demagnetization except for Bed 8 (Thermal).

to a temperature of 555 degrees C.

While good directional agreement within bed 2 was obtained by thermal cleaning for 209B, it cannot be certain that this was produced by the different cleaning technique. Specimens 206 to 208 were collected about 1 m from the base of a volcanic flow breccia, while specimens 209 to 211 were collected near the flow breccia top. It is possible that a volcanic clast was cooled, magnetized and tumbled into its final position before magnetic resetting could be accomplished. In this scenario, specimen 209A and 209B would represent different clasts or aggregates of clasts within the flow breccia. This hypothesis gains some credence by considering the demagnetization characteristics of specimen 209A.

The Zijderveld plot of 209A displays an arcuate trajectory over most of the 500 oersted cleaning range (figure 23b). This suggests the presence of two magnetic components with overlapping coercivity spectra. The harder component is most likely to be abberant. In contrast, most other specimens show nearly straight line decay trends towards the origin, beyond the low coercivity secondary component when present. Since 209A probably records 2 different directions of magnetization and the specimens from the same flow breccia top do not show the same hard direction of magnetizations, the simplest explanation is reorientation of 209B (tumbling?), after acquisition of the hard direction, followed by partial or complete resetting of the component. It is interesting to note that the dispersion among the specimens from the interior of the flow (k=1067.6) is much lower than the disperson of directions from the flow breccia top (k = 105.7).

These precisions are significantly different at the 95 percent confidence level. This suggests that (for at least this flow) the interior of the flow is a better recorder of the field direction than the flow top. This observation is consistent with the contention that clasts on a flow breccia surface can rotate after initial TRM acquisition.

The AF data from bed 4 (specimens 215A-217A) were rejected because they are not internally consistent and are statistically random in direction. Replicate B specimens from bed 4 were cleaned by the thermal technique. Again the B specimens are not internally consistent; however, they are comparable to their individual AF cleaned counterparts. The within sample consistency and within bed inconsistency indicates that the specimens are recording stable directions but the directions were acquired before their deposition in bed 4, a tuffaceous deposit. This may be considered a small sample conglomerate test, which indicates a relatively cold emplacement origin for the bed. Moreover, the conglomerate test demonstrates that there has not been a remagnetiza- of this unit.

AF cleaned specimens 225-228 failed to yield stable end point directions in two cases, probably due to insufficient cleaning. Two of the thermally cleaned counterparts gave moderately well-defined characteristic directions at 555 degrees C. These thermal data are included in the overall mean.

The precision of the mean of bed mean directions is relatively high ($k = 71.5$), which could possibly reflect incomplete averaging of secular variation. Insufficient bedding attitude variation precludes the fold test. Reversals are not observed; however, the geographic direction is significantly different from the present field. The only stability test of remanence direction is the intraformational conglomerate test. This test requires that there has been no remagnetization since the time of deposition of the tuff clasts. As the tuff clasts are penecontemporaneous with the local rocks of the formation, this stability test strongly suggests that the remanence is primary. It must be emphasized that the "conglomerate" test only comprises 3 samples. Therefore, additional stability tests are desireable before final acceptance of the data.

Pybus Formation

Two paleomagnetic sampling localities were collected from the Pybus Dolomite along the shores of West Channel of Pybus Bay on the southeastern coast of Admiralty Island, southeast Alaska (contained within the Sitka B-1 quadrangle, figure 19). This formation was sampled because it lies within the Alexander terrane, was deposited during the Permo-Carboniferous Reversed (PCR) polarity interval and therefore, has the potential to test the southward before northward tectonic motion hypothesized by Panuska and Stone (1981) and Stone and others (1982).

The Pybus Dolomite was named by Loney (1964). It consists of medium to thickly bedded fossiliferous dolomite with thin lenses and beds of chert. Loney (1964) assigned the Pybus Dolomite a Permian age based on brachiopod identifications made by Dutro and Girty. Girty's identifications yielded an Artinskian (Early Permian) age.

Sample Locations

One locality in the Pybus Dolomite (ALX.02) is located approximately 2 kilometers west of Long Island on the southwest shore of West Channel. The second locality (ALX.01) is situated on the northeast shore of West Channel approximately 7.6 kilometers WNW of the first locality. The beds at ALX.01 dip towards the west at about 30 degrees and the ALX.02 beds dip 50 to 60 degrees eastward thus providing a fold test. No stratigraphic up-down indicators were observed at either locality. The Pybus beds are considered to be in their upright position based on the stratigraphically lower position of the older Cannery Formation mapped by Loney (1964). Both outcrops contain stylolites; however, thin section examination suggests that very little loss of material, due to pressure solution, has occurred.

Paleomagnetic Data

Nine samples were collected from 5 beds at ALX.01 over a stratigraphic interval of about 5 meters. At ALX.02, 39 samples were taken from 13 beds over 20 meters of section. In addition, 5 cores were drilled from a dike that intrudes the Pybus Dolomite at ALX.02 and 7 cores were drilled from one bed along 20 meters of strike extending from the dike, in order to apply the baked contact test.

Initially, both Pybus Bay suites of samples were subjected to thermal cleaning. Preliminary analysis of the ALX.01 suite yielded a fairly well-defined component at blocking temperatures of 350 to 450 degrees C. Initial or NRM intensities were approximately 2 x 10^{-7} emu/cc. The mean geographic vector direction for ALX.01 (declination = 75, inclinataion = 61, k=33.3) (table 20a) is close to the present field direction, although it is significantly different.

The ALX.02 suites of samples were exceptionally weak. NRM intensities ranged from 2 to 5 x 10^{-8} emu/cc. Stable demagnetization end points were determined using Zijderveld plots; however, allowance was made for the high noise to signal ratio (see figure 24 for representative plots). To assess the variability of the measured remanence direction, duplicate or triplicate measurements at one demagnetizaion level were frequently made. Widely divergent replicate measurements were dismissed as being magnetically unstable. The plot of the

Table 20

Pybus Formation (ALX.01)

Longitude: 225.85
Latitude: 57.37

Bed	Demag Level*	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K	VGP Long	VGP Lat	K
1	400	1	83	59	--	133	84	--	238	49	--
3	450	1	86	62	--	146	83	--	236	46	--
4	400	1	47	48	--	19	84	--	236	68	--
5	400	2	95	68	19	189	68	19	220	18	7
	Mean		75	61		165	83		231	46	

	Geographic	Stratigraphic	VGP
K=	33.3	46.6	14.3
$\alpha 95$ =	12.1	10.2	18.5
N=	4	4	4

Note: These directions are interpreted as secondary directions.

*Thermal Demagnetization in degrees C.

Re-evaluated Pybus Data (ALX.01)

Bed	Demag Level	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K	VGP Long	VGP Lat	K
1	525 T	1	132	23	--	146	36	--	258	-8	--
5	450 T/ 500 AF	2	137	44	345	166	38	672	239	-11	522
	Mean		134	34		156	37		249	-10	

	Geographic	Stratigraphic	VGP
K=	28.8	51.0	36.6
$\alpha 95$ =	48.4	35.7	42.6
N=	2	2	2

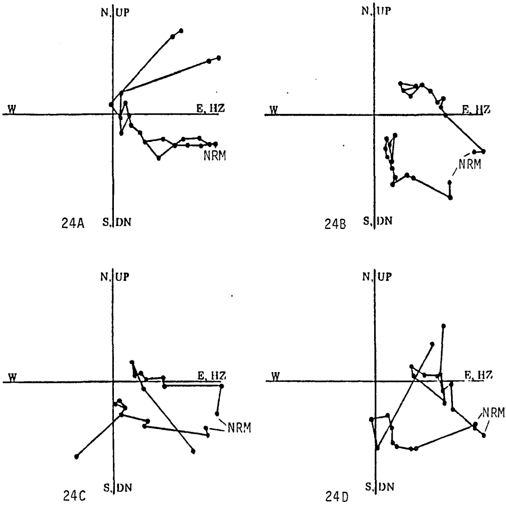


Figure 24 Thermal demagnetization vector diagrams for representative specimens of Pybus Dolomite. A crudely defined characteristic vector direction can be observed before encountering the aberrant intensity spikes. a) Specimen 164A. b) Specimen 197A. c) Specimen 200A. d) Specimen 202A.

preliminary selection gave a moderate cluster of points with several divergent directions (figure 25). The computed mean for this data set is declination = 144, inclination = 8 (k = 4.7). Quality control criteria were then applied to eliminate data with poor precision.

Due to the very low intensity of the Pybus samples, special care was taken to measure the sample holder's remanence and subtract this value from subsequent measurements. In this way, it was possible to obtain a background noise level of 3×10^{-9} emu/cc or lower. The first quality control criterion makes use of this noise level. For the least accurate measurement, this noise vector would be oriented at right angles to the specimen's remanence vector. An arbitrary value of 30 degrees from the true direction was selected as the lowest acceptable precision for any given measurement of direction. Assuming that the noise vector is oriented 90 degrees to the true remanence direction (the worst case), the measured remanence direction would be "off" by 30 degrees if the measured intensity were 6×10^{-9} emu/cc. Therefore, all data with an intensity less than 6×10^{-9} emu/cc were disregarded as they could be in excess of 30 degrees from the true direction. A similar quality control criterion was developed using instrument drift. The tolerance or instrument drift is the ratio of drift encountered during a measurement to the measured intensity expressed as a percentage. Using the 30 degree inaccuracy cut-off and assuming that the drift vector is oriented 90 degrees to the true vector, the maximum allowable tolerance is 50 percent (0.5 of the true vector or sin 30), without potentially violating the 30 degree cut-off.

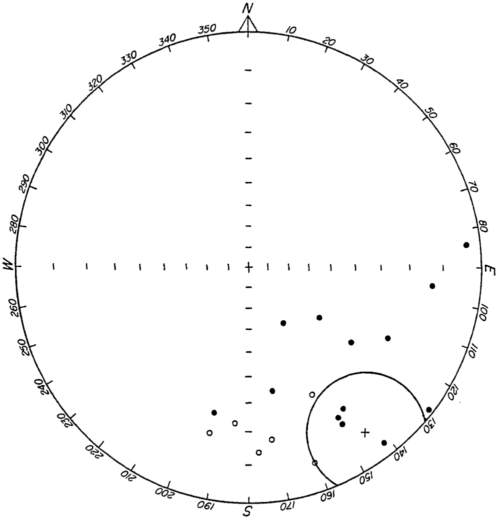


Figure 25 Stratigraphic vector directions for ALX.02, Pybus Dolomite (thermal data only).

After applying the measurement reliability criteria, 6 of the original measurements were rejected. The recalculated mean (declination = 154, inclination = 10, k = 10.7) is not signficantly different than the original mean; however, the precision (k) increased from 4.4 to 10.7. While the 30 degree value for maximum acceptable inaccuracy is arbitrary, this criterion was uniformly applied to all data. Thus, there is no subjective bias and this procedure is simply a noise reduction "filter".

The demagnetization level of measurements yielding reliable directions varied from 150 degrees to 450 degrees C with most being 300 degrees C or higher. These temperatures should not be interpreted as blocking temperatures as the intensity of most samples was driven into the noise level at intermediate temperatures, thus precluding a full thermal demagnetization. This does, however, raise some doubts as to the stability of the vectors. In order to evaluate the stability of the vectors in terms of coercivity, the replicate Pybus specimens were cleaned using AF techniques. Of the 28 replicates available, 3 specimens were taken from the dike, 2 were remagnetized by the dike, 14 specimens failed to yield a stable characteristic direction and 9 specimens responded well to the cleaning. The well-behaved specimens had coercivities of 100 to 300 oersteds. The mean direction of these acceptable vectors (with intensity and tolerance criteria applied) is declination= 149, inclination = 33 (k = 37.0). Considering the inaccuracies of measurement, this direction is rather close to the mean of the thermal data. The relatively high coercivities afford a greater degree of confidence in the stability of magnetization. Additional confidence is gained by a positive baked contact test. The best estimate of the mean direction of ALX.02 is the mean of the combined bed means of both AF and thermal data (figure 26, table 21).

In order to test for systematic errors in these data, two additional tests were performed. The first test consisted of zeroing the sample holder and recording the first 10 measurements of the empty holder that satisfied minimum drift requirements (no more than 5 x 10^{-5} gammas drift on any axis). If the holder's remanence has been removed from the measurements, there should be only a random noise direction remaining. The measured directions of the zeroed, empty holder are widely scattered and yield a k value of 1.4, which is statistically indistinguishable from a random distribution (Watson, 1956). Thus, there is no systematic measurement error due to the holder's remanence. The second systematic error check is designed to guard against the possibility of a drilling remanence overprint of the specimens. Although drilling remanence is commonly checked for, additional efforts were made for these samples because of their very low intensities. If such an overprint were acquired, it would be expected to align with the axis of the core. To test for such an alignment, the dispersion (k) of the orientation of the core axes was calculated and compared to the dispersion of the geographic vectors using the F-test. The precision of the ALX.02 core orientations (k = 3.4) is considerably lower than the precision of the corresponding geographic vectors (k = 13.2) and is statistically distinct. Therefore, it is very unlikely that the measured directions contain a significant axial (with respect to core specimens) overprint component.

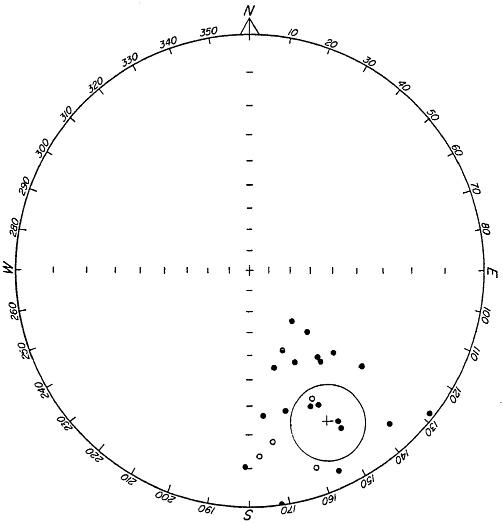


Figure 26 Stratigraphic vector directions for ALX.02, Pybus Dolomite. These directions comprise both AF and thermal demagnetization data with quality control criteria applied. See text for discussion of selection criteria.

Table 21

Pybus Formation (ALX.02)

Longitude: 225.90
Latitude: 57.28

Bed	Max. Demag AF/T*	N	Geographic Dec	Geographic Inc	K	Stratigraphic Dec	Stratigraphic Inc	K	VGP Long	VGP Lat	K
dike	200/	2	322	71	--	--	--	--	--	--	--
1	300/250	3	180	48	10	135	30	10	269	-5	19
2	100/	1	213	56	--	134	36	--	269	-4	--
3	300/	2	209	48	132	147	40	119	257	-6	110
4	100/	1	190	43	--	155	25	--	251	-17	--
dike	200/	1	57	74	--	--	--	--	--	--	--
5	150/300	4	219	53	30	153	42	30	245	-5	17
6	150/300	3	203	41	36	162	23	38	244	-19	41
7	100/300	2	224	62	10	149	22	10	258	-16	22
8	/200	1	181	55	--	149	14	--	259	-21	--
10	/150	1	139	53	--	128	1	--	283	-19	--
11	/450	2	174	23	7	168	-9	6	262	-42	12
12	/350	2	178	20	47	172	-9	46	236	-37	157
13	/450	2	167	6	40	169	-10	46	240	-37	43
	Mean		186	44		152	18		255	-19	
	K=		13.3			11.9			20.1		
	α95 =		11.1			11.8			9.0		
	N=		12			12			12		

Note: Dikes not included in calculation of the mean.

*Data include both Alternating Field and Thermal Demagnetization.

The initial evaluation of the ALX localities was that they failed the fold test, suggesting that one or both localities were remagnetized. However, on closer examination, it was noticed that the ALX.01 specimens had almost an order of magnitude stronger NRM intensities than the ALX.02 specimens. The ALX.01 specimens were then re-examined to determine whether the higher intensity might reflect a strong secondary component.

For the re-evaluation of the ALX.01 specimens the highest temperature interval component was selected for the mean if it satisfied reliability criteria. The measurements could not exceed the tolerance or intensity criteria previously established. The measurements had to be significantly different than the secondary component. The specimen could not show evidence of magnetic instability, such as abrupt changes in demagetization trend or drastically different directions (greater than 70 degrees) obtained for repeat measurements at a given demagnetization level. Finally the difference vector had to be significantly different from the secondary component, indicating removal of all of the secondary magnetization. If these reliability criteria are applied, only 3 specimens from 2 beds can be used in computing the mean (table 20b). The new mean stratigraphic vector direction is remarkably close to the ALX.02 stratigraphic mean. It is also interesting to note that the difference between the ALX.01 and ALX.02 mean vectors decreases from 40 degrees to 20 degrees (after tilt correction), suggesting a positive fold test.

The paleomagnetic data for the Pybus Dolomite passed some stability tests, while others are inconclusive. The geographic mean directions are significantly different from the present field and ALX.02 passes the baked contact test. Reversals are not to be expected as the Pybus Dolomite was deposited during the PCR magnetic interval. The fold test is suggestive of a stable primary remanence; however, it must be regarded as inconclusive since the mean direction for one limb (ALX.01) is based on very few data. The evidence at this point does suggest that the paleomagnetic directions of the Pybus Dolomite could well be a stable, albeit a weak, primary remanence. In view of the importance of data from the Permian of the Alexander terrane and the potential of the Pybus Dolomite to help fill this information gap, the need for conclusive stability tests and tighter control on magnetic direction must be emphasized.

Geological Constraints on Terrane Amalgamation

The pertinent regional geologic relationships must be established before inferring tectonic or paleogeographic significance from the palomagnetic data. This is particularly important because paleomagnetic data do not constrain ancient longitude; that is, once a paleolatitude has been determined, any position on the earth within that latitudinal band is (paleomagnetically) permissible. By evaluating the gelogical data, it is possible to infer the ages at which terranes were linked to each other (amalgamation) and to the craton (accretion). Linkage ages along with biogeographic data can help to provide paleolongitude control for more realistic paleogeographic reconstructions. In addition, a knowledge of terrane amalgamation ages will provide a time frame indicating when paleomagnetic data from one terrane may be applied to another. For the purpose of this discussion, only the major terranes south of the Denali fault will be considered (see figure 2 for terrane locations).

No clearcut structural contact between the Peninsular terrane and Wrangellia has yet been identified. The understanding of stratigraphic ties between the two terranes is severly hampered by the paucity of occurrences of Triassic and older rocks within the Peninsular terrane (Jones and Silberling, 1979). Jones and others (1977) distinguish the Peninsular terrane from Wrangellia on the basis that the Peninsular terrane lacks Traissic rocks older than late Carnian (which are present in Wrangellia). No volcanic rocks with proven ages similar to the Wrangellian Nikolai Greenstone (Ladinian to Carnian) have been reported in the Peninsular terrane. However, there are Triassic volcanics of apparent Norian age (Jones and others, 1977). Jones and Silberling (1979) suggest that Wrangellia and the Peninsular terrane retained their individuality through the Early Jurassic, since there is no evidence in the Wrangell Mountains for Early Jurassic volcanism which is well represented in the Peninsular terrane by the Lower Jurassic Talkeetna Formation. By Late Jurassic, or perhaps Middle Jurassic, the Peninsular and Wrangelia terranes show similar stratigraphies, suggestive of a common history (implying close geographic association) from that time on (Jones and Silberling, 1979).

Hudson (1979) discusses evidence from batholithic belts which tends to support a Middle Jurassic minimum age of amalgamation. The Aleutian Range-Talkeetna Mountains plutonic belt in the Peninsular terrane may be in part coeval with the Tonsina-Chichagof plutonic belt in Wrangellia (Hudson, 1979). Both belts have similar compositions and could be parts of the same magmatic arc (Hudson, 1979). If this is true, then the Wrangellia and the Peninsular terranes would have been joined by at least Middle Jurassic time.

Newton (1983) has studied the Upper Triassic molluscan faunas of the Wrangellia and the Peninsular terranes. These two terranes have similar bivalve species compositions from Norian (Upper Triassic) carbonates (Newton, 1983) This faunal similarity suggests that the Wrangellia and the Peninsular terranes were in relatively close proximity to one another (i.e. close enough for faunal exchange) in Late Triassic time, although they need not have been amalgamated (Newton, 1983).

The evidence points towards a Middle Jurassic minimum age of amalgamation of the Peninsular and Wrangellia terranes. However, it should be pointed out that the Triassic distinctions between the Peninsular terrane and Wrangellia are not compelling. Thus, it is possible that these two terranes are actually one and the same.

Jones and others (1977) have demonstrated marked differences between the Triassic stratigraphies of Wrangellia and the Alexander terrane. The Triassic of the Alexander terrane lacks the very thick basalts and the thick sequence of platform carbonates, which are characteristic of the Wrangellia sequence (Jones and others, 1977). This difference in Triassic stratigraphy, in addition to distinct faunas (Newton, 1983, Silberling and Jones, 1983), suggests that these terranes were geographically separated during Triassic time. The geographic separation between Wrangellia and the Alexander terrane is confirmed by distinct paleomagnetic paleolatitudes determined by Hillhouse (1977) and Hillhouse and Gromme (1980).

The contact between Wrangellia and the Alexander terrane in the eastern Wrangell Mountains has been identified as a major regional thrust along which Wrangellia was emplaced over the Alexander terrane (MacKevett and Jones, 1975). The suturing of these two terranes was apparently complete by the end of Jurassic time because the Gravin-Nutzotin belt, a Late Jurassic to Early Cretaceous assemblage of flysch and interbedded andesitic volcanic rocks, depositionally overlies both the Alexander terrane and Wrangellia (Berg and others, 1972). Thus, the age of Alexander-Wrangellia amalgamation is probably post Late Triassic and no later than Late Jurassic. The age of this amalgamation event can be further constrained by reconnaissance paleocurrent studies, made peripheral to this paleomagnetic investigation, combined with considerations of the regional mapping of MacKevett (1978).

Based on mapping by MacKevett (1978), the "Wrangellia-Alexander thrust" cuts upper Paleozoic and minor Upper Triassic rocks of Wrangellia, displacing them over the Devonian (Muller, 1967) and/or Ordovician (Read and Monger, 1975) Kaskawulsh Group of the Alexander terrane (figure 27). The thrust is apparently overlain by Lower Cretaceous rocks (MacKevett, 1978). This requires the age of thrusting to be Late Triassic to Early Cretaceous. If this thrust is part of a thrust system related to the thrust faults to the northwest, the age of thrusting can be determined more precisely. In the McCarthy C-5 and C-6 quadrangles MacKevett (1971, 1972) mapped thrust faults that cut rocks as young as Late Jurassic (the Root Glacier Formation) and are overlain by Lower Cretaceous (Albian) rocks. If the fault correlations are valid, the age of the Wrangellia-Alexander suturing corresponds to the Late Jurassic to Early Cretaceous orogeny recognized by MacKevett (1978). This interpretation is reasonable in view of the fact that no other Late Triassic to Early Cretaceous orogenic event is known in the Wrangell Mountains. Additional evidence bearing on the Late Jurassic amalgamation of the Wrangellia and Alexander terranes can be found in paleocurrent directions of the Root Glacier Formation.

Sixty-one orientations of ripples and large scale cross beds were

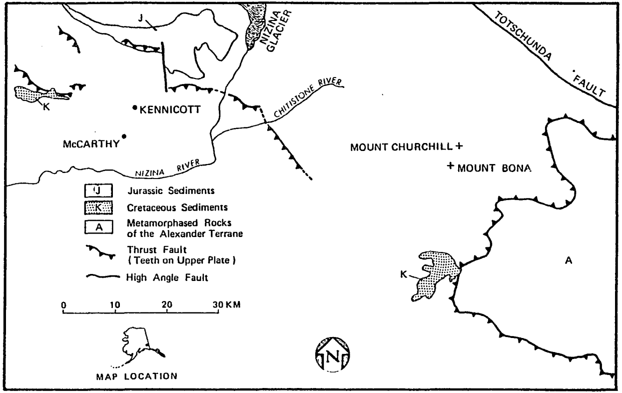


Figure 27 Generalized geologic map showing selected units and structures in the eastern McCarthy Quadrangle.

measured from thin to medium-bedded sandstones in the Root Glacier Formation. The measurements were taken from 6 outcrop localities approximately 3 km south of the Stairway to Heaven Icefall, along the eastern valley wall of the Root Glacier (McCarthy C-5 quadrangle) and 1 reconnaissance paleocurrent locality on the western Root Glacier valley wall (P.C. 0). These data, which span about 500 m of section, are plotted in figure 28 and tabulated in table 22. A WNW paleocurrent direction is indicated. The apparent bimodal distribution is probably the result of oversampling the limbs of trough cross beds. This contention is supported by the fact that trough axes and average directions determined from both limbs of the same troughs (where observable) are virtually identical to the grand mean. Although a greater geographic distribution of paleocurrent data could not be obtained, the large stratigraphic thickness sampled suggests that the mean direction is probably not a short lived, localized paleocurrent direction. However, more data are needed to prove this hypothesis.

The mean paleocurrent direction suggests that Root Glacier Formation sediment was shed from the vicinity of the Wrangellia-Alexander thrust. Paleocurrent evidence along with the penecontemporaneous onset of terrigenous sedimentation and the Wrangellia-Alexander thrusting suggests a causal relationship. This allows the speculation that the Late Jurassic collision of the Wrangellia and Alexander terranes generated an area of high relief, in the vicinity of the present thrust, which then supplied sediment to the Root Glacier Formation. Penecontemporaneous deposition of the Gravina-Nutzotin flysch on the craton-ward side of this composite terrane further suggests that the

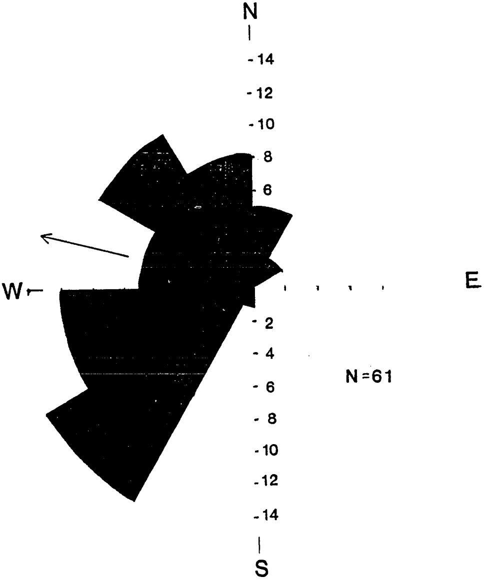


Figure 28 Paleocurrent rose diagram for the Upper Jurassic Root Glacier Formation.

Table 22

Paleocurrent Data
Root Glacier Formation

Locality	Azimuth	Standard Deviation	N
PC 0	290	20	3
PC 1+2	253	38	12
PC 3	235	13	5
PC 4	273	62	3*
PC 5	342	65	17
PC 6	290	37	21
All Data	285	54	61
Mean of Means	285	38	5

*Indistinguishable from random.

Late Jurassic orogeny was not the accretionary event but rather an arc-arc collisional amalgation event in an oceanic setting.

To the south of the Peninsular-Wrangellia-Alexander composite arc are two major tectonostratigraphic terranes, the Chugach and Prince William terranes (figure 2), which must also be included in any tectonic analysis. The Chugach and Prince William terranes are Cretaceous and early Tertiary units consisting of melange, flysch and volcanic rocks. Moore and others (1983) describe these units as accretionary complexes. As employed by Moore and others (1983), "accretionary" refers to subduction accretion and may or may not be related to (cratonal) accretion, in the sense of Jones and others (1983).

Jones and others (1981) describe the Chugach terrane in terms of two subterranes: a polymictic disrupted terrane and a coherent flysch terrane. The disrupted (melange) terrane consists of blocks of chert, limestone, volcanic, plutonic and untramafic rocks contained in a cherty, tuffaceous argillitic matrix. Radiolarian ages from the cherty matrix range from Late Jurassic to Early Cretaceous. The second subterrane is composed of graywacke, argillite and slate (flysch) of Late Cretaceous age (Campanian to early Maestrichtian). Pavlis (1982) argues that the disrupted terrane in the Chugach Mountains (the McHugh Complex) was emplaced against the composite arc terrane, along the Border Ranges fault by Early Cretaceous (Albian) time, on the basis of undeformed, radiometrically dated plutons which cross-cut the fault. Decker (1980) suggests a somewhat later age of emplacement on the basis that the mid Crectaceous metamorphosed Kelp

Bay Group was emplaced along the Border Ranges fault and cut by post-tectonic diorite sills of inferred Late Cretaceous or earliest Teritary age.

The Ghost Rocks Formation of the Prince William terrane consists of both disrupted and coherent flysch and volcanics of early Tertiary age, and is interpreted as offscraped trench deposits (Moore and others, 1983). The Ghost Rocks are cut by plutons, dated as 62-63 m.y., of the Kodiak batholith, which also intrudes the flysch subunit (Kodiak Formation) of the Chugach terrane (Moore and others, 1983). Therefore, by 62 m.y. the Chugach and Prince William terranes were linked together (Moore and others, 1983).

It thus appears that the Chugach melange can be linked to the composite arc terrane by mid to Late Cretaceous time and the Prince William can be linked to the Chugach flysch by early Tertiary time. Moore and others (1983) suggest that the Chugach and Prince William accretionary complexes formed adjacent to the older composite arc terrane to the north; however, they point out that no demonstrable relationships have yet been found which unequivocally make that link. Paleomagnetic data offer some support for this hypothesis. Plumley and others (1983) have determined a paleomagnetic paleolatitude of 40 degrees N for the Paleocene Ghost Rocks Formation in the Prince William terrane. The latest Cretaceous MacColl Ridge Formation (Wrangellia) paleolatitude determined in this study is 32 degrees N. Given the uncertainties of the paleolatitude determinations and the minor age differences between the formations, it is probable that the Prince William terrane (and by implication the Chugach terrane) was located at the same latitude as the Peninsular-Wrangellia-Alexander arc in early Tertiary time. The paleolongitude of these terranes is indeterminate at present. It must be emphasized that the paleolatitude agreement is a necessary but not sufficient condition for proving the paleogeographic ties between the arc and the trench deposits.

Based on the previous discussion, it can be established with reasonable confidence that by Late Jurassic time there existed an amalgamated composite arc terrane consisting of the Peninsular, Wrangellia and Alexander terranes along with the Gravina-Nutzotin overlap assemblage. There is circumstantial evidence for including the Chugach terrane in early Tertiary time, in this composite arc assmeblage. For the purposes of tectonic analysis, this composite arc (and trench ?) superterrane needs a separate designation. The term "Talkeetna superterrane" proposed by Csejtey and others (1982) is inadequate as it only comprises the Wrangellia and Peninsular terranes. It is here proposed that the "Southern Alaska superterrane" be defined as a Late Jurassic amalgamated superterrane consisting of the Peninsular, Wrangellia and Alexander terranes plus the Gravina-Nutzotin belt. It is also suggested that the Chugach and Prince William terranes be provisionally accepted as Upper Cretaceous and lower Tertiary (respectively) components of the Southern Alaska superterrane.

Paleogeographic Analysis

Wrangellia

Figure 29 is a representation of VGP positions for existing paleomagnetic data from the Wrangell Mountains, Nutzotin Mountains and the Alaska Range portion of Wrangellia. The Triassic paleomagnetic data were taken from Packer (1972), Hillhouse (1977) and Stone (1982). On the basis of known polarity of the Pennsylvanian and Permian data, the late Paleozoic and Triassic VGPs are considered to be estimates of the north geographic pole position with respect to the Wrangellia terrane. These data strongly suggest that Wrangellia occupied a northern hemisphere position (10-20 degrees north latitude) during the late Paleozoic and early Mesozoic as originally proposed by Panuska and Stone (1981). Although paleomagnetic data can provide paleolatitude constraints, in most cases, they offer no paleolongitude control; howevr, biostratigraphic data can provide such paleogeographic evidence.

Newton (1983) has demonstrated a faunal similarity between Norian (Late Triassic) bivalves from Wrangelia and North America. Three species of Septocardia were identified in Wrangellia which are endemic to the Americas and indicate tropical paleolatitudes (Newton, 1983). Although the Wrangellia faunas show the greatest similarity with the Auld Lang Syne and Star Peak Groups of Nevada (suggesting a northern hemisphere paleolatitude), a confident biostratigraphic hemisphere determination is precluded by insufficient faunal data from autochthonous rocks from both North and South America (Newton, 1983).

Silberling and Jones (1983) have identified three Late Triassic

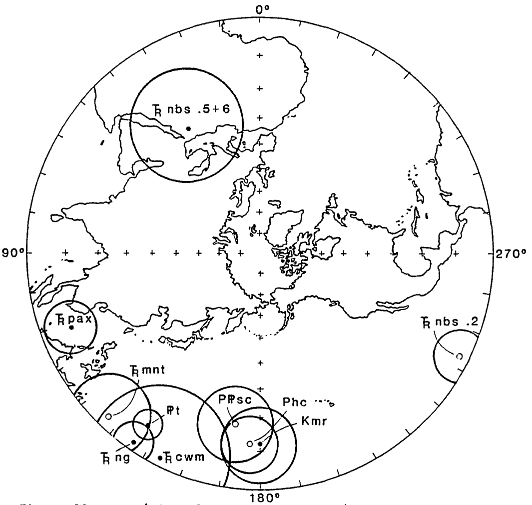


Figure 29 VGPs (virtual geomagnetic poles) for the Wrangell Mountains and nearby portions of Wrangellia. PPsc-Permian/Pennsylvanian Station Creek Formation, Phc-Permian Hasen Creek Formation, Pt-Pennsylvanian Tetelna Volcanics, TRng-Triassic Nikolai Greenstone (Hillhouse, 1977). TRnbs.5+6,TRpax, TRmnt, TRcwm and TRnbs.2-Triassic Nikolai Greenstone and equivalents (Stone, 1982). Kmr-Cretaceous MacColl Ridge Formation.

latitudinal zones in western North America based on species of the bivalve genus Monotis. The Wrangellian Monotis species are characteristic of low paleolatitudes similar to the species found in Nevada (Silberling and Jones, 1983). This Monotis fauna is apparently distinct from southern hemisphere varieties (Silberling and Jones, 1983).

The distribution of Septocardia and Monotis suggests that Wrangellia occupied a low latitude position in the eastern paleo-Pacific Ocean. Wrangellia must have been close enough to North America to allow faunal exchange, although they need not have been in contact (Newton, 1983). The biostratigraphic data from Late Triassic bivalves tend to corroborate the low paleolatitude estimate based on paleomagnetism. Moreover, the Monotis fauna of Wrangellia suggests that the northern hemisphere interpretation based on paleomagnetic data is correct.

The VGP for the Upper Cretaceous MacColl Ridge Formaion shown in figure 29 is also considered to be the north VGP. This VGP position corresponds to 32 degrees N paleolatitude. This polarity option is preferred because it requires only 30 degrees of latitudinal motion instead of 90 degrees since Late Cretaceous time. In addition, the late Campanian to early Maestrichtian (the age of the MacColl Ridge Formation) geomagnetic field is strongly biased towards normal polarity (Lowrie and Alvarez, 1981, Irving and Pullaiah, 1976). The polarity bias observed in the MacColl Ridge Formation (table 12) thus strongly favors the northern hemisphere polarity option. If the Moonshine Creek Formation data is a primary remanence, then the VGP plotted in figure 29 is also a known north VGP, since the Moonshine Creek Formation was deposited during the Cretaceous long normal polarity interval (Lowrie and Alvarez, 1981). The geographic locations of known and probable north VGPs for the southcentral Alaska portion of Wrangellia yield a relatively tight distribution within the southwest Pacific Ocean, with three noteable exceptions. VGPs from localities NBS.5 + 6 and NBS.2 from the Triassic Nikolai Greenstone are significantly displaced from the southwest Pacific cluster. It is possible that these aberrant directions are due to remagnetization; however, the existence of reversals in locality NBS.5 + 6 (Stone, 1982) strongly suggests that remagnetization has not occurred at this site. The fact that NBS.2 give a paleolatitude similar to localities with demonstrably stable primary remanence directions is circumstantial evidence, albeit weak, that it, too, is primary. The discordance of the NBS.5 + 6 and NBS.2 VGPs can be explained in terms of local teconic rotations generated by a strike-slip fault system, similar to teconic rotations observed in California (reviewed in McWilliams, 1983). This interpretation is reasonable in view of the proximity of these localities ot the Totschunda fault (Stone, 1982). This argument, however, cannot be used to explain the spurious Moonshine Creek VGP.

The Cretaceous Moonshine Creek Formation was deposited on the same structural block that contains the late Paleozoic Skolai Creek section, sampled for this study, and the Triassic Nikolai Greenstone studied by Hillhouse (1977). Thus, a post mid-Cretaceous rotation cannot account for the position of the Moonshine Creek VGP, since such rotation should affect all localities equally. A pre-latest Cretaceous rotation is apparently precluded by the lack of a substantial declination difference between the MacColl Ridge VGP and the VGPs obtained from the rocks beneath the Moonshine Creek locality. Therefore, with the exclusion of a relative rotation origin for the VGP discrepency, it is most likely that the Moonshine Creek Formation locality (WRG.18) has been remagnetized, as suggested by the close agreement of a recent geomagnetic field direction and the geographic vector direction.

The grouping of the late Paleozoic and Mesozoic Wrangellia VGPs suggests that the southcentral Alaska portion of the terrane behaved, for the most part, as a coherent tectonic block. If a structurally undisrupted (in terms of local rotations) Wrangellia can be demonstrated by subsequent studies, two important ramifications follow. The first implication is that the southcentral portion of Wrangellia has undergone approximately 130 degrees counterclockwise rotation (or 230 degrees of clockwise rotation), since latest Cretaceous time. If other portions of the terrane are similarly intact, oroclinal bending hypotheses (bending about a vertical axis over several hundred kilometers of arc length) could be tested. In addition, VGPs from the terrane's stable interior could then be used in paleogeographic reconstructions, to provide paleo-north direction as well as paleolatitude.

Alexander Terrane

The first apparent polar wander (APW) path segment for the Alexander terrane was constructed by Van der Voo and others (1980). The data were acquired from Middle Ordovician through Middle Pennsylvanian rocks from western Prince of Wales Island. The primary nature of the remanence is confirmed by numerous fold tests and reversal tests. A northern hemisphere location of the Alexander terrane is preferred on the basis that a minimum motion with respect to North America is required and that they both have similar VGP positions and generalized APW paths (Van der Voo and others, 1980). A paleolatitude versus time plot for this same data is at odds with the predicted trend, assuming, however, that the Alexander terrane was fixed geographically with respect to North America. A much better paleolatitude versus time fit between the Alexander terrane and North America is obtained for a paleo-position in western Nevada/northeastern California (Van der Voo and others, 1980). This is strong evidence in support of an original location of Alexander terrane near present day Califoria, a concept originally proposed by Jones and others (1972). Such a paleogeographic model is further strengthened by a provenance study (Girty and Wardlaw, 1984), which suggests that the Alexander terrane supplied detritus to the pre-Late Devonian Shoo Fly Complex, California. Based on the fit of Alexander terrane paleolatitudes for Ordovician through Middle Pennsylvanian time with a Nevada/California origin, Van der Voo and others (1980) concluded that the Alexander terrane was fixed with respect to ancient Nevada/California through Pennsylvanian time and subsequently was displaced to its present southeast Alaska position.

Hillhouse and Gromme (1980) have reported paleomagnetic data which bears upon the timing of accretion of the Alexander terrane. These authors have sampled the Upper Triassic Hound Island Volcanics in Keku

Straight and determined a 44 degree paleolatitude. This determination is in good agreement with the predicted paleolatitude, assuming that the Alexander terrane was in its present position with respect to North America. On this basis, Hillhouse and Gromme (1980) and Van der Voo and others (1980) suggested that Alexander rifted away from Nevada/California in post-Pennsylvanian time, moved approximately 15 degrees northward and accreted to North America by Late Triassic time. Such an interpretation, however, is difficult to reconcile with the accretion history of inboard terranes.

Based on paleomagnetic paleolatitude data Monger and Irving (1980) suggest Jurassic or Cretaceous age of accretion for the Stikine terrane in British Columbia (figure 30). Tempelman-Kluit (1979) proposed that the Late Jurassic deformation in the Stikine terrane is the manifestation of the arc-continent collision, which emplaced Stikinia. If the Triassic accretion age for the Alexander terrane is correct, then the Stikine terrane would have to be wedged behind the earlier accreted terrane, as proposed by Monger and Irving (1980), or thrust over the Alexander terrane. Neither scenario is appealing due to the mechanical and geometric problems associated with the "wedging" model and the lack of any evidence for the overthrust model.

In addition to problems with the Stikine accretion, a Triassic Alexander accretion is apparently precluded by paleomagnetic data from other southern Alaska terranes. Since Wrangellia, Alexander and the Gravina-Nutzotin terranes were linked by Late Jurassic time, any paleomagnetic data post-dating this amalgamation from any one terrane should apply to all. The very low paleolatitude found in the MacColl Ridge

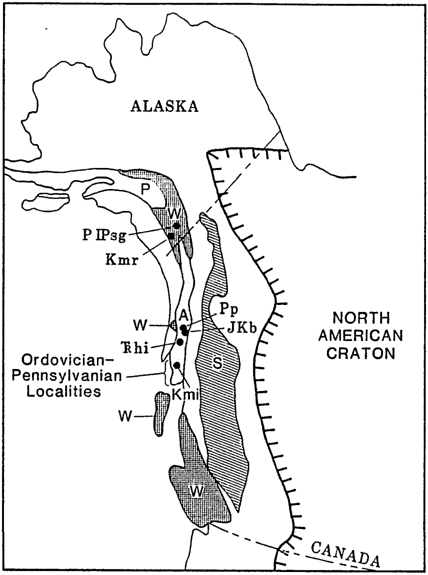


Figure 30 Map showing selected terranes of northwest North America. A-Alexander terrane, W-Wrangellia, P-Peninsular terrane, S-Stikine terrane. Key paleomagnetic sample localities are shown for reference. PPsg-Skolai Group, Kmr-MacColl Ridge Formation, Pp-Pybus Formation, JKb-Brothers Volcanics, Trhi-Hound Island Volcanics, (Hillhouse and Gromme, 1980), and Kmi-Marsh Island (Panuska and others, 1984). Ordovician-Pennsylvanian localities are from Van der Voo and others (1980).

Formation (Wrangellia) and suggested by the Brothers Volcanics (Gravina-Nutzotin Belt) imply that the Alexander terrane was far south of its present location in Cretaceous time. It is possible that the Alexander terrane accreted in Triassic, rifted from North America, amalgamated with Wrangellia and reaccreted in post-Cretaceous time; however, such a scenario is unduly complicated and difficult to defend.

The Triassic Hound Island data can be interpreted to represent a southern hemisphere position to circumvent the problems of accretion timing. The paleomagnetic data for Alexander taken as a whole tend to support such an inference. According to the data presented by Van der Voo and others (1980), when "last seen" in Early or Middle Pennsylvanian time, the Alexander terrane was moving south (figure 31). That is, in Devonian to Pennsylvanian time, Alexander moved from 21 degrees to 8 degrees north latitude, although this latitudinal motion is just barely within the limits of detectability. The data from the Pybus Formation (this study) is of known polarity and corresponds to a 9 degree south paleolatitude. This suggests that there was net southerly motion of the Alexander terrane in Devonian to Pennsylvanian time and requires southerly motion between the Devonian and the Permian. In addition, the Pybus Formation VGP (shown in figure 32) plots relatively close to the earlier Paleozoic VGPs. Since the Pybus is a known north VGP, this tends to confirm the polarity preference of Van der Voo and others (1980) and also suggests that the portion of the Alexander terrane sampled has undergone little or no differential rotation. If the premise of little differential rotation is correct, then a terrane-wide APW path can be constructed. VGP data from the

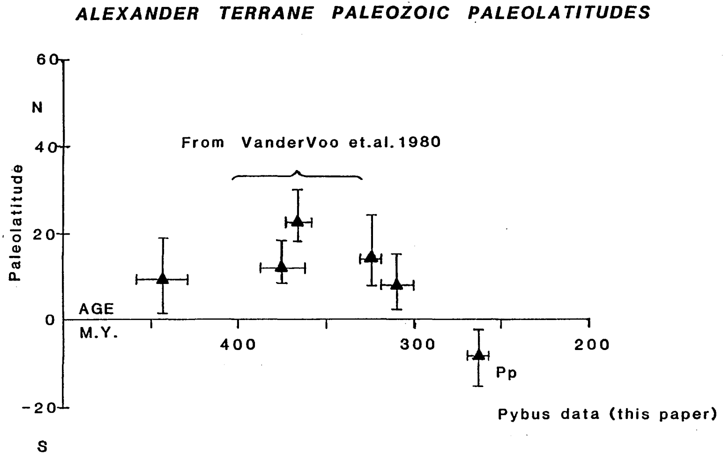


Figure 31 Paleozoic paleolatitudes for the Alexander terrane. Pybus data is from this paper. Other data are from Van der Voo and others (1980).

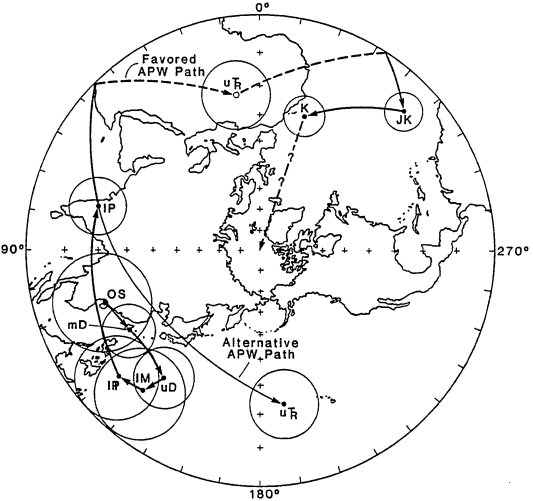


Figure 32 Alexander terrane VGPs and a possible apparent polar wander (APW) path. Solid dots are northern hemisphere. Circles are southern hemisphere. The northern hemisphere Late Triassic VGP and "alternative APW path" are not favored (see text). Ordovician through Pennsylvanian VGPs are from Van der Voo and others (1980). Permian VGP is from the Pybus Dolomite (this paper). Jurassic/Cretaceous VGP is from the Brothers Volcanics (this paper). Cretaceous VGP is from Panuska and others (1984).

Alexander terrane are shown in figure 32 with two possible generalized APW paths. The Paleozoic VGPs are almost certainly north poles because of the known polarity of the Pybus pole and the arguments of Van der Voo and others (1980) previously discussed.

Additional paleomagnetic data from Marsh Island (Panuska and others, 1984, in press), located 80 km south of Petersburg, can be used to constrain APW path models. The Marsh Island VGP (from Panuska and others, 1984, in press) and the Brothers Volcanics VGP are considered to be north poles because the Marsh Island rocks (100 m.y. K-Ar age) were deposited during the Cretaceous long normal polarity interval and the Brothers Formation is considered to be of similar age. Hillhouse (1983, written communication) has expressed concern that the Marsh Island rocks are remagnetized. Hillhouse found _in situ_ geographic vector directions from Paleozoic and Triassic rocks which are similar to the Marsh Island _in situ_ direction, suggesting remagnetization. These directions are close but the Marsh Island direction is statistically distinct in all but one case. Unfortunately, Hillhouse's data represents the mean NRM directions. Although AF and thermal demagnetization of some pilot specimens did not substantially change the direction (Hillhouse, 1983, written communication), it is not known whether magnetic cleaning of all specimens might alter the mean directions sufficiently to give a more definitive test.

A recent K-Ar age date on a lahar deposited within the Marsh Island section gives a 100 m.y. age, which is in excellent agreement with the Albian paleontological age determination (Decker, 1983, personal communication). Since the K-Ar age of the rock has not been reset, it is unlikely that the magnetic directions of the rocks can have been reset by heating above their blocking temperatures. The age quoted was determined for a hornblende separate (Blum, 1963, personal communication). The minimum argon blocking termperatures of hornblende is approximately 460 degerees C (Odin, 1982). While the Argon blocking termperature is below the magnetic blocking temperature of these rocks (Panuska and others, 1983), the possibility of high temperature viscous thermoremanent remagnetization cannot be unequivocally ruled out. The remagnetization curves of Pullaiah and others (1975) would allow the Marsh Island rocks to be remagnetized if they were heated to just below the Argon blocking temperature and maintained at that temperature for about 50 m.y. Perhaps the strongest test for magnetic stability of the Marsh Island rocks is the general agreement of direction with the Brothers Volcanics (which passes the conglomerate test). Thus, although there is only circumstantial evidence for the stability of the Marsh Island rocks, the Marsh Island data is tentatively accepted as a primary direction because there are two different lines of evidence suggesting stability.

The major concern in proposing an APW path is the correct polarity assignment for the Triassic Hound Island VGP. The southern hemisphere Triassic VGP is preferred for a variety of reasons. The southern pole forms a continuous, great circle APW path from Pennsylvanian to Triassic time. It requires only 76 degrees APW motion and 35 degrees of latitudinal motion of the area sampled since Early Permian. In addition, the southern hemisphere pole is only 62 degrees from the probable Brothers north VGP and only 53 degrees from the Marsh Island known north VGP. The alternative northern hemisphere VGP for the Hound Island rocks would require 102 degrees of APW motion with 53 degrees of latitudinal motion and would produce several pronounced kinks in the APW path. Thus, on the basis of simplest APW path and least latitudinal motion, the Hound Island VGP located in the southern hemisphere is considered the most likly estimate of the geographic north pole.

There is some paleontological support for the Triassic southern hemisphere location for the Alexander terrane. Silberling and Jones (1983) report a _Monotis_ fauna from the Alexander terrane which is unlike the North America cratonic _Monotis_ fauna and may represent southern hemispere middle paleolatitudes. In addition, Newton (1983) has identified _Septocardia_ cf. _S._ _pascoensis_ (Cox, 1949) in Alexander terrane Triassic rocks from Keku Straight. This species is markedly different from the North America _Septocardia_ and the sole previously reported occurrence is in Peru (Newton, 1983).

Although the Triassic biostratigraphic data tend to support a southerly before northerly tectonic motion for the Alexander terrane, the data are too sparse to deomonstrate conclusive ties. Similarly, the paleomagnetic data are too sparse to assess the likelihood of local tectonic rotations and clearcut stability tests for some localities are lacking. Thus, the apparent polar wander path presented in figure 32 is offered only as a working hypotheses, which will no doubt be modified by future studies. Nevertheless, tectonic analyses based on this curve are useful in pointing out correlations and conflicts and for generating testable tectonic models.

In addition to estimation of paleolatitude changes, the APW path for the Alexander terrane can be used to estimate absolute plate motions. The Alexander terrane VGPs for Pennsylvanian, Permian and Triassic times sweep out an approximate great circle path. Since each VGP must coincide with the north geographic pole of the earth during deposition of its sample locality and since each VGP must be displaced southward to "accomodate" the next VGP, this segment of the APW path must have conformed to a line of longitude. By measuring the angle between the "best fit" great circle segment of the APW path and the great circle connecting the geographic position of the sample locality and the various VGPs, the longitude of the terrane relative to the APW path (the reference longitude) can be obtained (figure 33). Since the great circle APW path is within the confidence limits of the Devonian and Mississippian VGPs, their relative longitudes were also estimated. These relative longitudes coupled with the corresponding paleolatitudes will give the absolute paleogeography of the terrane relative to the reference longitude (the APW path) (figure 34). The displacement of the terrane and the age differences between the VGPs allow the calculation of absolute plate velocities. The plate motion from Mississippian to Permian is 4 cm/yr, and the Devonian to Mississippian plate speed is approximately 2 cm/yr. These plate velocities can only be viewed as an order of magnitude approximation, since the combined error limits would permit velocities as high as 13 cm/yr to 5 cm/yr retrograde motion. The Permian to Triassic plate motion is approximately 13 cm/yr, with error limits allowing 8 to 18 cm/yr. Moreover, the absolute motion path shows that the Alexander terrane moved in a southerly direction from Devonian to Permian followed by

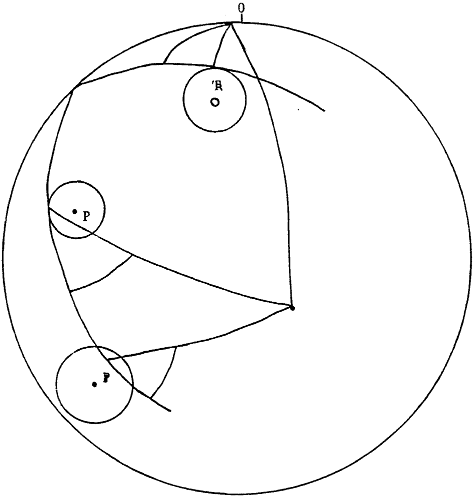


Figure 33 Best fit great circle to the Alexander terrane APW path. The angular measure between this great circle and the great circle connecting VGPs and the sampling locality provides an estimate of paleolongitude.

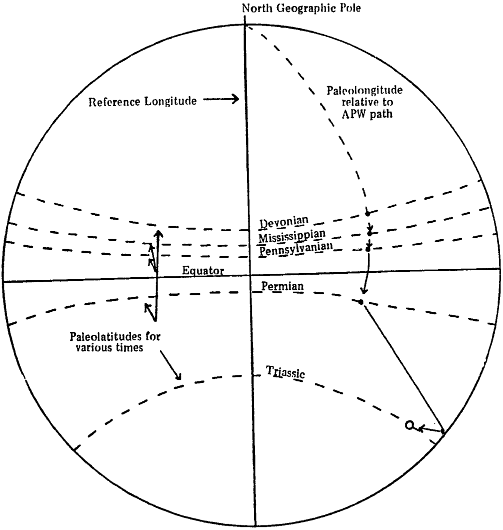


Figure 34 "Absolute" position of Alexander terrane through time relative to the reference paleolongitude. Reference longitude is the great circle Alexander terrane APW path rotated to conform to a line of longitude. Terrane positions are determined by the intersection of the paleolongitude, relative to the reference longitude (determined in figure 33) and the paleomagnetic paleolatitude. Terrane velocity may be calculated by geographic displacement divided by time.

a southeasterly motion from Permian to Triassic.

It must be pointed out that the preceding reconstruction of absolute positions and terrane velocities depends critically on the APW path being a great circle. Since the Alexander terrane APW path is not well established, the calculated velocities must be regarded as tentative.

Applications to Southern Alaska Tectonics

The Polarity Ambiguity

One of the serious problems that has hampered the construction of satisfactory paleomagnetic paleolatitude versus time trajectories for allochthonous terranes is the polarity ambiguity. In the absence of known polarities or complete APW paths, it is not commonly possible to assign an unambiguous polarity, and therefore the hemisphere of a paleomagnetic determination. These criteria include the least latitudinal translation required (Hillhouse, 1977, Hillhouse and Gromme, 1980, Stone and Packer, 1979, Van der Voo and others, 1980), the least amount of tectonic rotation required to account for the discordance of VGP's (Hillhouse, 1977, Van der Voo and others, 1980, Panuska and Stone, 1981) and the least number of required changes in the sense of latitudinal translation (Stone and others, 1982). The least total tectonic transport criterion is not by itself a very strong argument, especially if excessively rapid plate motions are not required. The least angular VGP discordance is a reasonable criterion; however, given the allochthonous nature of the terranes and their equivocal paleogeographic affinities, selection of a reference pole is often debateable. Moreover, the effects of local tectonic rotation, arc segmentation and oroclinal bending may be difficult to separate with a limited data base and may lead to erroneous conclusions. The least number of changes in latitudinal motion or APW path is perhaps one of the best criteria for polarity assignment. Continental APW paths tend to show an "orderly" progression of pole positions with relatively few "kinks" or accelerations. When such kinks or accelerations are observed, they can often be attributed to major tectonic events such as orogenies or opening of ocean basins (e.g., see Irving, 1979). On the basis of a limited number of rifting and orogenic events preserved in the record of a tectonostratigraphic terrane, the smoothest paleolatitude versus time or APW curve appears to be an acceptable guide to interpretation. Unfortunately, this presumes that terranes behave similarly to continents; the behavioral similarity is unknown at present.

Figure 35 represents an interpretive paleolatitude versus time trajectory for the Wrangellia and Alexander terranes as well as the other terranes comprising the Southern Alaska superterrane as they joined the amalgam. Several constraints and guidelines were employed to assign polarity. The Paleozoic Alexander terrane data were assigned polarities on the basis of the general similarity to the North American APW path, the known polarity of the Pybus Formation data and the gross stratigraphic similarity to the western United States, as explained in the previous chapter. A southern hemisphere Triassic paleolatitude for the Alexander terrane is inferred on the basis of the smoothest APW path, the least tectonic translation from Permian to Triassic time and on paleobiogeographic considerations (see previous chapter). The Paleozoic polarity for the Wrangellia terrane is known to be northern hemisphere as the sampled units were deposited during the Permo-Carboniferous reversed polarity interval. Triassic Wrangellia paleolatitudes are almost certainly northern hemisphere by correlation with the known polarity late Paleozoic VGPs along with supportive paleoboigeographic evidence. Since the geologic evidence strongly suggests that the

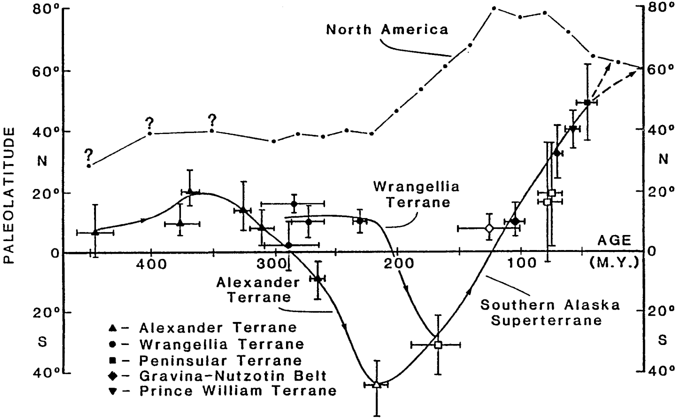


Figure 35 Preferred paleolatitude versus time curve for Wrangellia, the Alexander terrane and the Southern Alaska superterrane. Solid symbols indicate that polarity (thus hemisphere) is known with some confidence. Open symbols indicate equivocal polarity. The uppermost curve is a summary of paleolatitudes that the Southern Alaska superterrane would have recorded if it were part of North America since the Paleozoic. Data from: Hillhouse and Gromme (1980), Hillhouse (1977), Plumley and others (1983), and this study. See text for details of data.

Southern Alaska superterrane was assembled by Late Jurassic time, the Jurassic data from the Peninsular terrane (Stone and Packer, 1979), which includes samples of Middle and Late Jurassic age, may be applied to all terranes. A southern hemisphere position is inferred because of minimum motion considerations.

The data also show that Wrangellia and the Alexander terrane are clearly latitudinally separated in the Triassic. As geological evidence requires that the terranes be joined by Late Jurassic time, the Jurassic paleolatitudes almost certainly must be the same polarity as the Triassic Hound Island data from the Alexander terrane. The alternative to this (i.e., a southern polarity for the Alexander in Triassic and northern polarity for all terranes in Jurassic) would require a northward component of plate motion of at least 16 cm/yr, which seems rather high [The fastest plate motion today is on the order of 12 cm/yr (Addicott and Richards, 1982) and the fastest known plate motion is 13 cm/yr half spreading rate (Norton and Sclater, 1979)]. However, if the Alexander Triassic paleolatitude is challenged and a northern hemisphere location is preferred, then the least required motion argument would constrain the Jurassic data to be northern hemisphere polarity as well. The result of interpreting all Triassic and Jurassic data to be northern hemisphere is to generate a very "jerky" terrane motion (figure 36). Such a latitude versus time curve would require the Alexander terrane to reverse its latitudinal direction in the Permian (from southward to northward), again in Triassic (from northward to southward) and finally in Late Jurassic/Early Cretaceous (from southward to northward). Similarly, Wrangellia would have an

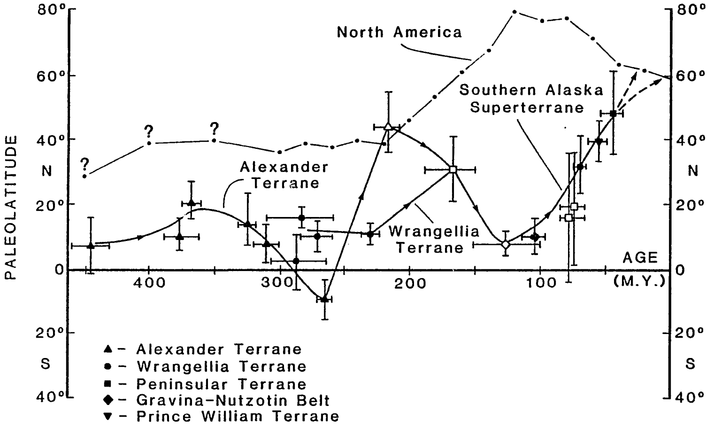


Figure 36 Alternative paleolatitude versus time curve for Wrangellia, the Alexander terrane and the Southern Alaska superterrane. This interpretation is not favored because of the "jerky" paleolatitude trends and the relatively frequent changes in the sense of paleolatitude motion (i.e. northward or southward). Data sources, symbols and North America reference curve are described in figure 35.

additional reversal in motion: namely a change in Jurassic time from northward to southward. The latitude versus time plot shown in figure 35 is favored because it requires only two changes in the sense of latitudinal motion changes (once for Alexander and once for Wrangellia) as opposed to the four changes required by the polarity interpretations illustrated in figure 36.

On the basis of the latitudinal motion changes just outlined, the APW path arguments and the biogeographic arguments outlined in the previous chapter, the latitude versus time trajectory in figure 35 is considered to be the most likely path of southern Alaska terranes. Based on this curve, the following tectonic model is put forward. Given the uncertainties of some of the paleomagnetic data and the circumstantial evidence for early Mesozoic polarity, this model can only be considered a working hypothesis.

Paleogeographic Model

The oldest known rocks (early Paleozoic) of the Southern Alaska superterrane occur within the Alexander terrane. The Alexander terrane probably originated as a continental volcanic arc adjacent to what is now California (Jones and others, 1972, Van der Voo and others, 1980). It is worth noting that before the southerly excursion, the Alexander terrane was at a paleolatitude that would allow it to participate in the Antler orogeny of the western United States. The concept that the collision of this terrane with North America could be the driving force of the orogeny needs closer scrutiny.

Based on paleomagnetic evidence, the Alexander terrane began to move in a southerly direction in Devonian to Mississippian time. It continued to move southward until it arrived at a latitude of 44 degrees south, in the Late Triassic. In addition, by assuming an origin of the Alexander terrane in Nevada, "absolute" motion considerations (figure 34) suggest that Alexander moved southward and then southeastward towards the general vicinity of the west coast of South America. This scenario allows an explanation of the South American affinities of Alexander terrane Septocardia (Newton, 1983) and Monotis (Jones and Silberling, 1983), by suggesting close Triassic geographic proximity.

During Pennsylvanian through Triassic time, Wrangellia remained at approximately constant paleolatitude. By comparison with North American reference poles (Irving, 1979), Wrangellia's paleolatitude was equivalent to the latitude of present day Las Vegas throughout the late Paleozoic and early Mesozoic, although the error limits on the paleomagnetic data permit any latitude between present day Oregon and central Mexico. Faunal evidence (Silberling and Jones, 1983, Newton, 1983) suggests an eastern Pacific paleogeography due to similarities with fauna from Triassic cratonic rocks in Nevada. Thus, it is tempting to speculate that Wrangellia was involved in the Late Permian-Early Traissic Sonoma orogeny. Jones and others (1977) suggested that the voluminous Nikolai Greenstone and equivalent rocks represented a Late Triassic rifting event. This interpretation is supported by sparse trace element data from the Goon Dip Greenstone (Nikolai equivalent) reported by Decker (1983). Accordingly, the Nikolai Greenstone couuld be related to a post-collisional (Sonoma orogeny) rifting event that severed Wrangellia from the North American margin. Following rifting, Wrangellia moved southward into the southern paleo-Pacific ocean.

By Late Jurassic, the Wrangellia and the Alexander terranes amalgamated in low to middle latitudes of the southern hemisphere. The amalgamation event was probably an arc-arc collision corresponding to the Late Jurassic orogeny in the Wrangell Mountains recognized by Mackevett (1978). The newly formed Southern Alaska superterrane then moved northward throughout the remainder of the Mesozoic. By latest Cretaceous time, the superterrane was at middle northern hemisphere paleolatitudes and by Paleocene the superterrane was located at approximately 40 degrees N. latitude (Plumley and others, 1983). The Southern Alaska superterrane then accreted to North America by closing a Tertiary ocean basin lying between it and the continent, intracontinental shortening, transcurrent faulting or a combination of these scenarios, as outlined by Moore and others (1983).

The Arrival Problem

Each of the accretion scenarios for southern Alaska terranes discussed by Moore and others (1983) has difficulty in accounting for various geologic relationships. If the Southern Alaska superterrane accreted by closing a Tertiary ocean basin, all evidence for this ocean basin must have been lost (or has not yet been discovered). Because of this lack of marine Tertiary rocks between the superterrane and the continent, Jones and others (1982b) favored a Late Cretaceous accretion age. Moore and others (1983) suggest the possibility that evidence for the "missing" Tertiary ocean basin could lie beneath the major thrust faults in the Alaska Range and northern Talkeetna Mountains. The currently recognized total strike-slip displacement, inboard of the Southern Alaska superterrane, is perhaps 1000 km (Moore and others, 1983). This value is considerably lower than the amount of convergence required by the paleomagnetic data. Some of the predicted convergence could be taken up by crustal shortening (Moore and others, 1983); however, the relative importance of this process is very difficult to assess at present. The major controversy concerning the late Mesozoic and Cenozoic southern Alaska tectonic history revolves around two key issues: when were the southern Alaska terranes emplaced and where is (are) the structure(s) that accommodated the convergence between these terranes and those parts of Alaska already in place? Paleomagnetic data provide important contraints for both issues.

The MacColl Ridge Formation data presented in this dissertation preclude a pre-Cenozoic emplacement age for Wrangellia and by implication, for the entire Southern Alaska superterrane. Early Tertiary paleolatitudes determined from the Alaska Peninsula, Kodiak Island and the Matanuska Valley (Stone and Packer, 1979, Plumley and others, 1983 and Stone, 1983) suggest that the Southern Alaska superterrane did not reach its present position until post-Eocene time. This scenario would allow southern Alaska to be the source of terrigenous sediments in the Zodiac Fan, presently on the Pacific Plate immediately south of the Aleutian Trench (Stevenson and others, 1983). In contrast, paleomagnetic data from interior Alaska localities, north of the Denali fault system and western Alaska localities, suggest little or no motion relative to North America since latest Cretaceous or early Tertiary.

Globerman and others (1983) sampled Lower Cretaceous tuffs near Ohagamuit, on the lower Yukon River, and determined that these rocks had been remagnetized, presumably in later Cretaceous or Paleocene time. Because the mean vector direction without tectonic tilt correction is not appreciably different from the expected Late Cretaceous or Paleocene field direction (assuming a present relative paleogeography), they concluded that these rocks were part of North America by the Late Cretaceous or Paleocene. Globerman and Coe (1983) found paleolatitudes consistent with a North American paleogeography in latest Cretaceous time for the Hagemeister Island area, Bristol Bay. Plumley and Coe (1982) determined a Paleocene VGP position from the Nowitna Volcanics near McGrath, which agrees with the North America Paleocene pole. Similarly, Hillhouse and Gromme (1982) present paleomagnetic data from Paleocene volcanic rocks within the Cantwell Formation in the McKinley Park area, north of the McKinley strand of the Denali fault, which strongly supports a North American paleogeography. These data pass the fold test and yield a VGP very close to Cretaceous and Paleocene North American paleomagnetic poles. The available paleomagnetic data suggests that the Southern Alaska superterrane was emplaced against portions of Alaska north and west of the Denali fault system, which were part of the North American continent by latest Cretaceous to Paleocene time.

Recent evidence from Eocene volcanic rocks (50-54 m.y., K-Ar ages) in the northern Talkeetna Mountains shows a paleolatitude predicted by its present position with respect to North America (Hillhouse and Gromme, 1983). These rocks lie south of the Denali fault, occur on both sides of the Talkeetna thrust (which juxtaposed Jurassic-Cretaceous flysch and Permian volcanic rocks of Wrangellia) and are considered to depositionally overlie Wrangellia (Hillhouse and Gromme, 1983). Assuming that the primary nature of the remanence and the correct terrane affinities can be verified in a more complete publication, these data have very important implications. Taken at face value, the MacColl Ridge data (this dissertation) and the Eocene volcanic data indicate that Wrangellia moved from 32 degrees latitude to 76 degrees latitude or a minimum of 4800 km within approximately 20 m.y. This displacement would require plate velocities of at least 22 cm/yr, which would rival or surpass the fastest known plate motions (Norton and Sclater, 1979, determined a 13 cm/yr half spreading rate in the Indian Ocean based on offset of Cretaceous marine magnetic anomalies). As an alternative to the rapid motion hypothesis, the data may be explained by postulating a tectonic boundary between the Eocene locality of Hillhouse and Gromme in the Talkeetna Mouuntains and the MacColl Ridge locality in Wrangellia. Such a boundary has not yet been recognized, but the possibilty exists that new tectonic discontinuities will be identified. To resolve this problem, it is imperative that more sampling be carried out across the Talkeetna Mountains in rocks of Cretaceous and Tertiary age.

Summary and Conclusions

The following is a brief statement of the major conclusions and findings that are a direct result of my work.

Approximately 800 paleomagnetic samples were collected from 24 localities in southcentral and southeast Alaska in order to elucidate the latitudinal translational history of Southern Alaska tectonostratigraphic terranes. Thirteen localities failed to yield resolvable characteristic directions or displayed strong evidence of remagnetization. The remaining 11 localities passed all the stability tests that could be applied and were used in the tectonic analysis.

Paleomagnetic data from the MacColl Ridge Formation, which passes both the fold test and the reversal test, demonstrates that the Southern Alaska superterrane was located 32 degrees N latitude in Maestrichtian time (latest Cretaceous). This precludes the possibility of a pre-Tertiary arrival of the terrane at its present location. The MacColl Ridge Formation paleolatitude is similar to the paleolatitude obtained for the Ghost Rocks Formation, Prince William terrane (Plumley and others, 1983) and is permissive of the interpretation that the Chugach and Prince William terranes were deposited as integral parts of the Southern Alaska superterrane. Relatively minor dispersion of Cretaceous, Triassic and late Paleozoic VGPs suggests that the Wrangell Mountains have sustained very little internal differential rotation and have rotated about 130 degrees counterclockwise as a coherent block.

Analysis of regional mapping provides evidence supporting a Late

Jurassic, amalgamation of the Wrangellia and Alexander terranes. The geologic evidence suggests that by Late Jurassic the Peninsular and Gravina-Nutzotin terranes were also linked to the Wrangellia and Alexander terranes. This composite terrane is here termed the Southern Alaska superterrane. The Chugach and Prince William terranes may also be part of this superterrane.

Paleomagnetic data from the Brothers Volcanics, which pass an intraformational conglomerate test, indicate an 8 degree (probably northern hemisphere) paleolatitude in Late Jurassic-Early Cretaceous time. These data plus the MacColl Ridge data strongly suggest that the Alexander terrane had not been emplaced at its present position relative to North America by Late Triassic time, as proposed by Hillhouse and Gromme (1980).

Data from the upper Paleozoic Skolai Group has apparently resolved the polarity ambiguity, indicating that Wrangellia occupied a 15 degree N latitude in late Paleozoic and Triassic time.

Data from the Permian Pybus Formation yield a 9 degree S paleolatitude determination for the Alexander terrane. If the primary nature of the Pybus remanence direction can be verified by subsequent studies, important ramifications follow. This will confirm the early and mid Paleozoic southward plate motion. The most likely APW path and paleolatitude trends favor a Triassic southern hemisphere mid-latitude position for the Alexander terrane.

Analysis of the available paleomagnetic data plus faunal evidence allows the following working hypothesis: The Alexander terrane began to move southward in Devonian or Mississippian time to attain a mid-southern hemisphere paleolatitude near the west coast of South America in Late Triassic time. Wrangellia began to move southward in late Triassic time. By Late Jurassic, the Wrangellia and Alexander terranes amalgamated in the southern hemisphere. The assembled Southern Alaska superterrane then moved northward until it accreted to North America in Tertiary time.

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