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True polar wander during the Permo–Triassic

Maria C. Marcano, Rob Van der Voo*, Conall Mac Niocaill

Department of Geological Sciences, University of Michigan, Ann Arbor, MI 48109-1063, U.S.A.

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As usual, and this should be obvious to anyone whose eyes have not been blinded by the false light of Western science, the pole that we see is not the real pole, for the real pole is the one that cannot be seen, except by some adepts, whose lips are sealed.

Umberto Eco, Foucault's Pendulum

Abstract

The apparent polar wander path for the Pangea supercontinent is about 35° long for the interval of 295–205 Ma, which means that in that interval Pangea rotated over an angle of 35° with respect to the rotation axis about an Euler pole located at the equator. If the rest of the world, largely comprised of the Panthalassa Ocean, rotated about the same Euler pole in the same sense as Pangea, then a good case can be made that true polar wander occurred during the Permo–Triassic. In contrast, if the Panthalassa Ocean moved in a sense opposite to that of Pangea, then true polar wander is not likely to have occured. In the latter case, convergence between Panthalassa and Pangea would have led to subduction of large amounts of ocean crust under the leading edge of Pangea. We have examined the geology of Pangea's leading edge for evidence of such subduction, in the form of Permo–Triassic plutonism, andesitic volcanism and deformation. No such evidence was generally found, except for areas very close to the Euler pole in the western U.S.A. and in displaced terranes that also were in near-equatorial paleolatitudes at the time. We conclude that true polar wander has most likely occurred during the Permo–Triassic at a rate of about 0.4° per million years.

1. Introduction

True Polar Wander (TPW) is the real displacement of an entire, rigid Earth with respect to its spin axis (Goldreich and Toomre, 1969; Gordon, 1987). This simple definition becomes a complex one as soon as we try to describe such displacement in a terrestrial reference frame because, of course,

* Corresponding author: Fax: +1-313-763-4690; e-mail: voo@umich.edu.

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the Earth is not rigid. If it were, TPW would merely be a physical consequence of the moment of inertia tensor of a spinning heterogeneous body. However, the mantle and core of the Earth are convecting, and the Earth's lithospheric plates are all in relative motion with respect to each other as well as with respect to the mantle. Thus, kinematic relations among the different layers of the Earth and their inherent components, which are still not well understood, require that we make certain assumptions and simplifications before TPW during geological times can be approximated.

There are three reference frames which are meaningful in a discussion of TPW: the Earth's rotation axis, the mantle hotspot framework and the mean-lithosphere framework. A disadvantage of the spin axis framework is that it is not directly recorded for geological time scales; we must therefore seek a proxy in the mean paleomagnetic pole for a given time period. For the Phanerozoic, this approximation is reasonably well-supported by the general agreement between paleomagnetic poles and independent latitudinal indicators, such as those related to paleoclimate (e.g. Van der Voo, 1993). In the mean-lithosphere framework, individual motions of the plates have no net rotation in common. In other words, TPW is the movement of the rotation axis in this framework and is directly related to the theoretical concepts of the net angular momentum of, or net torque on, the lithosphere (Simpson, 1975; Solomon et al., 1977). The drawback of the mean-lithosphere framework is that it relies on precise knowledge of the relative movements between all plates, which is difficult to obtain for converging plates, such as North America and the Pacific (see Acton and Gordon, 1994). The mean-lithosphere framework may also not be representative of the Earth as a whole. The hotspot framework has its disadvantages as well, one being that for pre-Cretaceous time the record is very scarce and another being the increasing suspicion that individual hotspots are not truly stationary with respect to each other (Molnar and Stock, 1987; Acton and Gordon, 1994; Christensen, 1998).

Two methods have been proposed that do not rely on a hotspot framework. One of the simplest approaches to assess TPW is the one proposed by McKenzie (1972) and subsequently calculated by McElhinny (1973) in which a common movement should show up in a resultant vector after summing the surface area-weighted displacement vectors for all plates, given that relative motions between plates are canceled out in this way. This method resulted in an estimate of TPW over the past 50 Myrs that was not significant. The other technique (Jurdy and Van der Voo, 1974; Jurdy and Van der Voo, 1975) involves a mathematical method for separating TPW from purely relative plate motions and yielded a result for the Tertiary consistent with McElhinny's. Moreover, the cumulative amount of TPW since the Early Cretaceous was also found to be insignificant, because it was less than the radius of the cone of 95% confidence about the mean pole (about 7°).

The use of the hotspot reference frame contrasts with the above two methods, equating TPW with the displacement of a plate's mean paleopole while keeping the hotspots fixed (e.g. Livermore et al., 1983; Livermore et al., 1984; Andrews, 1985). The basic assumption is that hotspot traces reflect plate movements over the stationary deeper mantle. By correcting the positions of the plates and their paleopoles for these movements, the displacement of the spin (= paleomagnetic) axis, can be calculated with respect to the mantle hotspots.

Consequently, any such displacement is just showing a non-coincidence of the present and past spin axes in this framework. Earlier results revealed up to 30° of total offset since the Early Cretaceous, being small for the Cenozoic and much larger for times prior to about 65 Ma. More recent results using this approach (Courtillot and Besse, 1987: Besse and Courtillot, 1991) for the past 200 Myrs yield around 10–20° of maximum shift, with the greater amount of TPW during the Mesozoic. Courtillot and Besse also recognized periods of rapid TPW and periods of standstill.

Comparisons between the paleomagnetic and hotspot frameworks have resulted in decreasing estimates of TPW, as better data have become available (Gordon and Livermore, 1987; Besse and Courtillot, 1991). This suggests that in part the older results might be attributable to inaccurate determinations for the fixed-hotspot model. Because it is increasingly likely that mantle hotspots are not stationary (Christensen, 1998), it remains advantageous to obtain a result that is independent of this framework, i.e. based only on mean-lithosphere and paleomagnetic axis comparisons. However, given that the mean-lithosphere framework relies on accurate knowledge about the uncertain relative plate motions in the Pacific–Antarctica–Africa circuit, such TPW determinations have been model-dependent and somewhat suspect.

For Jurassic and older times determinations of TPW depend entirely on comparisons between the lithosphere and the records of the paleomagnetic axis, i.e. apparent polar wander paths (APWPs). Even though TPW is potentially a component of any APWP for a given period and is known to be a rotation about an equatorial pole in common to all plates (Jurdy and Van der Voo, 1974), it cannot be addressed in a straightforward way for pre-Cretaceous times unless further assumptions are made about the movements of all the lithospheric plates, particularly those oceanic plates which have since been subducted.

Because these assumptions are difficult to verify, interpretations of TPW in earlier times are very speculative. Van der Voo (1994) proposed that TPW could have been rather large during the middle Paleozoic (up to 75° over 75 Myrs). Similarly, Kirschvink et al. (1997) suggested that about 90° of TPW is incorporated in long APWPs for an Early Cambrian interval of some 15 million years; alternatively, in the absence of TPW, plate motions must have been rather high (>60 cm/year). Given that TPW must be a common element in a set of APWPs, these authors based their hypotheses on the similar shapes or lengths of the APWPs for the major continents for that time period. However, these proposals are of necessity very tentative, as well as contentious (e.g., Torsvik et al., 1998); moreover, no knowledge exists about the movement of oceanic plates in those times.

Anderson (1982) observed that the geoid highs apparently correlate better with former configurations of the continents than with today's positions, suggesting that certain gravity anomalies could be related to the position of Pangea during the late Paleozoic–early Mesozoic. Anderson points out that gravity anomalies, in turn, must be based on density anomalies that may influence the rotation of the Earth and suggests that TPW is likely while these geoid anomalies are developing. In the model of Kirschvink et al. (1997), it is argued that the maximum and intermediate non-hydrostatic principal moments of inertia exchanged positions, triggering 90° of TPW. In contrast, Besse and Courtillot (1991) established a link between TPW and the frequency of reversals of the geomagnetic field. According to their model the Triassic was a time of frequent reversals and possibly a time of rapid TPW.

These studies suggest that assessments of TPW before about 200 Ma are important, even if speculative. Because hotspot tracks are not available for this time, our aim in the following analysis is to assess the likelihood of TPW during the time that Pangea existed as a supercontinent, taking into account possible lithospheric plate movements, for which direction and magnitude may be indicated by the geological record.

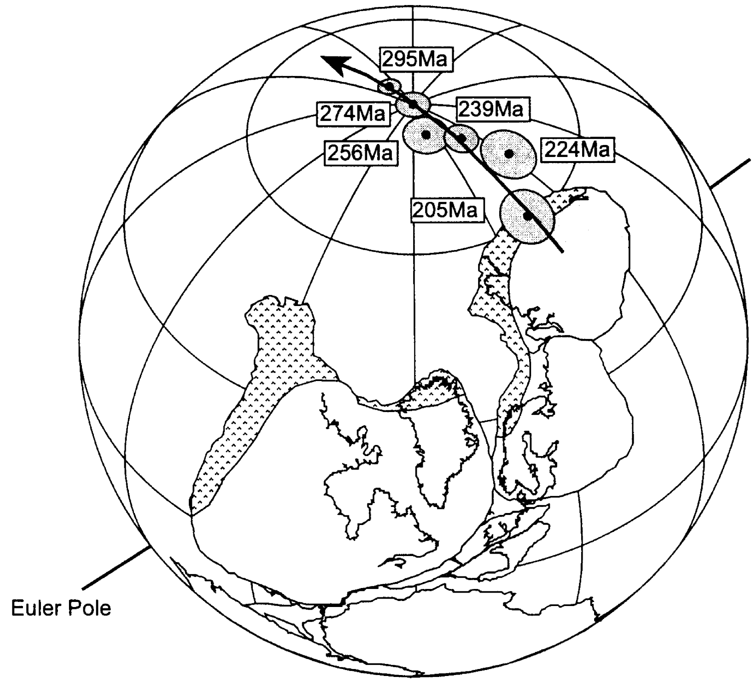

Fig. 1. Pangea and its apparent polar wander path 295–205 Ma, represented by the APWP derived for North America (Van der Voo, 1993), which indicates that a $\sim 35^{\circ}$ rotation of Pangea occurred with respect to the paleomagnetic axis about an equatorial Euler pole just west of California. The areas marked by a different signature at the leading edge of Pangea have been examined for geological signatures of Permo–Triassic subduction.

2. How TPW during the Permo–Triassic would be recognized

During the roughly 90 Myrs from Late Carboniferous to Early Jurassic all the major continents were assembled together in a quasi-rigid supercontinent, Pangea, that occupied about 70% of a hemisphere. It is generally assumed that this supercontinent constituted a single lithospheric plate, although in the next section we will discuss the implications of hypothesizing relative movements *within* the overall outline of the supercontinent. Most of the rest of the Earth's surface was a large oceanic domain, known as Panthalassa. The APWPs of the best-studied continents within Pangea, N. America and Europe, are smooth traces during this interval (Fig. 1), and represent the displacement of the paleomagnetic axis with respect to northern Pangea. The APWP of Gondwana forms a similar trace, which resembles the N. American APWP in gross outline, and which in a tight Pangea configuration falls close to the N. American APWP (Van der Voo, 1993; Fig. 5.32). Because the definition of TPW implies a rotation of the Earth about an equatorial axis, we will consider in our analysis the best-fitting great-circle segment through the APWP as the consequence of the displacement of Pangea with respect to the paleomagnetic axis, having been produced by a rotation about an Euler pole located in the equatorial plane.

For TPW not to occur, i.e. for the paleomagnetic axis to be stationary with respect to the lithosphere framework, compensatory plate movements are essential (Fig. 2). Reduced to extreme simplicity, this means that if one continental hemisphere (Pangea) moved north, the other oceanic

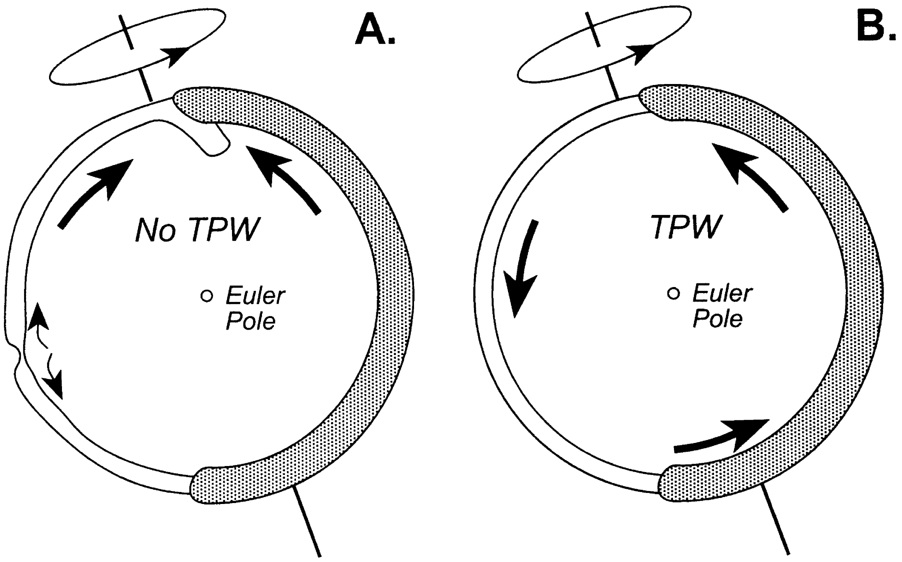

Fig. 2. Schematic illustrations of the hypothetical movements of a largely continental (Pangea, stippled) hemisphere and a largely oceanic (Panthalassa) hemisphere about an Euler pole perpendicular to the sections. In (A) the movements are compensatory and no true polar wander (TPW) would be likely, whereas in (B) the movements are in the same sense, implying the existence of TPW.

hemisphere (Panthalassa) must also have moved north in a rotation of equal magnitude, but opposite direction, about the same Euler pole (Fig. 2a). Conversely, if the continental hemisphere (Pangea) moved north and the oceanic hemisphere (Panthalassa) moved south in similar rotations about the same Euler pole, then the entire lithosphere shows a coherent motion with respect to the paleomagnetic axis [Fig. 2(b)]. The net result, as explained in the Introduction, could be taken to represent TPW.

From the length of the APWP trajectory (circa 35°) we calculate that the plate velocity of Pangea, with respect to the paleomagnetic axis, was about 4 cm/yr or less, depending on location. If Pangea moved north with respect to a stationary Panthalassa, about 50 million km^2 of oceanic crust must have been consumed or subducted at the leading edge of Pangea during the 90 Myr interval (Fig. 3). If Pangea and Panthalassa both moved north about the same Euler pole, albeit with opposite sense of rotation, we expect about 100 million km^2 of oceanic crust to have been subducted over this time interval. Clearly subduction of such copious amounts of oceanic crust would leave traces of an Andean-type margin along the leading edge of Pangea, or in marginal island arcs. In this latter scenario of Pangea–Panthalassa convergence no net rotation of the lithosphere would have occurred. However, if Panthalassa moved south with respect to the spin axis, then an overall net rotation of the lithosphere would have taken place, no evidence for subduction would be present at the leading edge of Pangea, and TPW could have been significant ($\sim 35°$ over 90 Myrs).

Therefore, we have examined specific geographic locations (Fig. 1) of the former leading edge of Pangea for possible evidence of the large subduction-related volcanism that would be required for no TPW to have occurred. In a later section, possible intra-oceanic sites of subduction will be

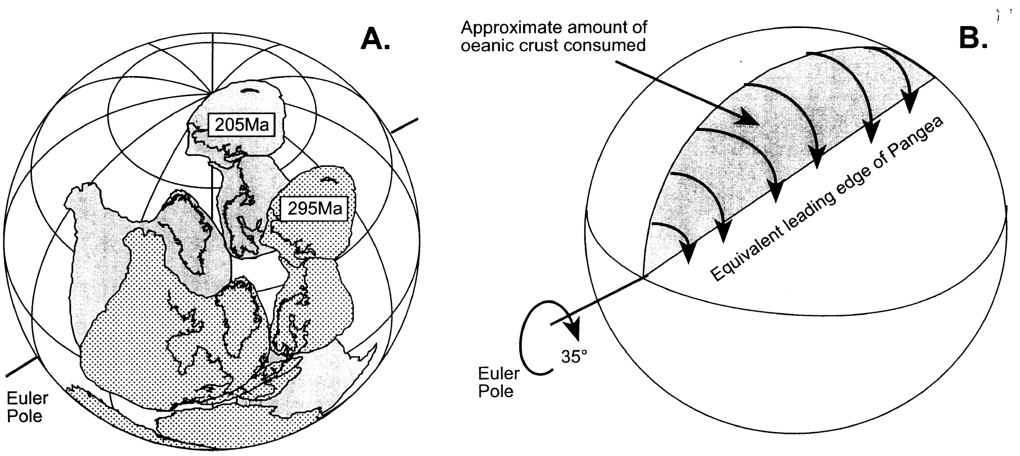

Fig. 3. (A) The positions of Pangea with respect to a paleomagnetic axis held fixed, at the beginning (295 Ma) and the end (205 Ma) of the interval considered. (B) The surface area of the oceanic (Panthalassa) crust and upper mantle that would be expected to have subducted if Pangea moved with respect to a stationary Panthalassa during the 295–205 Ma interval.

examined as well, assuming that we would find evidence for these as island arcs in far-traveled terranes now accreted to and incorporated into continental margins. While the paleopositions of such arcs are poorly known, if they existed at all, —and yet are necessary for assessing their relevance to TPW—we can nevertheless attempt to address how large these features may have been.

3. Implications resulting from relative movements within a quasi-rigid Pangea

In this section we will address the possible consequences for our TPW analysis resulting from Permo–Triassic relative movements between the major continents within Pangea, such as between Siberia and Baltica (= northern and eastern Europe), or between Gondwana and northern Pangea, as well as relative movements of smaller blocks, such as Iran, Tibet, or North and South China.

It is well-known that Siberia, Kazakhstan and Baltica collided to form the Ural mountains, but the details of their relative movements before final amalgamation in the Permian are not well known. However, the Permo–Triassic movements of Siberia/Kazakhstan with respect to Baltica were sinistrally transpressive before final amalgation (Echtler et al., 1997) and this implies that, if anything, we underestimate the total (averaged) northward movement of Pangea, because the APWP we use is for Laurentia and Europe. If the relative movements of Siberia and Kazakhstan were still significant during the interval of interest, it would mean that Pangea as a whole had a greater (averaged) northward movement than now represented by the APWP in Fig. 1.

Even less certain is the magnitude of possible north–south movements between Gondwana and northern Pangea during this interval. However, even when a relative shift from a Pangea B to a Pangea A configuration is granted (for discussion, see Van der Voo, 1993; ch. 5, or Torcq et al., 1998), it did not involve major north–south, but rather an east–west movement during the Permian or Triassic. The combined effect of the Pangea B–Pangea A transition may involve a reduction of at most some 5° in track length for the (averaged) Pangea APWP.

Likewise, the relative movements of blocks within the Tethys, have implications for the overall movement of the Pangean 'plate', but here the effects would be to enhance the (averaged) northward movement of Pangea, as all components, such as Iran, N. Tibet, N. China and S. China moved northwards during the Permo–Triassic interval (Van der Voo, 1993; Torcq et al., 1998).

In summary of this section, even if we grant that Pangea was only quasi-rigid and that internal displacements between its components took place, we estimate that the aggregate effect of these movements would not greatly reduce the (averaged) northward movement of Pangea, and therefore would not lead to very different estimates of TPW. In our analysis we will therefore ignore the effects of relative movements internal to Pangea.

4. Subduction-related features of Pangea's leading edge

The leading edge of the northward moving Pangea Supercontinent (Fig. 1) consists of the cratonic margins of eastern and northern Siberia, northern Europe (Baltica), northern Greenland and the northern and western margins of North America (Laurentia). Below we will examine whether these margins preserve evidence of oceanic convergence and subduction during the 295–205 Ma interval. Mesozoic and younger far-traveled terranes, accreted to these margins after 205 Ma will be discussed in a later section.

4.1. Northeastern Siberia

The modern northeastern margin of the Siberian Platform constitutes an important component of the leading edge of Pangea (Fig. 1). Toward the east of this margin a number of accreted terranes of cratonic and oceanic affinity are found that were docked to the Platform primarily during the middle to late Mesozoic (Churkin et al., 1985; Khramov and Ustritsky, 1990), i.e., after the periods of interest.

Siberia consists of a Precambrian–Paleozoic nucleus (the Anabar–Olenek and Aldan blocks of Fig. 4), surrounded by a deformed margin, which on the northeastern and northern sides comprises the Verkhoyansk Fold Belt and the Taimyr Peninsula. The Carboniferous to middle Jurassic rock sequences of these margins comprise clastic sediments deposited on a continental passive margin (Parfenov, 1991). Sedimentary environments from Carboniferous to Permian changed by northward progradation from typical deposits of lower fan and basin plain to widespread shallow water deltaic sequences (Khudoley and Guriev, 1994). No evidence is found in the rocks of Carboniferous through Early Jurassic age for subduction in this area.

Separated from the Verkhoyansk Fold Belt by the Adycha–Taryn Fault Zone (ATFZ in Fig. 4), slates of Triassic and Jurassic age are found in the Kular–Nera Belt, which borders the Omolon Massif and Platform. The low-grade metamorphic sediments of the Kular–Nera Belt are interpreted as continental rise deposits, subsequently tectonized during accretionary wedge formation in a collision between Siberia and Omolon (Parfenov et al., 1993), with the ATFZ as the collisional suture (Parfenov et al., 1988). Because paleomagnetic data indicate that the paleolocation of the Omolon Massif was at least 30° further southward than today's position with respect to Siberia

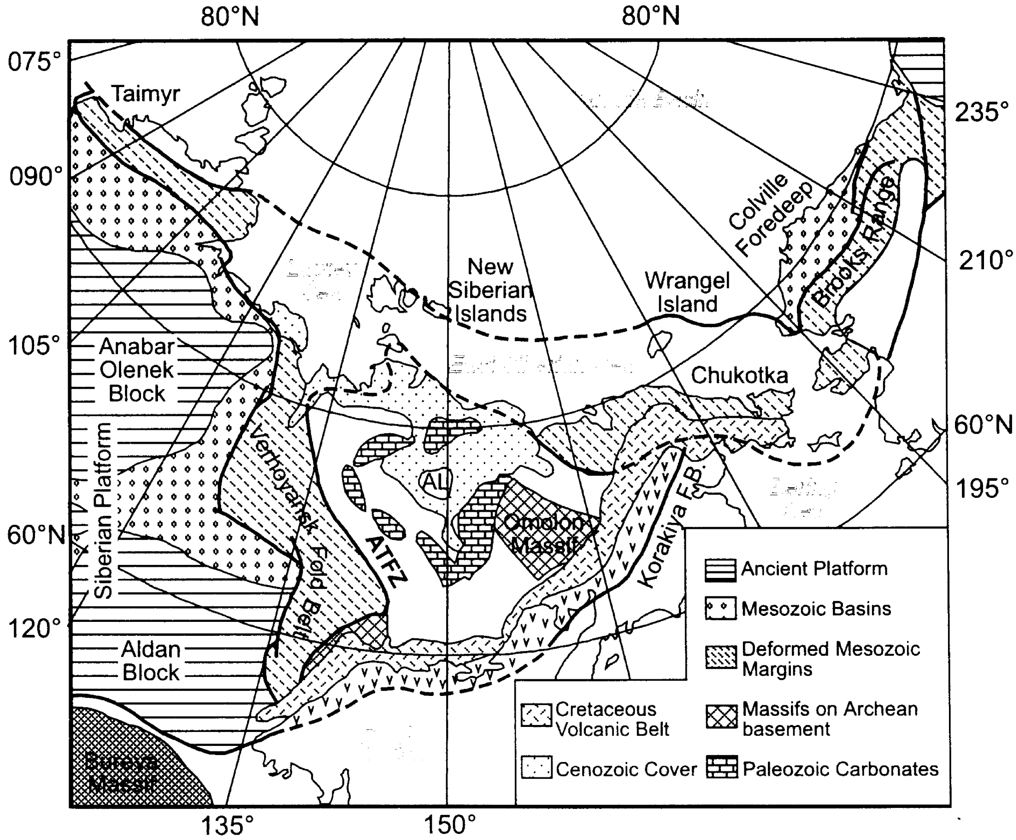

Fig. 4. Schematic geological map of Northeastern Siberia (for discussion, see text). ATFZ = Adycha–Taryn Fault Zone. The areas where Permo–Triassic island-arc activities have been documented are AL = Alazeya and the limited Udsko–Murgal'sk arc remnants in the Cretaceous Okhotsk–Chukotka coastal arc complex (v-pattern).

(Khramov and Ustritsky, 1990), this collectively implies that the Kular–Nera Belt and blocks east of there are far-traveled terranes. We will therefore discuss the evidence for Permian–Triassic subduction-related activity to the east of the ATFZ and the possible paleolocations of these terranes in a later section.

In the southern part of the Taimyr Peninsula, Permian granitic intrusions are found that could be related to subduction in late Paleozoic times. However, it seems most likely that these granites are related to closure of the Uralian Ocean between the Russian Platform and Siberia and not to subduction of Panthalassa under Siberia, because this part of the Taimyr Peninsula forms a direct northeastward continuation of the Urals (Hamilton, 1970).

4.2. Northern Europe

Almost all of the northern margin of Baltica is comprised of rocks of Paleozoic and Proterozoic age, representing the Proterozoic basement of the paleocontinent and the overlying allochthonous sheets that were thrust onto the Baltic margin during the Siluro–Devonian Caledonian orogeny.

Several units in the upper allochthon also yield evidence for an Early Ordovician (Finnmarkian) phase of deformation (e.g. Sturt and Roberts, 1991; Roberts, 1988). However, since the Caledonian orogeny no major phases of deformation have affected any of the rock units on the northern mainland of Scandinavia and no rocks of Permo–Triassic age are present.

In contrast, rocks of Permo–Triassic age are preserved in southern Svalbard, with up to 1600 m of evaporites, carbonates and clastics present (Cutbill and Challinor, 1965; Birkenmajer, 1981; Hjelle et al., 1986). These units are all inferred to have been deposited on a stable platform (Birkenmajer, 1981) and there is no evidence of any major volcanism or deformation during the time period of interest. Similarly, drill cores from the southern Barents Sea, south of Svalbard and north of Finnmark, have also revealed the presence of several hundreds of metres of carbonates, clastics and evaporites, which are all interpreted to have been deposited on a stable shelf (Bugge et al., 1995).

4.3. The Canadian Arctic and Greenland

Although rocks as old as Precambrian outcrop in the area, rocks of Paleozoic and Mesozoic age are predominant in Arctic Canada (the Franklinian and Sverdrup basins) and the northern Greenland fold belt. The older Paleozoic sequence was deformed during the Devono–Carboniferous Ellesmerian Orogeny, whereas the post-Devonian sediments were deformed during the Cretaceous and Paleogene (Sweeney et al., 1994). The succession of events that gave rise to the deposition of the rocks is similar for both areas, despite the fossil plate boundary nature of the Nares Strait, which separates northern Greenland from the Canadian arctic islands (Dawes and Kerr, 1982).

The Ellesmerian Orogeny, which affected Siluro–Devonian deposits in the Late Devonian–Early Carboniferous, is characterized by intrusions, metamorphism, regional folding and faulting. Neither its timing nor its relationship to plate tectonic interactions, e.g., between Asia and North America, are well understood. At any rate, it is well established that this orogeny occurred before the Late Carboniferous and younger periods of interest to our analysis.

The rocks deformed in the Ellesmerian orogeny are unconformably overlain by Early Carboniferous to uppermost Cretaceous sequences that outcrop primarily in Canada (e.g., the Sverdrup Basin). These late Paleozoic–Mesozoic strata include marine and non-marine terrigenous clastic sediments, carbonates and evaporites. Locally, gabbro dikes and sills and basalt flows are found, and are inferred to be mainly of Cretaceous age, although minor Permo–Carboniferous basaltic eruptives are reported in the Audhild and Esayoo formations (Balkwill, 1978; Balkwill, 1983; Jackson and Halls, 1988). The latter volcanics in the post-Ellesmerian section in the Sverdrup Basin may be related to the formation of grabens, initiated by extension of Carboniferous to Early Permian age. Transgression was followed by the progressive filling of the new depression by clastic sediments. The cycle ends with the deposition of mature to supermature quartzarenites. Evaporitic diapirism developed during the late Middle to early Late Triassic. The Sverdrup basin subsidence ended in the Campanian to Maastrichtian with the Eurekan orogeny.

4.4. The cratonic margin of Laurentia in Alaska and western Canada

It has long been recognized that nearly all of Alaska, and much of the Intermontane, Coast Mountains and Insular belts of Canada (for definition, see Cowan et al., 1997) are additions to the North American continent that occurred in post-Triassic times (e.g., Irving et al., 1996). A conservatively drawn Permo–Triassic North American margin is shown (Fig. 5, from Burchfiel et al., 1992) as located in southern British Columbia, to the west of what is often referred to as the Quesnellia 'terrane' (e.g., Monger et al., 1994), continuing southward into Washington, but many authors place this margin more to the east (Speed, 1994; Monger et al., 1994). The displaced and/or accreted areas to the west of the dashed line in Fig. 5 will be discussed in a later section.

During the interval of interest, i.e., from the Late Carboniferous through the Triassic, the continental margin in northwestern Canada (Yukon) was generally passive, with no known deformation and very little igneous activity. To the south, in British Columbia, the situation is different in the displaced terranes to the west of the Quesnellia 'terrane', and in Quesnellia itself. In the Slide Mountain area to the east of Quesnellia, and in Quesnellia itself, mid-Carboniferous to Permo–Triassic pillow basalts, gabbros, diorites and granites are found and are thought to be arc-related (Monger et al., 1994). On the other hand, volcanics in the Nicola Arc (in the Eastern Assemblage of Monger, 1977; Miller et al., 1992) of southern British Columbia (Fig. 5) are thought to be the remains of late Paleozoic rift basins (Struik and Orchard, 1985). Whether the arc-related rocks are indicative of subduction underneath the North American craton near British Columbia clearly depends on whether Quesnellia is autochthonous or whether it is a displaced terrane. Arguments for a southern-hemisphere Permian paleolatitude have been presented by Ross and Ross (1983) on the basis of the terrane's McCloud fauna, which is intermediate to the typical Tethyan and North American faunas; unfortunately, no paleomagnetic evidence exists for the late Paleozoic or early Mesozoic paleolatitudes of these areas.

4.5. The cratonic margin of Laurentia in the western U.S.A.

Figure 5 illustrates the main features that are of interest to our analysis, with the largest being the Golconda Allochthon in Nevada, which was emplaced during the Early Triassic Sonoma Orogeny above the older Roberts Mountains allochthon. During this orogeny, deep-marine sedimentary rocks of latest Devonian through Permian age and associated volcanics were emplaced (Speed, 1977). The Roberts Mountains thrust system, in turn, is related to the earlier Antler Orogeny that ended by the middle Mississippian (Miller et al., 1992). In southern California, the rocks in some areas (e.g., El Paso) were intensely deformed and metamorphosed in the Permian and intruded by Upper Permian granitoids (Burchfiel et al., 1992). In contrast to the areas discussed directly below, which often have an uncertain paleolocation, the El Paso area and the Golconda Allochthon unambiguously were, or became, part of the cratonic margin of North America during the Triassic.

It is debatable whether the Blue Mountains, Klamath Mountains and northern Sierra Nevada (Fig. 5) were in the same North American location during the late Paleozoic as they are today. However, because paleomagnetic evidence for relative displacements is lacking, and given that many authors (e.g., Miller et al., 1992; Burchfiel et al., 1992) include these areas within the late Paleozoic North American continental margin, we have decided to treat them conservatively the same way. Nevertheless, we note that these three areas have characteristic Tethyan or McCloud faunas that have suggested to some authors (for discussion, see Miller et al., 1992) that they were located farther south in the late Paleozoic to early Mesozoic.

South of the Washington–Oregon–Idaho tri-state border, the Blue Mountains contain late

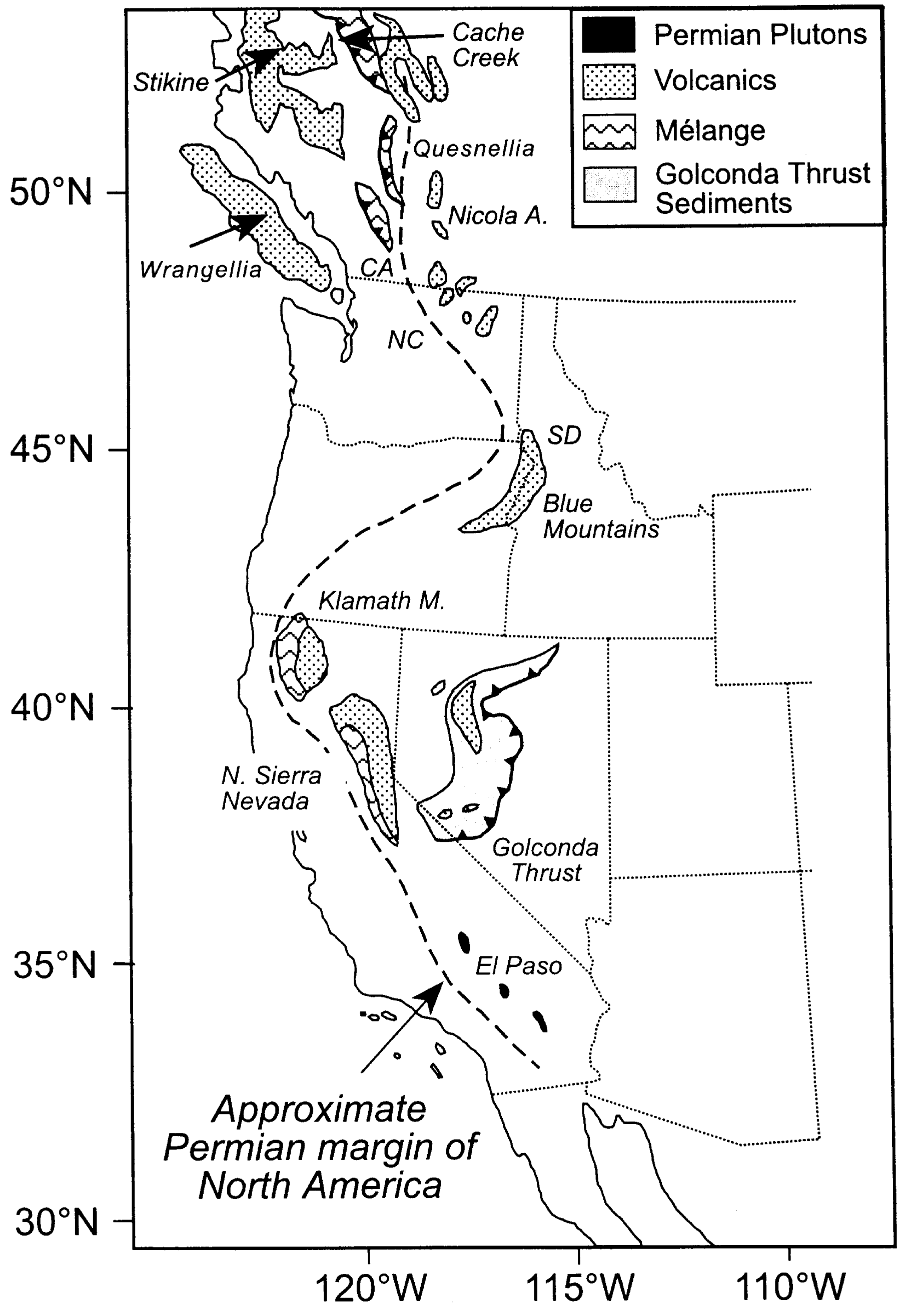

Fig. 5. Map of the western U.S.A. and southern British Columbia with those areas highlighted that reveal Permo–Triassic plutonism, arc-related volcanism or deformation. The approximate margin of the North American continent (after Burchfiel et al., 1992) may include suspect terranes for which displacements with respect to the craton are unresolved. CA = Chilliwack arc, NC = Northern Cascades, SD = Seven Devils arc.

Paleozoic to early Mesozoic terranes that are called the Grindstone terrane and the Seven Devils arc. The Grindstone sequence begins with clastic rocks containing terrigenous plant debris of Pennsylvanian age; plutonic and volcanic debris are also found. This part of the sequence is interpreted to have been deposited in a non-marine or shallow marine environment. However, these rocks seem to represent displaced blocks or olistoliths, so their stratigraphic value is questioned (Blome and Nestell, 1988). Similarly, overlying Lower Permian shallow water limestones with volcanic debris and blocks of Permian chert are thought to be olistostromes in an argillaceous matrix of Triassic age (Silberling and Jones, 1987).

Volcanic rocks of Late Permian age are present in the Seven Devils arc. The lower part of this section is composed by conglomerates, volcanic sandstones, keratophyric water laid tuffs, mafic breccias and spilitized flows. The Seven Devils arc rocks are overlain by volcanic and volcaniclastic rocks of Middle to Late Triassic age, followed by Upper Triassic platformal limestones. The various terranes within the Blue Mountains are thought to record the evolution of a single convergent margin (Miller et al., 1992).

Paleozoic or Triassic arc-related rocks also occur within the two easternmost and structurally highest lithologic belts of the Klamath Mountains: the Central Metamorphic Belt and the Eastern Klamath Belt (Cashman, 1980). In the latter, Devonian to Jurassic volcanic rocks outcrop in the vicinity of Redding, consisting of pyroclastic, siliciclastic and volcanic rocks. In the eastern Klamath Mountains the Devonian and Lower Carboniferous rocks are conformably overlain by Upper Carboniferous to Lower Permian fossiliferous, shallow water platformal limestones, although this contact may become disconformable in certain areas (Miller et al., 1992). The fossils in these limestones include fusulinids, corals and brachiopods and are characteristic of the McCloud fauna. The Permo–Triassic section is not continuous, and may lack the Early Triassic (Miller and Harwood, 1990). It consists of interfingered mafic and intermediate lava flows and waterlain pyroclastic rocks during the waning stages of arc-related volcanism, while the facies changes from shallow water to more basinal with turbidites and hemipelagic sedimentary rocks.

The northern Sierra Nevada rocks of interest are upper Paleozoic in the west and Triassic–Jurassic in the east, unconformably overlying Cambrian to Middle Devonian oceanic rocks of the Shoo Fly complex (Varga and Moores, 1981). Intermediate to mafic volcanic and volcaniclastic rocks constitute the base of the sequence and can be as young as Early Permian, followed by middle Permian shallow water and subaerial andesitic breccias, tuffs and lavas. The island-arc volcanism is thought to have been initiated in the late Early Permian, reaching a maximum in the middle Permian. Harwood (1988) suggested that the basaltic magmatism represented by the oldest volcanics overlying the Shoo Fly complex was coeval with uplift and erosion, pointing out the presence of normal faulting and extension during Late Carboniferous and Early Permian times. In some areas of the Sierra Nevada the Permian volcanic rocks are overlain by Upper Triassic limestones above a local unconformity, and elsewhere, disconformably, by Lower Jurassic rocks.

If the Blue Mountains, Klamath and northern Sierra Nevada areas were indeed already part of the North American continental margin in the Permian and Triassic, then their arc-related volcanism is roughly coeval with, and probably related to, the Sonoma Orogeny characterized by the more easterly Golconda overthrust. Their origin can then be ascribed to convergence between Pangea's northwestern margin and parts of the Panthalassa ocean to its west.

4.6. Little or no Permo–Triassic subduction on Pangea's leading edge

Summarizing what we have discussed thus far, it appears that little or no evidence exists for ocean-continent convergence on Pangea's leading edge, with the only exceptions being found in the western U.S. and Canada. It is worthwhile, however, to keep in mind that this area is a part of Pangea's margin that is very close to the Euler pole describing Pangea's movement with respect to the rotation axis (Fig. 3). This means that one would not expect to see much effect of the motion of Pangea with respect to Panthalassa, if any occurred along the lines of the model in Fig. 3. Yet, of all the regions that we examine on the leading edge of Pangea, this area shows more evidence, even if limited, for convergent ocean-continent plate interactions during the Permo–Triassic than any of the other areas.

Apart from the geology of the western U.S., the evidence suggests that, if any plate convergence took place along the leading edge of Pangea, whatever subduction occurred left little or no evidence in the form of major plutonism and compressive deformation. Moreover, the evidence that exists is primarily in the western U.S., where it suggests a different Pangea–Panthalassa convergence than that being hypothesized in Fig. 3.

5. Displaced terranes

Several displaced terranes, mostly in western North America, show evidence for arc-related volcanism, as well as compressive deformation, during the 295–205 Ma interval. In the subsections below, we will discuss this evidence and also address the—often poorly understood—evidence for where these terranes were located in Panthalassa. Most of the latter evidence comes from paleomagnetism but, as is well-known, this evidence pertains, at best, to paleolatitude, whereas paleolongitudes remain indeterminate with paleomagnetic techniques. This is a major limiting factor in our analysis and examination of northward ocean-floor subduction within Panthalassa.

5.1. Northeast Siberia

The location, origin and relations among the many oceanic and continental slivers and fragments accreted to the Siberian craton are still a matter of debate. Fujita and Newberry (1982) and Churkin et al. (1985) point out a clear relation between some blocks of northeastern Russia (i.e. Chukotka with its Triassic and occasionally Permian clastic shelf deposits, for location see Fig. 4) and terranes in northern Alaska; both have boreal faunas (Parfenov, 1991). In contrast, upper Paleozoic Tethyan faunal assemblages have been reported in some other terranes, e.g., around the Omolon Massif (Fig. 4), similar to the fauna in several displaced terranes in the North American Cordillera (Belasky, 1994); these faunas suggest significant northward movements subsequent to the Paleozoic. Accretion to Siberia occurred along the Verkhoyansk Mountains in Late Jurassic–Late Cretaceous times. In support of these relative displacements, paleomagnetic data for the Omolon block (for location see Fig. 4) indicate a large northward relative movement with respect to the Siberian Shield. Triassic paleolatitudes of about 40° (Khramov and Ustritsky, 1990) can be compared to the near-polar Late Triassic paleolatitude for the Verkhoyansk margin of the Siberian Shield (see Fig. 3), and suggest that Omolon was located some 3000 km or more from Siberia.

However, the paleolongitude of the terrane is unknown; examination of Fig. 3 indicates that a 40° paleolatitude can imply a position with respect to Siberia ranging (in present-day coordinates) from far to the west–southwest, to the southeast, or even far to the northeast.

Be that as it may, there is very little evidence for volcanism or compressive deformation during the 295–205 Ma interval in the northeast Siberian terranes. The exceptions are found in two ancient island arcs, which have been recognized as active during Permian and/or Triassic time. These are the localized Alazeya arc area (AL in Fig. 4) and the Udsko–Murgal'sk arc remnants in the Cretaceous Okhotsk–Chukotka Volcanic Belt along the coast of the Sea of Okhotsk (v-pattern in Fig. 4). The Alazeya arc is composed primarily of volcaniclastic sandstones, conglomerates and tuffs, indicating proximity to an active island arc. According to Parfenov (1991), oceanic crust from Panthalassa was consumed under or near the Alazeya terrane from Carboniferous through Early Jurassic times. The Udsko–Murgal'sk arc rocks consist of Upper Permian or younger volcanic outcrops wrapped up as thrust remnants in predominantly Cretaceous volcanogenic and sedimentary sequences (Andrei Khudoley, pers. comm., 1996). The upper Paleozoic/lower Mesozoic rocks include high-pressure metamorphic rocks, volcaniclastic turbidites and island-arc volcanics. Parfenov and Natal'in (1986) believe that the Udsko–Murgal'sk arc became active in the Late Permian as a convergent boundary close to the Siberian Platform, but the paleomagnetic evidence for large displacements of the Omolon Block (Khramov and Ustritsky, 1990) and areas farther outboard (e.g., Didenko et al., 1993) renders this rather unlikely.

5.2. *Alaska*

Alaska is a collage of a large number of accreted terranes. Moore et al. (1994) recognize four main types of accreted terranes in this zone. The first type includes blocks of continental nature that have a clear affinity with the continental margin of North America. They show a similar development as the North American cratonic margin with non-marine to marine sedimentary sequences deposited during much of the late Paleozoic and early Mesozoic at a west-facing passive margin. A number of these continental terranes now occur as thrust sheets of shelf and slope strata that were detached from their crystalline basement. Examples include the Arctic Alaska, Porcupine, Yukon–Tanana, Seward, Nixon Fork and Ruby terranes, which were deformed during the late Mesozoic and early Cenozoic. These are representative of the North American margin evolution for the time of interest. The next oldest orogeny in some of these terranes is the Ellesmerian during the Devonian and Early Carboniferous. Just like their North American continental margin counterparts, such terranes do not show any evidence for subduction during the times of interest to us; clearly these terranes were part of a subsiding passive continental margin from the Late Devonian to the Early Cretaceous.

The second type of terranes is named 'continentalized' in recognition of their evolution from an oceanic origin to a more stable continental nature. This type is represented by volcanic rocks of late Paleozoic to Early Jurassic age, which have been highly deformed and intruded by younger rocks in later Mesozoic times, resulting in a thickened and rather stable crust. The Peninsular, Alexander, and Wrangellia terranes in southern Alaska are good examples. Within those terranes of Paleozoic–early Mesozoic age, we find sedimentary, volcanic and plutonic rocks that are clearly related to island arc activity, but most of this activity is either older (e.g., in the Alexander terrane) or younger than the time interval of interest. Moreover, paleomagnetic results from these terranes indicate that they were located at near-equatorial paleolatitudes in the late Paleozoic or Triassic (Hillhouse, 1977; Van der Voo et al., 1980; Haeussler et al., 1992); if their paleolongitudinal positions are assumed to be adjacent to the North American craton, then they were located very close to the Euler pole of Fig. 3.

The third type of terrane defined by Moore et al. (1994) is thought to have originated at the ocean-continent interface, possibly adjacent to the North American passive margin. They contain Paleozoic to Jurassic clastic sediments in northern Alaska (Slate Creek, Prospect Creek and Venetie terranes) and mostly upper Cretaceous or younger terrigenous clastic deposits in southern Alaska, where examples include the Chugach, Prince William, Yakutat and Pingston terranes.

The fourth and last type of terrane comprises remnants of oceanic crust and seamounts, and marine platforms. Many of the ocean-floor rocks are Devonian to Late Triassic or Early Jurassic in age and comprise diabase, pillow basalts and cherts with tuff, argillites, graywackes and limestones. Some are only preserved as ophiolites. A number of Jurassic and younger island arcs developed on these terranes of intra-oceanic origin, which testifies to a late Mesozoic period of maximum convergence within Panthalassa. Older exceptions are found in the small Susitna terrane, where arc development started in the Late Triassic, and in the quite localized Strelna terrane (between the Chugach and Alexander–Wrangellia terranes), where convergence is inferred to have occurred in Late Carboniferous (to Early Permian?) times (Moore et al., 1994). The paleolatitudes of these terranes are largely unknown.

Summarizing, it appears that nearly all of the Alaskan terranes, regardless of whether they were displaced over large or short distances, are lacking in indicators of Permo–Triassic Panthalassa–Pangea convergence. A few intra-oceanic terranes (e.g., Strelna) show island-arc activity during the 295–205 Ma interval, but their extent is quite minor and their paleolocations within Panthalassa are unknown.

5.3. North American Western Cordillera

In this section we will briefly describe the Late Carboniferous to Late Triassic geology of the relevant displaced terranes, paying particular attention to evidence for ocean-continent or ocean-ocean convergence and compressional deformation events, in addition to what we know about the paleogeography of the terranes. Only evidence relevant to the implications regarding TPW from Late Carboniferous to latest Triassic time will be discussed.

5.4. The Western Intermontane Superterrane: Stikine and Cache Creek (see Fig. 5)

The western part of the superterrane (for definition see Cowan et al., 1997) contains well stratified Lower Devonian to Middle Jurassic volcanic and sedimentary strata, as well as plutonic rocks that are probably coeval with the volcanism. Above a post-Mississippian–pre-Permian unconformity, the succession in Stikine consists of mafic volcanic rocks grading up into argillites, which in turn are overlain by Permian carbonates. These carbonates are the most distinctive Paleozoic unit and consist of a dark, thinly bedded, argillaceous, pyritic micritic to calcarenitic limestone and a light calcarenite that is locally dolomitic.

There are limited data on the pre-Jurassic volcanic rocks of the Stikine terrane and this precludes a precise interpretation in terms of tectonic setting. It has been speculated that they represent arc-related rocks (Monger et al., 1994), and the carbonates are thought to represent banks surrounding the volcanic edifices. Nonetheless, very uniform and tabular carbonates in western Stikine suggest deposition on a uniform and stable substratum.

The Cache Creek terrane consists of serpentinized peridotite, tholeiitic and alkalic basalts, chert, argillite and limestone that are generally interpreted as remnants of ocean floor and deep-marine sedimentary rocks. The sequences range in age from Paleozoic to Late Triassic (Monger, 1984). An important characteristic is the occurrence of conspicuous limestone blocks ranging in age from Carboniferous to Late Triassic, which are commonly associated with alkalic pillow lavas. These blocks contain equatorial fauna intermediate between the North American and Mc Cloud assemblages. Monger interprets them as reef deposits accumulated on ocean seamounts. The structure and metamorphic facies (progression to blueschists) suggest that the terrane represents an accretionary complex.

The paleogeographic locations of Stikine and Cache Creek since Late Cretaceous times are disputed (see the discussions by Beck, 1992 and Cowan et al., 1997), with some paleomagnetic arguments favoring northward displacements of about 1100 km since 90 Ma (e.g., Irving et al., 1996), and geological as well as other paleomagnetic arguments favoring little or no displacement (e.g., Butler et al., 1991). Jurassic paleolatitudes from Stikine are difficult to interpret in terms of relative displacements with respect to North America (Beck, 1992), because of uncertainties about the cratonic reference poles.

5.5. The Insular Superterrane: the Wrangellia, Alexander and Taku terranes

The parts of this superterrane have long been recognized as far-travelled. Extensive and reliable paleomagnetic evidence documents that they have been displaced northward, over at least 2500 km and possibly more, since Late Triassic times (Symons, 1971; Hillhouse, 1977; Yole and Irving, 1980; Panuska and Stone, 1981; Haeussler et al., 1992; see also McClelland et al., 1992; Irving et al., 1996 and Cowan et al., 1997). The results collectively indicate a paleoposition for Wrangellia close to, or even south of, the Euler pole of Fig. 3.

Middle Devonian through Triassic rocks in these terranes, moreover, record a tectonically stable and perhaps rift-related environment of deposition. The various parts of Wrangellia, now dismembered and dispersed between southern Alaska's Wrangell Mountains, Queen Charlotte Islands, Vancouver Island and possibly Hell's Canyon in Idaho, include a characteristic sequence of Upper Triassic tholeiitic basalts, supporting the interpretation of a rifting environment. Arc-related rocks are found but are either older or younger than the periods of interest.

5.6. Northern cascades

This area is part of the Coast Mountains Orogen of Cowan et al. (1997). In northwestern Washington, the Chilliwak terrane contains Mississippian rocks, succeeded by argillaceous limestone containing early Upper Carboniferous fauna of Tethyan affinity (Miller et al., 1992). Volcaniclastic sandstone, conglomerates and argillites disconformably overlie the limestones and contain both shallow marine and terrestrial fauna of latest Carboniferous or Early Permian age. Lower Permian rocks include micritic and calcarenitic limestones, which are in turn overlain by, and interfingered with, Permian basalt flows and intermediate to siliceous tuffs. The presence of faulting, basaltic volcanism and the development of carbonate platforms suggests extension and possibly rifting that might have affected this terrane in the Early Permian. Lower Permian sequences are followed by undated volcanic rocks that are in turn overlain by upper Triassic argillite and siltstone.

6. Discussion and conclusions

To examine whether TPW has been a possibility during the interval of 295–205 Ma, we have discussed the geology of the leading cratonic edges of Pangea, suspect terranes at these continental margins, and displaced terranes now accreted to these margins, in terms of evidence for subduction of Panthalassa's ocean crust. The evidence for subduction is primarily restricted to Cordilleran terranes and the western margin of cratonic North America in the U.S.A. Less abundant arc-related rocks in displaced terranes in northeastern Siberia and southern Alaska have also been documented.

The terranes for which displacements have been documented, and in which evidence for subduction is abundant, were generally located in low paleolatitudes, near the equator or in the southern Panthalassic hemisphere. Longitudes for these terranes are entirely unknown. For the suspect terranes, unfortunately no quantitative evidence is available for ancient paleolatitudes, but their Tethyan or McCloud faunal affinities have been interpreted as indicating low latitudes (Ross and Ross, 1983). We recognize, however, that this interpretation is not universally agreed upon. That part of the leading edge of cratonic Pangea that shows ocean-continent convergence was also located in low latitudes (0–25° North; see Fig. 3) during the time interval of interest. No evidence, however minor, for ocean–ocean or ocean–continent convergence has been found in latitudes greater than 40° North (see Fig. 3A for context). Generalizing, we can say that the bulk of the observed island-arc or active-margin activity has been restricted to latitudes more southerly than 25° North.

This implies that even if one assumes northward compensatory movements of parts of Panthalassa in accordance with the Euler pole of Fig. 3, these oceanic parts comprise at most about 70% of one hemisphere, or about 35% of the total surface of the Earth. Referring to Fig. 2 (left), it is clear therefore that an imbalance existed during the 295–205 Ma interval between the areal extent of the northward moving, largely continental Pangea hemisphere (stippled in Fig. 2) and the areal extent of that part of the largely oceanic Panthalassa hemisphere that possibly moved northward (white plate in Fig. 2). To be fully compensating, this smaller northward-moving part of Panthalassa would then have to be moving at twice the speed of the Pangean plate.

The above is one extreme in a range of scenarios, the other extreme being that—even though parts of Panthalassa were undoubtedly subducting—the oceanic hemisphere had no consistent northward movement about the Euler pole at all. This is not as unlikely as it may appear at first. Subduction under continental margins, such as deduced for the area involved in the Sonoma Orogeny, is usually more orthogonal than highly oblique. In other words, it seems *a priori* more likely that oceanic crust was subducting in a (paleo-) easterly direction under the western margin of North America in the Permo–Triassic, than in a (paleo-)northerly direction.

It is also worth mentioning that in this paper we have ignored subduction evidence from the *trailing* margin of Pangea, because its existence can neither establish nor refute the existence of

TPW. However, it has long been known that there was Permo–Triassic subduction under the New Zealand, Australian, Antarctic and northern Argentinian margin of Gondwana (the 'Samfrau' belt) and this indicates that within Panthalassa an oceanic plate of unknown extent was moving towards Pangea's trailing edge (Visser and Praekelt, 1998). If this plate was large, which we do not know, and given the likelihood it moved in the same sense as Pangea (but even faster), it would enhance, not detract from, the possibility of TPW.

Despite the lack of knowledge about paleolongitudes of any of the possibly displaced terranes of island-arc character, and an equally lacking knowledge about the orientations of these arcs and, hence, the directions of subduction underneath, we can conclude that TPW probably occurred in the 295–205 Ma interval, and was possibly as large as 35°. This implies a TPW rate of about 0.4°/m.y.

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