Palaeoenvironments and elemental geochemistry across the marine Permo-Triassic boundary section, Guryul Ravine (Kashmir, India) and a comparison with other North Indian passive margin sections

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Abstract
The Guryul Permian-Triassic sediments were deposited on a passive margin in a partially enclosed basin, at a latitude of about 45°S along the southern margin of the Neotethys Ocean. The closest modern analogy is the Japan Sea rotated into a southern hemisphere orientation. Guryul element variations are subdued, mostly lithological, and take place across the change from the dominantly sandy shallow-water Permian Zewan Formation into the dominantly shaly deeper-water latest Permian-Triassic Khunamuh Formation, and not at the palaeontological Permian-Triassic boundary. The overall geochemistry of the Guryul sediments is very similar to that of the Ulleung Basin in the Japan Sea, where the sediments are 40%-60% eolian from loess and re-deposited loess in the North China Basin. The overall Guryul geochemistry indicates that the sediments were derived from dominantly silica-rich continental rather than silica-poor sources although with some more silica-poor inputs at times. Element ratios suggest increasing aridity and wind abrasion across the Permian-Triassic boundary. Various geochemical redox proxies suggest mainly oxic depositional conditions, with episodes of anoxia, but with little systematic variation across the extinction boundary. Productivity proxies suggest a slight decrease across the Permian-Triassic boundary, and at least one more decreased interval in the Early Triassic. The similar geochemistry of other localities indicate that the Guryul geochemical changes are regional in extent and representative of the North Indian Permian-Triassic passive margin. The lack of consistent element geochemical changes across the boundary accompanied by significant C, S, and other isotopic changes suggests that atmospheric and oceanic chemistry rather than physical changes, such as provenance, climate and sea-level changes, drove the Permian-Triassic environmental changes and extinctions at least on the mid-latitude Tethyan shelf of northern India.

KEYWORDS
Extinction, geochemistry, Kashmir, Permian-Triassic

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1 | INTRODUCTION

The greatest extinction in the geological record took place at the end of the Permian around 252 Ma (Baresel et al., 2017), supposedly killing over 90% of all marine species, (though this may be be 81% according to Stanley, 2016), and about 70% of all terrestrial species (Erwin, 2006). Both gradual environmental and catastrophic events have been used to explain the extinction but so far with little consensus (Wignall, 2015). One problem is that relatively few thick sections allow individual late Permian events to be separated. For example, at the Global Stratotype Section and Point (GSSP) for the Permo-Triassic boundary at Meishan, Southeast China, the change of rock type (limestone to marl) and biostratigraphic boundary are separated by less than 0.3 m (Yin et al., 2001).

In other sections, like Shangsi (South China), East Greenland, and Kashmir (India), the change of rock type and the biostratigraphic boundary are separated by thicker sediments, which allow distinct events to better separated than at Meishan (Nakazawa et al., 1975; Wignall et al., 1995; Stemmerik et al., 2001; Wignall and Twitchett, 2002). The palaeontology, stratigraphy, and some isotopic characteristics of the Kashmir sections are reasonably well-known (Nakazawa et al., 1975; Brookfield et al., 2003, 2013; Wignall et al., 2005; Brookfield and Sun, 2015; Tewari et al., 2015; Huang et al., 2019). Geochemical studies across the boundary in Kashmir are limited. Algeo et al. (2007) analyzed the elements and some isotopes in irregularly vertically spaced samples from −8 to +8 m from the boundary, which concentrated on the limestones. Brookfield et al. (2010) analyzed the Platinum Group Elements in 0.3 cm spaced samples from −0.5 to +1.3 m across the boundary, which indicated no abnormal extraterrestrial components. Kumar et al. (2017) analyzed five widely spaced samples from −3 to +6 m across the boundary, which indicated increased weathering, greater reworking, and a change from oxidizing to reducing conditions across the boundary, with a continental crust source. Mir et al. (2016) analyzed eight samples for major elements, indicating low to moderate degrees of weathering with little change, and a moderate degree of weathering for plagioclase feldspar, but unfortunately did not give any detailed stratigraphic information, other than formation, so no trends can be proposed.

The aim of this paper is interpret the palaeoenvironments, compositional variability and geochemical changes in elemental distributions from the Upper Permian through the Lower Triassic at the expanded Guryul ravine section in the context of palaeoenvironments inferred from sedimentology, petrology and palaeontology, and using analogous Quaternary to Recent oceanological situations and histories for comparison. Individual studies like this one, by themselves, should neither be extrapolated to evaluating regional and worldwide changes, nor to evaluate possible causes for the end Permian extinctions, although some comparisons have been made with other sections on the Permian–Triassic north Indian passive margin (see Section 12).

Samples, 5 cm thick, were taken at regular 30 cm intervals in the Khunamuh shales, and at irregular intervals in the underlying Zewan Formation wherever shaly lithologies were present. The 30 cm spaced sampling did not intersect the lensitic bioclastic limestones of the lowermost Khunamuh Formation material which was introduced from shallower water by storms or tsunamis (Nakazawa et al., 1975; Brookfield et al., 2013), probably in broad shelf channels as they are not found in nearby exposures to the southeast. This study therefore considers the geochemistry of only the finer-grained sediments as indicator of the local environments.

2 | GEOLOGICAL SETTING

In the late Palaeozoic, Kashmir formed part of the northern Gondwana continental margin and lay on the southern side of the Palaeoethys Ocean adjacent to Oman (Figure 1). Rifting and eruption of early Permian basalts (Sakmarian around 290 ± 3 Ma) (Shellnutt et al., 2015) were followed by separation of blocks off the northern edge of Gondwana, the formation of a Neotethys Ocean and rapid thermal subsidence of the northern Gondwana margin in the later Permian and Triassic (Brookfield, 1993). The Permian–Triassic sections of the northern Indian passive margin are now in separate structural blocks, which have been displaced southwards relative to the Indian craton, by as much as 2,000 km in the case of the High Himalaya (Brookfield, 1993; Dipietro and Pogue, 2004; see Section 12).

During the latest Permian, Kashmir was at about 45°S, in the area of northerly warm ocean currents, and persistent westerly winds from the Karoo Basin, and far south of the Pangea mega-monsoon area (Parish, 1993; Figure 1). Through the Permian, global warming transformed these temperate latitudes from cold periglacial, through warm and wet, into hot and dry (hyperarid) environments (Chumakov and Zharkov, 2003). Earliest Triassic biotas suggest a hot equatorial dead zone, flanked by cooler areas with arid to semi-arid environments on land above ~30° latitude and warm water biotas in the sea (Sun et al., 2012). During the last Changhsingian stage of the Permian, tropical marine temperatures seem to have rocketed from 20 to 32°C culminating in temperatures of 38°C in the earliest Triassic (Sun et al., 2012). Arid climates occupied up to 80% of the land, and average summer temperatures in the Karoo Basin could have been >40°C (Chumakov and Zharkov, 2003; Kiehl and Shields, 2005). If the marine temperature changes were true, then thermal expansion associated with them would increase the total volume of the oceans, raise late Permian sea level by over 20 m, and account for at least part of the latest Permian marine transgression (Southam and Hay, 1981).
Clastic sediment to the Kashmir area was supplied from the Karoo Basin area to the southwest where, in the latest Permian to earliest Triassic, drainage from the southern Cape mountains was roughly northeast across arid braided alluvial plains into arid floodplains and playa lakes, and coastal sabkhas in East Africa and India, into the western Neotethys (Visser, 1995). Although complicated by topography inherited from earlier Permian rifting, the main controls on Permian–Triassic Karoo sedimentation were source tectonics and climate, with an increasingly hot climate with fluctuating precipitation (Nichols and Daly, 1989; Catuneanu et al., 2005). From the late Permian into early Triassic, changes from meandering to low sinuosity rivers were accompanied by general drying of the floodplain habitats and redistribution of vegetation belts (Smith, 1995), although an interval of large lakes straddle the actual boundary (Yemane and Kelts, 1990). Coarser sediments are generally immature feldspathic sandstones (Pace et al., 2009).

### 3 | GURYUL RAVINE SECTION

The well-known locality at Guryul Ravine lies about 15 km east of Srinagar, Kashmir, and shows continuous, well-exposed sections across the Permo-Triassic boundary (Figure 2). The section (Figure 3) was a candidate for the GSSP for the Permo-Triassic boundary (Yin et al., 2001). The fossil defining the base of the Triassic, Hindeodus parvus, first appears at +2.0 m, at the base of unit E2 of the Khunamuh Formation, co-incident with the first appearance of the earliest Triassic ammonoid, Otoceras woodwardii Griesbach (from current knowledge—conodont studies are in progress on the bioclastic limestones of the Zewan–Khunamuh transition). Anoxia, as determined by populations of small pyrite framoids, first appears at 1 m above the base (Wignall et al., 2005), with maxima at +2 and +4.3 m corresponding with low species numbers and with an oxic horizon between (Huang et al., 2019).

The Zewan Formation is divided into four main units (Nakazawa et al., 1975; Figure 3).

- **Unit A** consists of calcareous sandstones (arenites) with minor micaceous shales with a basal extraformational conglomeratic bed (A1) passing up into thick bedded sandy limestone (A2), unfossiliferous bioturbated sandy shales (A3) and calcareous sandstones and thin shales (A4). Except for A3, the beds are fossiliferous with brachiopods, crinoids and bryozoans. These indicate a normal shallow marine environment.

- **Unit B** consists of thick unfossiliferous, frequently bioturbated, micaceous sandy shale, muddy sandstone and alternations of sandstone and thin shale (B1); overlain by a thin, similar but fossiliferous unit of alternations of micaceous sandstone and sandy shale with many brachiopods and a small number of bivalves, crinoids, bryozoans and foraminifera (B2); in turn overlain by bioturbated micaceous sandstone and sandy shale (B3). The presence of trace fossils without body fossils in most of Unit B indicates a brackish environment (Buatois et al., 2005), but with periodic marine incursions shown by the beds with a normal marine fauna: these deposits, however, are not “deltaic” as they are only 20 m thick.

- **Unit C** consists of rhythmic alternations of calcareous sandstone and bioturbated, micaceous sandy shale. The sandstones show parallel and cross-laminations and graded structures in beds 10–20 cm thick and there are several shell-beds with gastropods, bivalves, brachiopods and other fossils (Nakazawa et al., 1975). These indicate a storm-influenced normal shallow marine environment.
Unit D is dominated by thick-bedded, parallel and cross-laminated, calcareous fine to medium-grained, well-sorted, often argillaceous, very sandy limestones, or calcareous quartz sandstones, interbedded with sporadic thin calcareous shales. The upper beds show synsedimentary deformation (Figure 4A). Most of the quartz grains are angular to sub-rounded and first-cycle (Figure 4B, 4), but a few are very well-rounded, indicating a desert source for those (Nakazawa et al., 1975). Some beds show unidirectional slumping (Figure 5A) suggesting appreciable slopes, possibly disturbed by earthquakes. Graded beds show consistent trends from parallel lamination to festoon (hummocky) cross-lamination typical of deposits of waning storms or tsunamis (Figure 5B) (Brookfield et al., 2013; Tamura et al., 2015). The large-scale convolute bedding seen in beds 41 and 43 and in the nodular beds 44–46 are attributed to liquefaction and slumping during seismic shaking (Brookfield et al., 2013). These indicate a storm or tsunami-influenced shallow marine environment, perhaps shallower than unit C, especially as the topmost beds contain abundant chert grains possibly derived from erosion of underlying Zewan cherty limestones (Figure 5C).

Although most of unit D has a normal marine fauna, bed D46-2 contains only the agglutinated foraminifera, Nodosinella, and ostracods (Figure 5C), indicating a restricted marginal marine environment (Murray and Alve, 2011) and possibly very warm as ostracods have the highest tolerance of any metazoans to elevated temperatures (Mason, 1939). The topmost Zewan bed (D46-3) is a bioclastic limestone identical to those in the lower Khunamuh Formation above and the major, and sharp environmental break between shallow and deeper marine shelf takes place below this bed—although for convenience and ease of location, the top of bed D46-3 is taken as the 0 point for measurements.

Unit E is the lowest part of the Khunamuh Formation and consists of dominantly argillite at the base with increasing limestone interbeds upward. The contact with the underlying Zewan Formation is taken below the first argillite (Bed 47), which has a very sharp contact with the
**FIGURE 3**  Permian to lower Triassic section at Guryul (after Nakazawa et al., 1975), with location of analyzed samples, description of units, and inferred palaeoenvironments. Age of Panjal Trap is from youngest zircon U/Pb age of Shellnutt et al. (2011). Age of Permo-Triassic boundary is from Mundil et al. (2004)
bioclastic bed below (Figure 6A and B). Subunit E₁ (beds 47–51) is black sandy and silty shales with discontinuous limestone lenses. Its lower 0.8 m (beds 47–48) is dominated by medium dark grey (N4) massive silty mudstones with hummocky cross-stratified bioclastic limestone lenses containing late Permian faunas (Figure 6B) (Nakazawa et al., 1975; Brookfield et al., 2003). These bioclastic limestone lenses contain scattered small angular chert and larger rounded limestone clasts (Figure 6C) in a sandy shelly carbonate matrix (Figure 6D). Subunit E₁ changes, over a 20 cm transition of dark grey (N3) hackly silty mudstones, into grayish black (N2) platy silty mudstones with persistent micritic limestone beds (Figure 7A) which dominate the section up to the +20 m level where detailed sampling ended (Table S1). The mudstones are laminated and bituminous with abundant (up to 50%) quartz silt and with patches of quartz sand which may be infillings of small burrows (Figure 7B); the limestones often have broken
ammonoid fragments (Figure 7C). Many quartz silt grains in the argillites are well-rounded (Figure 6B top, Figure 7B) which is an effect of shape sorting by wind (sorting by abrasion does not occur in silt sizes) as water transport has no effect upon the shapes of quartz silt grains (Mazzullo et al., 1992).

The topmost Zewan (D46-3) and lower two Khunamuh pebbly bioclastic limestones (intrabioclastic grainstones) contain abundant angular and well-rounded quartz sand but no chert grains (Figure 6C) and show intense wave reworking with hummocky cross-stratification and grading to finer-grained tops. They were deposited by successive very large waves; calculations of wave amplitudes and heights are more compatible with tsunami than storm waves (Brookfield et al., 2013). There may have been significant erosion associated with these bioclastic lenses. Current conodont information suggests that the latest Permian preparvus zone of unit E1 is highly condensed at Guryul (Korte and Kozur, 2010). The equivalent of unit E1 at Barus and Mandakhpal is 6 m thick compared with 2 m at Guryul, and has no bioclastic limestone beds (Nakazawa et al., 1975; Nye et al., 2016; Huang et al., 2019).

These beds are sharply overlain, and interbedded with, calmer water grayish black silty argillite dominated by
pelagic organisms (Brookfield et al., 2003) (Figure 7). The characteristics of these lower Khunamuh Formation beds indicate a deep marine shelf environment subject to periodic input of shallower marine bioclastic sediment during storms and/or tsunamis for E1 and simple reworking and concentration of pelagic fossils for the overlying beds E2 and above.

According to current knowledge, unit A is Capitanian, units B to lower C are Wujiapingian, units upper C to E1 are Changxingian, and units E2 to F and above are Induan (Figure 8) (Algeo et al., 2007; Tewari et al., 2015). There is a progressive replacement in species upwards in the Zewan Formation and lowermost Khunamuh Formation, from beds A to E1, but with actual increase in diversity from beds B to D (from 3 to 21) and only a slight drop in E1 (18), although these species are mostly in the bioclastic beds (Figure 6). The end-Permian mass extinction, with a drop in diversity from 18 to 8 species, and marked by an almost complete change in species (with only one, *Hindeodus typicalis*, crossing the boundary), occurs either in the middle of E1 or at its top (bed 51). The FAD of *H. parvus* defining the base of the Triassic is at the base of bed 52, together with *O. woodwardii*, which names the basal Triassic ammonite zone (Figure 8). The conodont-defined Permian–Triassic boundary may thus, at Guryul, mark the main end Permian extinction. Limited palynology is consistent with this (Tewari et al., 2015). The impoverished palynoflora of the Zewan Formation is non-diagnostic. However, unit E1 palynoassemblage of spores, pollen and fungi is typical of the Permian–Triassic transition flora of Gondwana (Jha et al., 2018); the absence of marine acritarchs in the Khunamuh is puzzling though.

The environmental changes based on sediments and fossils suggest a change upwards in the Zewan and lower Khunamuh formations, in a basically clastic marine environment, from a shallow marine, periodically river-influenced shelf (unit A) through brackish with periodic marine incursions (unit B) through storm-influenced shallow marine (unit C), through storm-influenced marginal marine (unit D) to deep normal marine shelf (unit E), with slumping indicating slopes in the top Zewan (units C and D) (Figure 3). The main changes also suggest a major rise in sea-level between units D and E from

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**FIGURE 8** Biostratigraphy of the Guryul Permian–Triassic boundary section showing stratigraphic range of fossils and number of species in each subdivision (data from Nakazawa et al., 1975). Bryo – bryozoans. Gastr – gastropods. Note that there is little extinction at the Zewan-Khunamuh Formation boundary and the main extinction (and new species appearance) is at the FAD of *Hindeodus parvus*, and the first inferred anoxic level marked by pyrite framboids. Anoxic levels from Huang et al. (2019)
a shallow to marginal marine to a deeper shelf environment. The total thickness of the Zewan of less than 100 m is far too thin to be a marine deltaic successions and the petrology of the sandy beds indicates first cycle clastic input with at least some reworking of eolian grains. This interpretation forms the basis for the choice of modern analogies for geochemical comparisons.

4 | RECENT ANALOGIES

In order to interpret the depositional environments and geochemistry of the Guryul Permo-Triassic sediments, it is considered essential to compare them to the oceanography and deposits in analogous Recent seas and their inferred history over at least the Quaternary. In the Permo-Triassic, northern Gondwanaland from India to Arabia was a northwest aligned passive continental margin on the western side of a large ocean (Figure 1). The palaeoceanographic situation of this margin has no exact modern equivalent, but it is comparable to the Recent western North Atlantic and western North Pacific margins, if these are horizontally flipped into a southern hemisphere position. The North Atlantic passive margin has no large offshore islands unlike the Permian western Panthalassic Ocean. The western North Pacific, however, has numerous magmatic arcs on continental fragments that, although related to subduction, are still similar in their oceanic effect to the rifted island in the Palaeotethys ocean, and their associated marginal back-arc oceans with their passive margins rest partly on continental crust (Liu et al., 2014). Translating the western North Pacific margin to a southern hemisphere position gives a palaeoceanographic situation very similar to the southern hemisphere northeastern Gondwanaland margin (Figure 9).

Of the North Pacific marginal seas, the Sea of Japan encompasses the same latitude (~40°) as the Guryul sections, has predominantly thinned continental crust with oceanic crust in its southern part (Yoon et al., 2014), and is likely to be the closest in basic climate and oceanographic character to the Permian northern Indian passive margin (Figure 10). Initial extension and rifting began in the Sea of Japan in the early Miocene with deposition of volcanlastic sandstones and siltstones along the basin margin (Kaneoka, 1986). Accelerated deepening of the Yamato Basin during the early Miocene and of the Japan Basin during the early middle Miocene was accompanied by deposition of fine-grained, diatomaceous and calcareous clay-rich pelagic and hemipelagic muds (Jolivet et al., 1994).

The Sea of Japan is enclosed by shallow sills, which formed a barrier to open ocean influx during low sea level at ~125 m of the last glacial maximum, and led to laminated muds and anoxia at times in the deeper part of the sea (Bahk et al., 2000; Kim and Kim, 2001). This may explain the comparable anoxia developed in the latest Permian off the northern India margin in Spiti (Williams et al., 2012). There are no major rivers flowing from the continent into the Japan Sea and thus no thick delta deposits, unlike the Yellow and East China seas to the south. The sub-basins rest on thinned continental crust, as do the northern India Permo-Triassic sediments (Brookfield, 1993), and the Ulleung Basin in particular has similar characteristics, like a ramp slope and fine clastic Quaternary cover that resemble the Guryul Permain-Triassic sediments. The vertical change in sediments during the Quaternary post-glacial rise in sea level can also be compared with the vertical changes up the Guryul section, even if the sea-level rise at Guryul is not glacio-eustatic.

The climates to the southwest of the Sea of Japan (on the transposed map) vary from dry winter continental with hot to cool summers and with cool semi-arid in the interior (Peel et al., 2007) (Figure 10). In winter, strong southwesterly winds transport redeposited loess from the North China Basin into the Japan Sea, and this eolian material forms over 50% of

![FIGURE 9 West Pacific marginal basins flipped to southern hemisphere orientation comparable to Permo-Triassic northern Gondwanaland. Inset top right shows paths of typhoons (from Yoon et al., 2014)
the detrital sediment in the Ulleung Basin (Anderson et al., 2019). In summer, weaker northeasterly monsoon winds carry Japan arc-derived sediment into the sea. The result is laminated silts and clays with higher contents of eolian and arc-derived materials respectively (Irino and Tada, 2000). The Guryul laminated Khunamuh argillites, with quartz silt laminae alternating with quartz-poor argillites may originate in the same way, although the laminae were not analyzed separately (Figures 4C, 6B, and 7B). Colder phases also cause greater aridity in the North China Basin, with the results that contemporary Japan Sea deposits contain around 60% of eolian sediment, in contrast to around 40% in interglacial periods (Irino and Tada, 2000). Tropical cyclones also produce periodic coastal upwelling as they track south along the coast (Figure 9 inset top).

The sea surface temperatures in the Sea of Japan are partly controlled by the Tsushima warm current, which enters via the Tsushima Strait and exits mostly at the Tsugara Strait, and partly by the Liman/Korean cold currents, which flow along the Primorye and Korean coasts. These meet along the SubPolar Front, a zone of rapid sea surface temperature change, which varies little in position through the year (Figure 10). The Sea of Japan has a very strong contrast between the warmer surface layers (3–28°C) and the very cold deep layers (0–1°C), because the deeper layers are essentially isolated by the shallow straits sills and annually replenished by the fall overturn of colder water in the north (Leonov, 1960). Summer surface temperatures range from 28 to 20°C with winter temperatures from 13 to 3°C and thus exhibit a seasonal range of about 15°C (Figure 10). The surface temperatures control the continental shelf faunal diversities which range, for (bivalves) from 313 species for South Korea to 157 species for Middle to North Primorye (Lutaenko, 2012) (Figure 10). Salinity decreases slightly northward from 34‰ at Tushima Straits (34°N) to 33‰ at the Strait of Tartary (50°N). The Subpolar Front position and surface circulation varied as the warm conditions of the interglacial periods changed into the cold conditions of the glacial periods. During interglacial periods with higher sea level, surface temperatures were higher than today and the Tsushima Warm Current strength increased due to deepening and widening of the Tsushima Strait; while the South Polar Front was displaced northward bringing warmer water to the central areas. During glacial periods of lower sea level (to about −100 m), the Tsushima Warm Current almost disappeared causing restriction and anoxia in the Japan Sea (Choi et al., 2012).

An increasing wind-transported dust component in all western Pacific seas corresponds with the Central Asia aridification event starting in the late Miocene at ~7.8 Ma (Wei et al., 2003; Liu et al., 2016). Millennial-scale variations in eolian sediment content can be used to infer Milankovitch cycles in the Japan Sea sediments (Yoo et al., 2003) and may be detectable in the Guryul section with more detailed and continuous sampling (Brookfield et al., in press).

Such oceanographic conditions and variations should be expected along the northern Indian Permo-Triassic margin.
The oceanographic complexities and space and time variations shown by modern seas and oceans ought to be considered in any interpretation of ancient seas and oceans. Both the Sea of Japan and the Permian–Triassic basin off northern India are bath-tub oceans with only limited connections with the open ocean across shallow sills (Williams et al., 2012; Chang et al., 2016).

The shelf, slope and basin sediments of the Ulleung Basin are therefore used as possible close analogs for the Guryul sediments (Um et al., 2009, 2013). These are supplemented by a loess sediment analysis (standard reference sample GSS-8) from the source of much of the Ulleung Basin sediments, the loess plateau west of Beijing (Song et al., 1987; Govindaraju, 1994). There is little geochemical variation across this plateau (Eden et al., 1994), although there is significant variation during the last 2.5 Ma, because of increasing aridity (Chen et al., 2001).

5 | SAMPLING

In 2007, samples were collected from the Guryul ravine exposure (latitude 34°04’26.7”N, longitude 74°56’44.8”E, altitude 1,800–1,900 m) at 30 cm intervals from the lower 20 m of the Khunamuh Formation, which includes the Permian–Triassic boundary, and sporadically at suitable shaly beds 70 m downwards into the Zewan Formation (Figure 3). The bed notation of Nakazawa et al. (1975) was used, and the major lithological boundary separating the Zewan and Khunamuh formations was used as a datum, which is easily identified, and lies about 2 m below the palaeontologically defined Permo-Triassic boundary (Figure 3). Even the most careful repeated section measurements, however, vary by 20% at the 95% confidence limit (Dennison, 1972).

Large samples were collected to take into account how grain size affects the size required for representative petrographic and geochemical sample analysis; and to allow others to re-analyze the same samples if they wish. For example, since many trace elements are concentrated in scarce individual grains, then the size of a representative sample increases with grain size. For the fine-grained sediments analyzed here, less than 5 g of sample is required for analytical consistency (Wickman, 1962).

Descriptions of the samples are in Table S1. The few Zewan Formation samples analyzed were identified in the field as micaceous fine grained clayey calcareous sandstones and non-calcareous sandy/silty mudstones in hand specimens; but their SiO2 content of less than 70% from chemical analyses identifies them all as mudstones (Moore and Dennen, 1970). All Khunamuh Formation samples analyzed are dark grey (N3) to grayish black (N2) non-calcareous sandy/silty mudstones and olive gray (5Y 4/1) to brownish gray (5YR 4/1) very fine-grained calcareous sandstones and sandy limestone.

At present, fairly large samples of the horizons collected in 2007 are housed at the Postgraduate Department of Geology of Jammu University, India, and splits are available to any workers who may wish to reanalyze the samples (please contact G.M. Bhat).

6 | ANALYTICAL METHODS

The samples were visually inspected and any veins and weathered surfaces were removed prior to grinding in a tungsten ball mill to a fine powder.

For major elements (except Na2O, MgO, and MnO) whole rock samples were powdered (<60 mesh) and homogenized by Spex® ball mill and analyzed by a Spectro XEPOS Bench Top Energy Dispersion X-ray fluorescence (XRF) at the Environmental Analytical Facility (EAF) of the University of Massachusetts (UMass) at Boston. Three to five grams of powdered sample was placed in XRF capsules and measured in triplicate under He purge for powdered samples. The USGS standards SDO-1 and SGR-1b, as well as NIST 1d were used as a three-point calibration curve. The USGS standard SCO-1 was repeatedly measured as an unknown to check the calibration curve and results were within 5% difference of certified values. All standards were processed in the same way as unknown samples. Reported errors represent the propagated error of replicate measures of each element.

For Na2O, 17 mg of powdered samples were completely digested in 7 ml Savillex digestion bombs using a combination of steps consisting of concentrated HF (1 ml) and HNO3 (3 ml), H2O2 (1 ml) and HNO3 (3 ml), and HNO3 (3 ml). Samples were dried down between each step, and the final solution was acidified with 0.5 ml of HNO3 and diluted to 50 ml with Milli-Q water. Sodium values were quantified using a PerkinElmer Optima 3000XL Inductively Coupled Plasma Optical Emission Spectrometer at the EAF at UMass Boston. The USGS standards SDO-1 and SGR-1b, as well as NIST standard 1d were used to produce a three-point curve for calibration and run after every 10 unknown samples. A separate digestion of SGR-1b from the sample used in the calibration curve was measured as an unknown. The unknown SGR-1b was repeatedly measured throughout the analytical run to monitor accuracy, and values were within 5% difference of the certified values.

For MgO, MnO, and trace elements, 17 mg of powdered whole rock samples were completely digested using the same method as the Na digestion. Trace elements, Mn and Mg were measured using a PerkinElmer ELAN DRCII inductively coupled plasma mass spectrometer at the EAF at UMass Boston. The USGS standard SDO-1 was serially diluted to produce a 3-point calibration curve and calibrations standards were run every 10 unknown samples. The USGS standard SGR-1b was repeatedly measured...
throughout the analytical run as a check of the calibration. Standards were digested and processed in the same way as the unknown samples. Indium was used as an internal standard and was added to each standard and unknown to monitor instrument drift. The SGR-1 check standard values were generally within ten percent difference of known values.

To determine total organic carbon (TOC), the methods of Wolbach and Anders (1989) were used on selected samples to supplement the limited data of Algeo et al. (2007). Reduced carbon residues (organic + elemental carbon, or bulk carbon) were isolated from selected samples using HCl and HF/HCl demineralization and splits of these residues combusted to CO₂ for mass spectrometric analysis to determine the stable carbon isotope composition. Organic carbon in the remaining sample splits was then selectively destroyed using acid dichromate oxidation at 160°C, and the total elemental (inorganic) carbon residues (TIC) measured. The amount of destroyed organic carbon (TOC) was then determined by difference using mass balance.

Whole rock REE values were normalized to chondrite values from Taylor and McLennan (1985). Ratios of light rare earth elements (LREE), middle rare earth elements, and heavy rare earth elements (HREE) were calculated using chondrite normalized values of La, Sm and Yb respectively. The Eu anomaly is calculated by Eu/Eu* = Eu/(Sm*Gd)¹/² and the Ce anomaly is calculated by Ce/Ce* = Ce/(La*Pr)¹/².

Loss-on-ignition was not used to calculate carbonate content as the results are strongly dependent on the amount of clay present, with greater loss with higher clay content (Veres, 2002). The classification used in the figures defines limestone as above 70% CaCO₃ (>30% Ca), calcareous mudstones as between 10% and 70% CaCO₃ (4%–30% Ca), and non-calcareous mudstone as below 10% CaCO₃ (4% Ca). This is a simplification of the Colorado School of Mines classification. There is a clear bipolar division between limestone and non-calcareous mudstones at Guryul with very few calcareous mudstones.

Guryul major and most minor elements were normalized to Al. When comparing element proportions in samples with variable carbonate and silica contents, an accepted practice, in Recent ocean sediment analysis, is to normalize element concentrations to the Al content, because it reflects the aluminosilicate fraction of the sediments, which moves little during diagenesis (Calvert and Pedersen, 1993). There are some problems, particularly where detrital fractions are lower than 3%–5% (discussed in Riquier et al., 2006), which is not the case with the Guryul sediments.

All Guryul sediments contain appreciable amounts of fine quartz sand and silt (Figures 4B,C, 5C, 6B,D and 7B; Brookfield et al., 2003), and since the coarse-grained sediment fraction does not affect fine fraction element measurements if the fine fraction is greater than 5% (Cauwet, 1987), then the analyses reported here do measure the element concentrations, even if diagenesis and metamorphism did some small-scale redistribution.

Plotting profiles of the element/Al ratios identifies anomalies due to biology or geochemistry as deviations from this background value. Major, minor and trace element analyses and ratios are compared with those from Recent sediments in the Japan Sea where possible, specifically around the Ulleung Basin. Unfortunately, no available comprehensive published geochemical analyses are available for sediments in the Japan Sea, nor are there reference to supplementary data tables in papers that subject the data to statistical manipulation; and it has not been possible to get data from relevant Institutions. Specifically there are no analyses of carbonate-rich sediments available, even from other SE Asian seas; so there is no comparison with the few carbonate-rich Guryul analyses, except for the inner shelf sample from offshore Korea with ~12% carbonate (Tables S2 and S3).

6.1 Data Availability Statement

Hand and some thin section descriptions of the sediments analyzed are on Table S1. Major and trace element analyses, and major and trace element correlation coefficients are on Tables S2–S4. Element/aluminium ratios and other element/element ratio figures, which are only partly discussed in this paper, are on Figures S1–S8. Complete spreadsheets with relative standard deviation (RSD) values are available from the first author.

7 RESULTS

The first question to ask is whether the results are true and accurate. Not all values are acceptable as measured by the RSD which is the absolute value of the coefficient of variation, here expressed as a percentage. Only those analyses with RSD’s of less than 10% were accepted. While most elements have RSD’s of less than 5% some elements, like Ta, Sb, and Be, and most of the Cd, Mo, Te, Hg and Platinum Group elements have few acceptable RSD values, or are below detection limits. Only a few samples were analyzed for P and S and reported by Algeo et al. (2007).

The major element abundances are determined by lithology (Figure 11), while the major element/Al ratios are more variable (Figure 12). Thus, Si is mostly low in calcareous beds and higher in argillites and sandstones where the values are comparable with the reference loess but mostly lower than the Ulleung sediments. The Si/Al ratio fluctuates inconsistently up the section and is generally lower than the Ulleung sediments, reflecting the lower quartz silt content of the Guryul sediments, although the limestone Si/Al values are often higher than in the shales, reflecting relatively higher amounts of quartz particles (Figure 12). Titanium is
low in the Zewan Formation sediments but rises to values comparable with, and higher than, the reference sediments in the Khunamuh Formation. The Ti/Al ratio rises through the lower Khunamuh and is generally higher than Ulleung sediments, but comparable with the loess. Iron, Mg and K tend to be higher in the argillites, and Mn (and of course Ca) higher in the limestones. The Fe/Al and Mn/Al ratios are much higher in the calcareous beds, and so Fe and Mn are not related to clastic but to carbonate content. The Mg/Al and K/Al ratios vary little and compare well with the Ulleung sediments, except that, in the limestones, Mg/Al is higher and K/Al lower. Sodium is very low except in the Zewan sandstones, while Na/Al is similarly comparable with the loess value in these sandstones, but lower than the Ulleung Basin sediments and very much lower in the Khunamuh argillites and limestones.

There is little consistent change in element or element/Al values, either across the Zewan/Khunamuh lithological boundary, or across the Permo-Triassic boundary, although Ti, Fe, Mg and K values increase, as do Ti/Al ratios. Sodium and Na/Al ratios decrease. Around the top of the Zewan Formation, Mn/Al, Ca/Al and Na/Al ratios reach limestone values in sandstones and argillites with very low Ca contents (Figures 11 and 12).

Similarly, changes in the trace element/Al values tend to correlate with lithology and compare with reference sediments where comparisons are available (Figures S1–S8).

The limited X-ray analyses of Algeo et al. (2007) show that quartz, calcite and clay minerals (illite with subordinate chlorite) dominate the mineralogy of the sediments, which also show marked separation into carbonate-rich (the limestones) and almost carbonate-free lithologies (the argillites). The metamorphism has obscured the original clay mineralogy of the sediments. But, by using the methods of Kackstaetter (2014) which calculate mineralogy and clay mineralogy from geochemical analyses, some idea of the original clay mineralogy can be worked out (Figure 13). The Zewan Formation is dominated by sericite with subordinate illite and small amounts of kaolinite and chlorite in lower to middle units. The Khunamuh Formation is dominated by kaolinite with subordinate illite but with smectite and chlorite in the middle part of the analyzed section—possibly due to volcanic input.
as some beds around the 10 m level contain identifiable volcanic lapilli (Table S1).

Two general REE patterns were determined based on the chondrite normalized REE spider diagrams. Each Guryul Ravine sample is assigned an REE pattern based on the spider diagram pattern and the \([\text{Eu}/\text{Eu}^*]_{\text{Chondrite}}\) ratio (Figure 14; Table S3). REE pattern 1, in general, has higher \([\Sigma \text{REE}]_{\text{Chondrite}}, [\text{La}/\text{Sm}]_{\text{Chondrite}}, \text{and [La/Yb]}_{\text{Chondrite}}\), and lower \([\text{Eu}/\text{Eu}^*]_{\text{Chondrite}}\). REE pattern 2 is the opposite with generally lower \([\Sigma \text{REE}]_{\text{Chondrite}}, [\text{La}/\text{Sm}]_{\text{Chondrite}}, \text{and [La/Yb]}_{\text{Chondrite}}\) and higher \([\text{Eu}/\text{Eu}^*]_{\text{Chondrite}}\). An interesting difference between the groups is the very negative Eu anomaly for REE pattern 1 with values ranging between 0.40 and 0.72 while REE pattern 2 is generally less negative with values ranging between 0.66 and 1.08. An upsection trend is seen in the REE patterns with the samples from beds A3 to E2/56 (the Zewan Formation and lower 6 m of the Khunamuh Formation) being dominated by REE pattern 1; then a transition to REE pattern 2 between E2/56 and E3/63 (6 and 11 m of the Khunamuh Formation). A second transition occurs between E3/63 and E3/67 (11 and 15 m of the Khunamuh Formation) back to the dominated REE pattern 1 before transitioning back to an REE pattern 2 at E3/67 for the last 5 m sampled of the Khunamuh Formation. Based on significant REE correlations with Al, and the distribution of the different REE pattern assignments for Guryul samples, the REE patterns are driven by lithological changes with REE pattern 2 corresponding to limestone lithologies.

Overall, there are no strong trends in either elements or element/Al ratios through the section, indicating no major changes in provenance. Sporadic large fluctuations are almost entirely due to lithology (e.g. Figure 12; Figures S1–S7, Tables S2 and S3).

8 | POST-DEPOSITIONAL ALTERATION

The Guryul sections has suffered severe post-depositional changes. The sediments are metamorphosed to sub-green-schist facies as all lithologies contain macroscopic chloritoid
(Figure 4C) and conodonts have colour indices of >4, both indicating temperatures >300°C (Kramm, 1973; Epstein et al., 1977; Savage and Bassett, 1985; Sweet, 1992; Königshof, 2003). Furthermore, the A-CN-K plot indicates significant K-metasomatism (see Section 9.1).

Nevertheless, the small-scale stratigraphic variability in most analyses and ratios suggests little migration and homogenization of the samples. Chemicals in fine-grained impermeable sediments do not move far (Mathieu et al., 2000). The precipitation of illite in the pore space of shales and mudstones greatly reduces their hydraulic conductivity, or permeability, by several orders of magnitude and produces closed system diagenesis for burial environments, particularly those >60°C (Nadeau, 2011). Trace elements can, however, be redistributed on at least the hand specimen scale in some shales (Lev et al., 1999), as trace elements migrate from the shales into the pore waters during or subsequent to illitization and related decomposition of organic matter (Hannigan and Basu, 1998). This is obvious from the macroscopic chloritoid porphyroblasts in many specimens. For carbonates, diagenesis is more open and compositions can be significantly altered (Reinhardt et al., 2000), for example the very high Sr/Al ratios in the Khunamuh limestones (Figure S4). Nevertheless, many of the more refractory elements, such as Zr and REE, change little even in the limestones.

9 | PROVENANCE AND CLIMATE

The composition of clastic sediments and rocks is controlled by; source rock composition, chemical weathering, abrasion, sorting during transport and diagenesis (Johnsson, 1993; Rollinson, 1993).

Although the presence of periodic large lakes in the late Permian need to be taken into account (Yemane and Kelts, 1990), in the hyper-arid to semi-arid climates, that prevailed in the Permo-Triassic tropical and temperate latitudes (Fluteau et al., 2001; Preto et al., 2010), chemical weathering processes are limited, so that petrography and heavy-mineral analyses can reconstruct sediment sources and dispersal paths up to thousands of kilometres away (Garzanti et al., 2012).

The ultimate source of the Guryul sediments must have been from the Andean mountain ranges along the southern edge of Gondwanaland via the African, Antarctic and India Permo-Triassic basins—distances of over 2,000 km (Figure 1). From the lithology and existing petrography, the Guryul Permo-Triassic sediments are moderately mature sandstones, mudstones and limestones deposited on a post-rift passive margin (Brookfield et al., 2003). But, apart from the topmost Zewan sandstones, which are arkoses with abundant plagioclase, indicating an arid source environment from
a cratonic basement block (Cox and Lowe, 1996), the grain sizes of the Guryul sediments are mostly too small to identify individual minerals and use petrography to infer provenance and climate.

Geochemistry of the fine fraction, however, can help determine these factors, since the proportion of coarse-grained sediment does not affect the distribution of elements in the fine fraction, if the fine fraction is greater than 5% (Cauwet, 1987), which is the case here, apart from the bioclastic limestones. Mudstones are the most useful because of their thorough homogenization before deposition, post-depositional impermeability and higher abundance of trace elements (Taylor and McLennan, 1985; Condie, 1993; McLennan et al., 1993; Bracciali et al., 2007).

9.1 | Major elements

Various indexes can be used to infer weathering, provenance and climate. The most popular and accepted is the chemical index of alteration (CIA) which is used to determine weathering (Nesbitt and Young, 1982; Bahlburg and Dobrinski, 2011).

\[
\text{CIA} = \frac{\text{Al}_2\text{O}_3}{(\text{Al}_2\text{O}_3 + \text{CaO} + \text{K}_2\text{O} + \text{Na}_2\text{O})} \times 100
\]

where CaO* is Ca exclusive of carbonates and the values are in molar proportions to emphasize mineralogical relationships.

The CIA values above 70 indicate a moderate to high degree of chemical weathering, while those below indicate a lower degree of chemical weathering (Fedo et al., 1995; Yan et al., 2010). The CIA, however, reflects the integrated weathering history in a drainage basin and is unreliable as a quantitative proxy for the intensity of chemical weathering (Li and Yang, 2010). Both weathering and sources initially determine the composition of clastic sediments deposits (Johnsson, 1993) and both thus affect the CIA (Fedo et al., 1995). In addition, since the CIA is based on the mobility of the major cations, it should be combined with ratios of immobile elements such as La, Th, Sc and Zr (Bhatia and Crook, 1986; McLennan et al., 1993) (see trace element Section 9.2 below). Another limitation in CIA use is post-depositional K addition to older clastic rocks, so a K-free index, the chemical index of weathering (CIW) can be used, also with moles (Harnois, 1988).

\[
\text{CIW} = \frac{\text{Al}_2\text{O}_3}{(\text{Al}_2\text{O}_3 + \text{CaO} + \text{Na}_2\text{O})} \times 100
\]

Both the CIA and the CIW indices reflect mostly the amount of feldspar relative to clay minerals, and therefore neither index is strongly sensitive to the type of parent rock. Furthermore, both require calculation of non-carbonate Ca. A better discrimination of parent rock types would include Fe and Mg.

The Index of Compositional Variability (ICV) includes Fe, Mg and Mn, does not require calculation of non-carbonate Ca, and uses oxides rather than moles (Cox et al., 1995).

\[
\text{ICV} = \left( \frac{\text{Fe}_2\text{O}_3 + \text{K}_2\text{O} + \text{Na}_2\text{O} + \text{CaO} + \text{MgO} + \text{MnO} + \text{TiO}_2}{\text{Al}_2\text{O}_3} \right)
\]

where \(\text{Fe}_2\text{O}_3(t)\) = total iron and CaO includes all sources of Ca.
Average basalt and average granite have very different ICV values of 2.20 and 0.95 (Li, 2000). Therefore, mudstones with the same degree of weathering (the same CIA) may have different ICV values, indicating composition of the source area; although high carbonate sediments give misleading values in all cases.

Guryul CIA, CIW, ICV values are compared on Figure 15. The carbonate-rich sediments are omitted except for ICV, which are shown to illustrate their unreliability, or, more accurately, the chemical instability of carbonates. The range of Guryul CIA values for samples with less than 2% of carbonate is from 62 to 75 (average 68) suggesting intermediate chemical alteration of the parent material (Borges et al., 2008) and are somewhat higher than the Ulleung Basin sediments. There is a slight variation in CIA from around 60 in the lower Zewan, rising to 60–70 in the top Zewan and Lower Khunamuh before dropping to 60 above about +15 m (Figure 15). Analyzed CIA (and CIW) values are similar to those calculated for some Zewan sandstones and Khunamuh shales by Akhtar et al. (2016). They are identical to the variation of the Chinese loess plateau sediments in the last 2.5 Ma (Chen et al., 2001). Contemporary Karoo late Permian palaeosols have less weathered CIA values of around 60 (Kus et al., 2017), while latest Permian shallow marine sandstone in the Salt Range have more highly weathered late Permian values of around 75, dropping to around 70 in the early Triassic, but showing the same trend as at Guryul (Hassan et al., 1999). These values are consistent with the Karoo soils and associated non-marine sediments being the source for both the Salt Range and Guryul argillites with the formers higher CIA than the references; but are markedly lower above the basal argillites in the Khunamuh Formation (Figure 12). The Na/Al ratios of the Zewan Formation sandstones are comparable with Ulleung Basin and loess values, but these ratios drop to very low values in the Khunamuh Formation (Figure 12). These criteria suggest lower rates of chemical weathering above the basal beds of the Khunamuh Formation, becoming more mature from the top Zewan into E2 and 3 of the Khunamuh before becoming less mature at the top of E3 and F.

The Ca/Al and Na/Al ratios in clastic sediments are indices of chemical weathering (Cox et al., 1995). Both are naturally high in the Guryul limestones. In the Guryul clastics, Ca/Al ratios in the Zewan Formation are compatible with, or higher than, the reference sediments; but are markedly lower above the basal argillites in the Khunamuh Formation (Figure 12). The Na/Al ratios of the Zewan Formation sandstones are comparable with Ulleung Basin and loess values, but these ratios drop to very low values in the Khunamuh Formation (Figure 12). These criteria suggest lower rates of chemical weathering above the basal beds of the Khunamuh Formation, becoming more mature from the top Zewan into E2 and 3 of the Khunamuh before becoming less mature at the top of E3 and F.

The K/Al ratio in sediments is generally used as an index of chemical maturity, but can also be used as an indicator of the original composition of sediments in the early stages of chemical weathering characterized by Ca and Na, because the K/Al ratio is markedly different in various feldspars, mica and clay minerals (Cox et al., 1995). K-feldspar is more resistant to chemical weathering than plagioclase, and during the weathering of K-bearing silicates, part of the K is tightly bound in the illite clay lattice, thus making it less mobile than Na in the weathering profiles. Furthermore, ancient fine-grained sediments with K/Al ratios less than ~0.35 point to minimal alkali feldspar abundance relative to other minerals in the original siliciclastics (Cox et al., 1995). Guryul K/Al ratios tend to be at or less than this (Figure 12), which fits with an average tonalite composition of the source. On the other hand, significant K metasomatism is obvious, and affects the inferred maturity of the sediments (Figure 16).

The Na/K, K/Fe and K/Fe + Mg ratios can similarly be used to infer source characteristics and weathering.
The Na and Na/K ratios reflect the maturity of clastic sediments. The Na/K ratios are used both to determine the degree of chemical weathering and the rate of sedimentation (Nicholls and Loring, 1960). Any Na/K ratios above 1 indicate immature sediments, while those below 1 indicate more mature sediments, due to the greater chemical rate
of weathering of plagioclase feldspar. The Na/K ratios of the lower Zewan sediments are comparable with the reference Ulleung sediments, but the top Zewan sandstones and Khunamuh argillites are lower, except for some calcareous beds near the top which approach 1 (Figure 17). The K‐metasomatism will, however, make all these N/K ratios lower. Despite this, the relative changes in Na/K indicate that the sediments become more mature from the top Zewan into E2 and E3 of the Khunamuh before becoming less mature at the top of E3 and F.

The K/Fe and K/(Fe + Mg) ratios reflect the contribution of felsic versus mafic rocks and the rate of chemical weathering of K‐felspar (Nesbitt et al., 1997). In areas of diverse rock sources like the present Atlantic margins, relatively low K/Fe values of tropical areas (0.26) contrast with higher K/Fe values (0.43) for arid regions (Govin et al., 2012). Guryul K/Fe and K/(Fe + Mg) ratios fluctuate over several orders of magnitude independently of lithology, with the most pronounced peaks at +6.4, +9.3, +13.7, and +17.7 m, and with many values above 0.4 indicating arid conditions (Figure 17). Although both these ratios are skewed by the K‐metasomatism, the fluctuations, in the argillites especially, like other fluctuating element ratios may be caused by alternating cooler and warmer phases since the sediment maturity does not change much during this time (Figure 15).

Titanium is insoluble, is concentrated in heavy minerals, like ilmenite (FeTiO₃) and rutile (TiO₂), and thus can be used as an indicator of detrital input (Calvert and Pederson, 2007). Because Ti content may be quite variable among different rocks, even when Al content remains relatively constant, the Ti/Al ratio is very useful in provenance identification for various sediments (Li, 2000; Sheldon and Tabor, 2009). Guryul Ti values are generally around Ulleung Basin values for the Zewan Formation but progressively increase in the lower 5 m of the Khunamuh Formation before fluctuating from average to very high values in the argillites but not the limestones above that (Figure 11). The Ti/Al ratios for Guryul increase through the topmost Zewan and lowermost Khunamuh formations from ~0.06, characteristic of the lower Zewan and Ulleung basin sediments, to the much higher values of ~0.1, characteristic of the Khunamuh Formation and loess (Figure 12). Such high Ti/Al ratios occur in sediments eroded from bare desert and proglacial plains by wind where physical destruction of Ti‐bearing minerals, such as ilmenite and rutile increases (Chen et al., 2013). Both ilmenite and rutile are softer than quartz (7) and zircon (7.5) and can break down faster to provide Ti‐rich dust (Anand and Gilkes, 1984) (Table 1). The Zewan Ti/Al ratios are close to those of the contemporary alluvial sediments of the continental Karoo Basin, and the near shore marine sediments of the Salt Range, where there is little or no change in Ti/Al (0.05–0.07) across the Permo‐Triassic boundary (data from Hassan et al., 1999; Coney et al., 2007).

In the Arabian Gulf, wind supplies dust enriched in Ti, Fe and Mg from Arabian deserts to offshore sediments during the summer monsoon. High Ti/Al, Fe/Al and Mg/Al ratios in these sediments correlate with drier conditions during stadials (Sirocko et al., 2000). In the Khunamuh Formation, high Ti/Al ratios correspond with high Fe/Al and Mg/Al ratios, which suggests increased wind supply at these levels (Figure 12).

The major elements and their ratios to Al thus indicate a dominantly continental source but with increasing input from basic rocks in the Khunamuh Formation, where significant amounts of sediment were supplied by wind, based on Ti/Al, Fe/Al and Mg/Al ratios in these sediments correlate with drier conditions during stadials (Sirocko et al., 2000). In the Khunamuh Formation, high Ti/Al ratios correspond with high Fe/Al and Mg/Al ratios, which suggests increased wind supply at these levels (Figure 12).

The major elements and their ratios to Al thus indicate a dominantly continental source but with increasing input from basic rocks in the Khunamuh Formation, where significant amounts of sediment were supplied by wind, based on Ti/Al, Fe/Al and Mg/Al ratios, possibly due to a very arid phase in the earliest Triassic, or simply to the loss of vegetation and hence increased soil erosion at this time.

9.2 | Trace elements

Trace elements and trace element/Al ratios reflect source characteristic as well as weathering and transportation (Holland, 1984; Piper and Calvert, 2011).

Guryul trace element/Al and some element/element ratios are on Figures S1–S7. Most Guryul trace element/Al ratios show little systematic stratigraphic variation, although a few (B/Al, Sc/Al, Sr/Al) increase and others (Y/Al, As/Al) decrease from the Zewan into the Khunamuh Formations (Figures S1–S7). Some (Li/Al, Co/Al, Ni/Al, Zn/Al, As/Al, Se/Al, Sr/Al, Y/Al, Sn/Al, Pb/Al, Th/Al, U/Al), with high
Ca/Al and thus greater amounts of carbonate relative to the reference sediments, increase by up to several orders of magnitude at several levels (at −4 to 0 m, at +7 to +10 m, and at +16 to +20 m); while others (B/Al, Ga/Al, Ge/Al, Tb/Al, Zr/Al, Sb/Al, Ba/Al) decrease at those levels (Figures S1–S7).

Elements like Zr, Nb, La, Hf, Th and U are enriched in felsic rocks, whereas Sc, V, Cr, Co, Cu, Ni and Zn, are more concentrated in mafic rocks due to the much smaller size of their ions allowing them to enter early crystallizing pyroxenes (Schock, 1975; Taylor and McLennan, 1985; Condie

**FIGURE 17** Plots of Guryul Na/K, K/Fe and K/(Fe + Mg) plots. Symbols and notation as Figure 11. Note gradual drop in Na/K and fluctuating but generally increasing K/Fe and K/(Fe + Mg) in argillites from Zewan to Khunamuh Formations in argillite
and Wronkiewicz, 1990; Garver et al., 1996). Pyroxenes are more readily broken down chemically and physically than zircons, and Sc is concentrated in residual deposits, like bauxites during weathering (Das et al., 1971). Guryul Y/Al, Nb/Al, La/Al, Hf/Al, Th/Al and U/Al ratios generally decrease, Sc/Al increases, and Zr/Al, Cr/Al, Co/Al, Ni/Al, Cu/Al and Zn/Al show a slight increase (Figures S1, S2, S7). The actual ratios are mostly close to, or higher than, the reference sediments, particularly the loess. This suggests that the main source was continental upper crust, with a slightly increased input of basic rocks material in the early Triassic (the source of which may have been the Siberian Traps) and with a significant amount of wind transported material, as in the Ulleung Basin. The Cr/Al, Co/Al, Ni/Al, Cu/Al and Zn/Al ratios all increase across the Zewan/Khunamuh contact, which may mark an increase in basic-derived versus acid-derived material (Figure S2).

Ratios such as Th/Sc, Zr/Sc, Cr/V, Cr/Ni, Cr/Th and Co/Th along with REE distribution also help in provenance determination (Taylor and McLennan, 1985).

The Th/Sc ratio is used as a reliable provenance indicator, because both elements are insoluble under surface conditions, transported in detritus and therefore preserve the characteristics of their source (Taylor and McLennan, 1985). Thorium is enriched in felsic rocks, while Sc traces mafic source components: old upper continental crust Th/Sc ratios are ~1.0; recycled sediments are >1.0 and young undifferentiated arcs are <1.0 (McLennan et al., 1993). Zewan Guryul Th/Sc ratios are above 1, those of old upper continental crust, and drop to values of generally 1 and less in the Khunamuh Formation, below the Chinese loess sediment and Ulleung shelf sediment which suggests contributions from basic igneous sources (Figure 18).

The Zr/Sc ratio is a useful index of zircon enrichment (or depletion) as Zr is practically confined to zircon, whereas Sc is not, but preserves a provenance signature like REEs (McLennan et al., 1993). High Zr/Sc ratios indicate zircon concentration during sediment reworking, while low Zr/Sc ratios indicate removal of zircons in some way. Throughout the Guryul section, the Zr/Sc ratios are low compared to the only available loess reference sediment, due to Zr depletion (Figure 18). Scandium is mostly in the relatively soft mineral monazite (which also has cleavages) and is easily broken down to dust; whereas zircon remains in the physically resistant zircons (Table 1). Preferential abrasion of monazite compared with the harder zircon by wind action can explain the low ratios. The Ti/Zr ratio is also a good indicator of eolian transport as the softer Ti-bearing minerals get broken down to fine particles and concentrated in atmospheric dusts relative to the hard Zr-bearing minerals. This can be seen in the increase in Ti/Zr from the Chinese loess plateau (~10) to the Ulleung Basin (~20–40), where much of the sediment supply is from dust blown into the Japan Sea by the strong southwest (on the transposed view) winter winds (Figure 10). Guryul Ti/Zr ratios are generally much higher, although fluctuating greatly and sometimes very high in the limestones (Figure 18). The fine grain size of the sheetflood sandstones of the early Triassic in the Karoo Basin source (Pace et al., 2009) indicate water reworking of essentially coarse loess deposits under arid to hyper-arid conditions as in their western European northern latitude equivalents (Brookfield, 2004). A regional evaluation of such geochemistries is needed, however, as complexities of source, amount of chemical versus physical weathering, and reworking can greatly change original compositions (Moreno et al., 2006).

Similarly, the Cr/V and Cr/Ni ratios, if detrital, measure the enrichment of Cr over other ferromagnesian minerals (for which V and Ni are proxies) and shows the relative importance of ultrabasic sources of chromite (mainly mantle peridotites of ophiolites) in which Cr is concentrated (Dickey, 1975). Both ratios rise from the Zewan into the Khunamuh, which might show an increasingly ultrabasic ultimate source for the Khunamuh mudstones and limestones (Figure 18).

On the other hand, all three elements, V, Cr and Ni are partly controlled by primary productivity and redox conditions (Piper and Isaacs, 1996). Guryul Cr and V, however, show low correlation of 0.14 (Table S4) and two orders of magnitude differences among their ratios to TOC (Tables S2 and S3). Guryul Cr/V ratios tend to rise slightly in the Khunamuh Formation, but remain close to reference sediment values and do not exceed those (~1.5) associated with anoxic conditions (Schaller et al., 1997), except in Khunamuh E1 and E2 where other indications of anoxia are present (see Section 10.4). The Cr/Ni ratios of argillites, on the other hand, rise markedly to values much higher than reference sediments, and show orders of magnitude fluctuations, that may reflect eolian transport (Figure 18). Chromite, the main Cr-bearing

| Mineral     | Composition | Hardness | Cleavage |
|-------------|-------------|----------|----------|
| Xenotime    | YPO₄        | 4.5      | 2 perfect |
| Pentlandite | (Fe,Ni)S₂₈ | 3.5–4.0  | None     |
| Chromite    | (Fe,Mg)Cr₂O₄| 5.5      | None     |
| Monazite    | (REE,Th)PO₄ | 5–5.5    | 2 good/poor |
| Sphene      | CaTiSiO₅    | 5–5.5    | 2 poor   |
| Ilmenite    | FeTiO₃      | 5–6      | None     |
| Magnetite   | Fe₂O₄       | 5.5–6.5  | None     |
| Columbite   | Fe₂Nd₂O₆    | 6        | 1        |
| Rutile      | TiO₂        | 6–6.5    | 1        |
| Cassiterite | SnO₂        | 6–7      | 2 good   |
| Quartz      | SiO₂        | 7        | None     |
| Zircon      | ZrSiO₄      | 7.5      | None     |
mineral, is relatively soft (H = 5.5) and light (SG = 4.5–4.8), so, like ilmenite, it can be abraded and its Cr concentrated in dust. Pentlandite, the commonest Ni mineral is even softer (H = 3.5–4) but as light (SG = 4.5–5), so physical abrasion should also preferentially concentrate Ni in dust. Desert dusts should have lower Cr/Ni than aqueous sediments; as they do, values of 0.5 (Cr/Ni) are typical for Chinese desert-derived atmospheric dusts (Zhang et al., 2003), while equivalent continental shelf marine sediments have values around 3 (Zhao and Yan, 1992). The high, but fluctuating, Guryul Cr/Ni values, which are mostly above the reference sediment values (2–3), suggest high but variable wind supply if these ratios are controlled by transport (Figure 18).

Cheminically and physically resistant heavy minerals in which elements like Sc, Ga, Y, Zr, Nb, Hf, Th and REE tend to be concentrated (Bhatia and Crook, 1986; McLennan et al., 1993) are affected by differential physical and chemical breakdown and by hydraulic sorting (Hayashi et al., 1997; Armstrong-Altrin, 2009). Silicate dusts from desert basins with a long history of physical weathering and transport are depleted in some elements because the minerals containing them are resistant to both physical and chemical weathering, do not break down into small dust particles and get concentrated in the coarser sand fraction, which is left behind (Castillo et al., 2008).

The Y/Ni ratio measures a proxy for HREE elements (Y) against a proxy for ferromagnesium minerals (Ni). Yttrium in sediments is mostly in heavy resistate minerals, such as zircon, xenotime and garnet and should follow Zr, which is concentrated in zircons. Guryul Y/Al follows Zr/Al and Nb/Al values, and are similar to the only available loess value (Figure S5). The co-variance of Zr and Y indicates that heavy minerals with these elements were preferentially removed during transport and deposition. The Y/Ni ratio monitors a proxy for HREEs

![FIGURE 18 Plots of Th/Sc, Zr/Sc, Ti/Zr, Cr/V and Cr/Ni ratios for Guryul samples. Symbols and notation as Figure 11. Note: Drop in Th/Sc and increase in Cr/V and especially Cr/Ni for argillites across the Permo-Triassic boundary; Zr/Sc less than reference loess; Ti/Zr much higher than loess and comparable to, and higher than, Ulleung Basin sediments](image-url)
(Y) versus the general level of ferromagnesium trace elements (Ni). A plot of Cr/V versus Y/Ni shows the importance of ophiolitic versus continental sources. Guryul values tend to cluster in the granite and intermediate fields (Figure S8). Guryul Co/Th, Cr/Th and Cu/Zn ratios tend to rise above the Zewan Formation and are close to reference sediment values, where these are available, but then commence rapidly extreme fluctuations in the Khunamuh Formation indicating fluctuating sediment sources with a trend towards more basic sources in the Khunamuh Formation (Figure 19). Sediments derived from silicic sources have Cr/Th ratios from 4 to 15; while those derived from basic sources have ratios from 25 to 100 (Madhavaraju, 2015). On these criteria, the Zewan Formation is derived from continental crustal sources with an upward trend to more basic source input in the Khunamuh Formation.

The Rb/Sr ratio is also a proxy indicator of chemical weathering intensity (Dasch, 1969). Rubidium generally coexists with K in the K-rich minerals, such as K-feldspar, and biotite, etc.; while Sr tends to enrich in Ca-bearing minerals, such as limestones and the Ca-bearing silicates such as hornblende and plagioclase. Since Ca-bearing minerals are easier to breakdown chemically compared with K-bearing minerals, the K-Rb pair and Ca-Sr pair are easy to fractionate during natural processes. Chemical weathering can leach the Ca-Sr much more easily than K-Rb, which leads the residue enriched in K-Rb but depleted in Ca-Sr. The increase in chemical weathering intensity rapidly leaches Sr compared to Rb (Nesbitt and Young, 1982); and the Rb/Sr ratio increases with increasing chemical weathering and can thus be used to monitor the degree of source rock weathering (McLennan et al., 1993). In Chinese loess/paleosol profiles, higher Rb/Sr ratios correlate with degree of weathering, and thus to stronger monsoon precipitation on long-term timescales (Chen et al., 1999). The dissolved material, enriched in Sr but depleted in Rb, is eventually transported into lakes/oceans causing lower Rb/Sr ratios in lakes/oceans where sediments with higher fractions of chemical/biogenic deposits have lower Rb/Sr ratios. This can easily be seen in the Guryul Rb/Sr plot, where the carbonates have much lower Rb/Sr ratios than the argillites (Figure 19).

**FIGURE 19** Plots of Co/Th, Cr/Th, Cu/Zn, Rb/Sr, Nb/Ta for Guryul samples. Symbols and notations as Figure 11. Note increase in Co/Th, Cr/Th, Cu/Zn across Permo-Triassic boundary, and lack of consistent variation in Khunamuh Formation; but sporadic very high values in Nb/Ta.
addition, higher rates of chemical weathering can be seen in the argillites and this corresponds with higher CIA and CIW values (Figure 15).

Likewise, with increase in chemical weathering intensity, K will normally show depletion against Rb (Wronkiewicz and Condie, 1989), thus leading to a higher Rb/K ratio. Guryul Rb/K ratios are, however, amazingly constant between ~40 and 60, are comparable with Ullueng Basin sediments, and do not show the increased weathering of the Rb/Sr ratios (Figure S4).

Niobium and Ta are chemically very similar elements. Both are at trace levels in pyroxene, amphibole, biotite, rutile, ilmenite, sphene, cassiterite and zircon, and thus in heavy minerals (Parker and Fleischer, 1968). The Guryul Nb/Ta ratios are generally below 20 and compatible with reference sediments, but there are a few very high values above 50 in Khunamuh E2 and 3, which are superchondritic (Figure 19). Such high Nb/Ta values in marine sediments are only found in deep-ocean ferromanganese crusts (Hein and Koschinsky, 2014). Although the significance of such Nb/Ta fluctuations is poorly known (Green, 1995), such high Nb/Ta values can also come from weathering of carbonatite intrusions of which South Africa has many (Pfänder et al., 2012).

The REE pattern of fine-grained siliciclastic sediments and some elemental ratios, especially Eu/Eu*, reflect the composition of the source area (McLennan et al., 1993; Mongelli, 2004). Sediments derived from quartzose recycled sources and deposited within intracratonic basins or passive margins are strongly enriched in the light REE, have relatively modest middle to heavy-REE slopes, and often show REE curves that represent the composition of their terrigenous source rocks (Xu et al., 2011). Large negative Eu anomalies are common, indicating mechanical fractionation of plagioclase out of the system during weathering and transport (McLennan et al., 1993).

Guryul total REE vary from 52 to 328 (Table S3) and plots are mostly flat with increasing LREE enrichment (Figure 14). The HREE are depleted which is typical of shales that have undergone heavy mineral separation and loss—additional evidence for eolian abrasion and sorting at source. The La/Sc, La/Sm, La/Th and La/Yb ratios show relative proportion of felsic and mafic sources (Taylor and McLennan, 1985). Guryul ratios are controlled by lithology, and show little change up section, except that decreasing La/Sc values across the Permian–Triassic boundary indicate greater basic source material (Figure S7).

Climate change in sources regions, without source composition changes, can be determined by variations in original clay mineralogy and fluctuations in Ti/Al, Ti/K and Ti/Sc ratios. Among clay minerals, kaolinite is characteristic of intense tropical and desert weathering and, because of its distribution is referred to as the low-latitude clay mineral (Griffin et al., 1968). Kaolinite, however, need not indicate humid conditions at a contemporary source. For example, in the Sahara, kaolinite weathered from kaolinite-bearing Cenozoic rocks was stored in lake basins, river beds and soils during humid periods. During subsequent dry phases, fine-grained dust was mobilized from the desiccated lakes, rivers and soils resulting in maximum eolian uptake and transport of kaolinite into adjacent seas (Ehrmann et al., 2017). Sericite and kaolinite dominate the Zewan Formation, possibly due to hydrothermal alteration of sources by the Panjal Trap eruptions (Figure 13). Illite and kaolinite dominate the Khunamuh Formation (Figure 13) and indicate desert conditions in the source as Guryul was at temperate, not tropical latitudes around the Permian–Triassic boundary close to the polar limit of lowland deserts (Sarnthein et al., 1982). Guryul Ti/Al, Ti/K and Ti/Sc ratios increase from the Zewan into the Khunamuh Formation and then fluctuate somewhat at higher than reference sediment values—which suggests increased wind abrasion of suitable Ti-bearing heavy minerals (Figure 12; Table 1).

The depletion of hard abrasion-resistant heavy minerals in Guryul sediments can be clearly illustrated on a Th-Sc-Zr/10 plot, frequently used to infer tectonic settings (Bhatia and Crook, 1986). Such discrimination diagrams, however, work successfully less than 60% of the time (Armstrong-Altrin and Verma, 2005). This is the case for the Guryul sediments which plot far away from the passive margin field, towards the Th-Sc line and nowhere near any plausible tectonic fields, although the Luochan loess is not so depleted in Zr (Figure 20). The same is true for the Th-Co-Zr/10 plot.

9.3 | Provenance and climate summary

The palaeotectonic situation of the Guryul section, from lithology, petrology and geochemistry, is a passive margin. In these situations, much of the clastic sediment comes from continental basement and cover. The source of the Guryul sediments lay to the southwest, across the African and Indian continental basins, where the sediment ultimately derived from the Gondwana Andean magmatic arcs running along its southern edge. In the late Permian, Kashmir lay downwind of the mid-latitude westerlies which were continually reworking and transporting fine sediment from the semi-arid to arid plains of the Karoo and Indian basins into the southwestern Neotethys Ocean (Figure 1). The Zewan Formation contains well-sorted coarse arkosic sandstones with angular quartz sand (Figures 4B and 5C) laid down in shallow marine environments, whose immaturity confirms the desert conditions in the hinterland. The Khunamuh Formation shows no coarse clastic input and the fine sediments have many of the characteristics of the loess from the Chinese basins adjacent to the Japan Sea. The dominance of kaolinite may be due to wind reworking of kaolinite deposited during humid phases.
in these source basins. The greater hardness of heavy minerals, such as zircon, restrict mechanical breakage and thus concentrate High Field Strength Elements associated with them, such as Zr, Y, into eolian lag deposits, reducing the input of such elements into the adjacent basin.

Compared with the five modern geochemical sub-environments of the SW Japan Sea, the Guryul sediments resemble the inner shelf (type IV) and coastal and upper slope (type III) clusters, although the sediments are very variable, in their relatively low Mg, Ti, Fe, relatively high Cr, variable Zn, Co, and low Fe/Al ratios (Cha et al., 2007).

10 | PALAEOENVIRONMENTS FROM GEOCHEMICAL PROXIES

Selected elements and element ratios are useful as environmental proxies and can be used as geochemical proxies for some environmental variables, such as temperature (Mg/Ca, Sr/Ca, Li/Ca, and oxygen isotopes), salinity (B/Be, Li/Ca), productivity (Ba$_{bio}$, Cd/Ca) and redox conditions (V, Cr, Mo, U, Re, Cd, Re/Mo, etc.) both today and in the past (Geen et al., 1992; Morford and Emerson, 1999; Henderson, 2002; Wilde et al., 2004; Tribovillard et al., 2006; Calvert and Pedersen, 2007; Gallego-Torres et al., 2010).

10.1 | Temperature

The element ratios indicated above need analysis of individual shells and can not be done for the metamorphosed Guryul sediments, nor can oxygen isotopes be relied upon. Elsewhere, oxygen isotopes of conodonts indicate dramatic global warming in the latest Permian of tropical marine surface waters from possibly 20 to 35°C with high temperatures persisting for at least 5 Ma into the Triassic (Sun et al., 2012; Chen et al., 2013).

10.2 | Salinity

Estimates of palaeosalinity from the tolerances of recorded fossils suggest that the mid-Zewan Formation (unit B) may have been brackish but the rest of the section, except the top of unit D, has undoubted normal marine biotas, although a more detailed palaeoecological investigation would be useful (Figure 8).

Boron and Li/Ca ratios have been used to infer palaeosalinities (Frederickson and Reynolds, 1960; Walker, 1968; Furst, 1981; Dominik and Stanley, 1993). Boron in the clay mineral size fraction increases with salinity at the time of deposition (Frederickson and Reynolds, 1960). At Guryul, Boron is strongly associated with Al ($r = 0.940$) and K ($r = 0.960$; Table S4), and the B/Al ratios of many Guryul argillites, carbonates and sandstones are an order of magnitude greater than the Chinese loess, the only reference sediment comparison available (Figure S1). Nevertheless, there are lower B/Al ratios at various places both in argillites and limestones (e.g., at ~0 m, +10 m) which may be due to lower salinity. There is no relationship between B and TOC: the Guryul B/TOC ratios range from 16 to 1,200 (Tables S1, S2). On the other hand, low B/Al ratios in clays also correspond with lower pH (Hingston, 1964) which is interesting in view of the possible ocean acidification at the Permo-Triassic boundary (Clarkson et al., 2015).

Li/Ca ratios increase with increasing salinity, opposite to the temperature effect (Marriott et al., 2004; Ourbak et al., 2006). The Guryul Li/Ca ratios of the sandstones in the abnormal salinity unit D are higher than those in the Khunamuh Formation limestones (Figure S9), which may be a salinity effect, as the Nanpanjiang Basin temperatures are lower at this horizon which should cause an increase in Li/Ca ratios (Sun et al., 2012).

10.3 | Productivity

Productivity can be estimated from TOC, P and biogenic Se and Ba proxies (Schoepfer et al., 2015), although all have their problems (Anderson and Winckler, 2005). Productivity can be estimated from phosphorous content directly, but is not usually preserved well in sediments. The few available Guryul values are low, mostly less than 1%. Although possibly caused by metamorphism, the low values (<1%) of both P$_2$O$_5$ and TOC, and the limited range of P$_2$O$_5$/TOC indicate that they are related and possibly original values (Table S2).
Selenium follows sulfur in its geochemistry and is concentrated in organic matter, and selenium maxima correlate with maxima of primary productivity in Recent environments (Cutter and Bruland, 1984). The Se/TOC ratio has thus been tried as an indicator of productivity in both Recent and ancient marine sediments, but with little success as many of the proposed interpretations of the Se/TOC ratios are speculative (Mitchell, 2011). The Se values for Guryul are constant at 1.5 ± 0.5 μg/g and comparable with those of Recent sediments (Table S3). In contrast, Guryul Se/TOC ratios are variable and very high at 2–20 due to very low TOC, compared with shale averages of 2, which suggests that significant amounts of TOC may have been lost by low-grade metamorphism in some of the Guryul sediments (Figure 21).

Marine barite (BaSO₄), the main carrier of particulate Ba in the water column, is related to the marine carbon export flux. Barite precipitation in the water column is associated with decaying organic matter, which results in a positive correlation between barite and excess Ba (the Ba not carried by detrital material), and excess Ba and carbon fluxes. Based on this, a positive correlation between excess Ba and carbon export was established in modern marine sediments and used to infer palaeoproductivity (Bains et al., 2000). The Ba/Al ratio in modern seas gives an estimate of palaeoproductivity if compared with the detrital Ba/Al ratio (McManus et al., 1998; Reitz et al., 2004; Liguori et al., 2016). Guryul Ba values and Ba/Al ratios vary a lot but are compatible with the Ullueng samples (Figure 21, Table S3). Estimates of productivity (Bₐᵢₒ) have been done for Permo-Triassic sections in many places and show an increase in Baₐᵢₒ from latest Permian into the early Triassic in most places, except for South China (Shen et al., 2015). For Guryul, there is no consistent trend in Shen et al.’s (2015) data, but the samples are mostly several metres apart and thus a random sampling of the sediments. Our Guryul calculations on more closely
spaced samples are more consistent with Shen et al.’s (2015) detailed Spiti section, which shows great and rapid fluctuations in latest Permian to early Triassic.

The $B_{\text{bio}}$ values presented here were calculated with the formula of Dymond et al. (1992):

$$B_{\text{bio}} = B_{\text{total}} - \left( A_{\text{total}} \times \frac{B_{\text{Al}}}{B_{\text{as}}} \right)$$  \hspace{1cm} (1)

where $B_{\text{bio}}$ is biogenic barium, $B_{\text{total}}$ and $A_{\text{total}}$ the measured amounts and $B_{\text{Al}}/B_{\text{as}}$ the detrital ratio. Several assumptions must be made if $B_{\text{bio}}$ is calculated from Ba and Al data (Paytan and Griffiths, 2007):

1. All Ba not associated with terrigenous aluminosilicates is related to carbon export.
2. All Al is terrigenous.
3. The water column is oxic.
4. The assumed Ba/Al ratio is representative for each sample.

The values of Ba/Al$_{\text{as}}$ are the critical factor. Most calculations of $B_{\text{bio}}$ from whole rock Ba and Al data are also based on assumptions about the average crustal Ba/Al ratio (Dymond et al., 1992). Although the modern Ba/Al$_{\text{as}}$ world average is ~0.0037 (Reitz et al., 2004), using a regional Ba/Al$_{\text{as}}$ ratio is essential (Pirrung et al., 2008). The Ba/Al$_{\text{as}}$ for Guryul can be estimated using a crossplot of Al and Ba, as done in Shen et al. (2015) producing a value of 0.0030 using the lowest argillite value (Figure 22). The Ba/Al$_{\text{as}}$ value estimated here for Guryul is lower compared to the 0.0041 estimate determined by Shen et al. (2015) based on a lower resolution Guryul dataset, in which the samples range from −8 m to +11 m at 2 m spacing except within 0.5 m of the 0 mark. If, however, the lowest two Guryul argillite samples #42 and #63 are excluded, the ratio is ~0.004 (Figure 22). Of course, a standard Ba/Al detrital ratio may not be appropriate for the entire section, despite the apparent lack of provenance changes. The low Ba/Al ratios of the carbonates may reflect lower productivity during deposition of carbonate and a few argillites and sandstones near the top of the Zewan Formation, but it may also reflect decreased Ba detrital input, especially as many of the carbonate samples are below the 0.003 line (Figure 22). Calculated $B_{\text{bio}}$ can be used to infer relative changes in palaeoproductivity (Dymond et al., 1992). The $B_{\text{bio}}$ values for Guryul vary greatly due to very low $B_{\text{bio}}$ values for limestone lithologies and thus inferred low palaeoproductivity during these intervals. In particular, both argillites and sandstones zigzag from high to low values near the top of the Zewan Formation, with number 8 (~−10.2 m), having the highest $B_{\text{bio}}$ (1.115) of the entire analyzed section (Figure 21). Otherwise, the excess Ba ($B_{\text{bio}}$) values tend to vary between 200 and 500 ppm, which, for the argillites, are comparable with the values for the Ulleung Basin sediments. There is a decreasing trend through the transition from unit D to E1 and then little change in the argillite values, but lower values for the limestones from E2 upwards (Figure 21). These values are compatible with palaeoproductivity calculations for the latest Permian of SW China of ~200 to 500 ppm. from radiolarian shell accumulations (Gu et al., 2007) and ~36 to 219 ppm. from Cu contents (Zhang et al., 2007).

Palaeoproductivity can also be estimated from Beryllium (Be) Thorium (Th), and Protactinium (Pa). Beryllium and Pa are more soluble than Th and can be advected by ocean currents (Kumar et al., 1995) to be removed in areas of high particle flux causing a positive correlation between Pa/Th, Be/Th and productivity (Henderson, 2002). Protactinium was not measured. In general, the limited Guryul Be/Th ratios (varying from 0.05 to 0.5) follow the pattern of $B_{\text{bio}}$, providing some confidence that both proxies may reflect relative changes in palaeoproductivity at Guryul (Figure 21). One important difference between the two proxies is the lack of Be/Th depletion in the limestones that is present in the $B_{\text{bio}}$ data.

In summary, relative palaeoproductivity changes based on $B_{\text{bio}}$ and Be/Th values broadly suggest a slight decrease in productivity across the Permo-Triassic boundary and at least two intervals based on $B_{\text{bio}}$ of lowered productivity during the Early Triassic. These results are also compatible with the palaeoproductivity calculations for the Opal Creek section in the subtropical eastern Panthalassic Ocean margin from selenium isotopes (Stücksen et al., 2015).

**FIGURE 22** Plot of Ba against Al of Guryul samples to determine detrital Ba/Al ratio (after Shen et al., 2015). Ba/Al detrital value is ~0.003
soluble under oxidizing conditions and less soluble under reducing conditions, resulting in authigenic enrichment in oxygen-depleted sediments (Wignall, 1994). Therefore, strong enrichments in the ratios of these elements to Al can be indicative of reducing conditions. For Guryul, V/Al and Cr/Al vary little through the section (Figure S2), but Mo/Al and U/Al vary markedly. The Mo/Al ratio tends to be lower than the lowest reference sediments (~0.2), except for high values greater than 0.1 at ~5.8 to ~0.1 m, 4.46 m and +7.3 m (Figure S5), but U/Al is very low for most Guryul samples (Figure S6), which affects the use of U in redox calculations.

Authigenic Th/U ratios are often used as an index of bottom water anoxia (Wignall and Myers, 1988; Wignall and Twitchett, 1996; Martin, 2004). Thorium is mostly in the detrital fraction, has only one redox state (Th4+), and its concentration in sediments is unaffected by redox conditions. Uranium, however, is also carried partly in solution as uranyl carbonate complexes, is adsorbed on organic matter, is a redox-sensitive metal, and is readily removed from seawater as insoluble U(VI) under reducing conditions, thus concentrating U relative to Th in anoxic facies (Bone et al., 2017). An increase in anoxic sedimentation reduces the concentration of U in seawater solution as more U is sequestered in organic-rich sediments. In carbonates, the U concentration is related to the U concentration of the seawater in which they are deposited, an increase in anoxic sedimentation results in an increase in the Th/U ratio of carbonate sediments. The authigenic U can be calculated if the detrital U is known. In ‘normal’ mudstones, where U is assumed to be entirely detrital, the Th/U ratio is 3; so, the detrital U can be estimated by dividing the Th value by 3 (Wignall and Myers, 1988). Subtracting the calculated detrital U from the measured U will give a value for authigenic U in the sample. Thus:

\[
\text{Th}/\text{U}_{\text{authigenic}} = \frac{\text{Th}_{\text{measured}}}{\text{U}_{\text{measured}} - \text{Th}_{\text{measured}}/3}
\]

The Th/Uauthigenic values were calculated for Guryul samples, but give mostly negative values (ranging from ~602 to +57) because the measured Th values (1–30 μg/g) are mostly high compared with relatively low Umeasured values (1–3 μg/g), giving mostly high negative Uauthigenic values. The Th/U ratios (1–15) are also variable, and somewhat independent of lithology, although Zewan sandstone and Khunamuh argillites tend to have higher values than the limestones (Figure 22; Table S3). Using the basic Th/U plot, and the dysoxic/oxic boundary at Th/U ~3 in shales, then not only were most of the Khunamuh argillites and some limestones in the dysoxic field, but so were almost all the Zewan sandstones, which have no other evidence for dysoxia (Figure 23). Disregarding the Zewan sandstones, there is a rapid shift towards dysoxia (just after a short oxic phase) above the Late Permian Event Horizon (LPEH) at Guryul as there is elsewhere (Brennecke et al., 2011). Using Th/U ratios as indicators of anoxia, however, is more complex than earlier studies indicated (Algeo et al., 2011), and the importance of organic matter in U geochemistry is only now being thoroughly investigated (Cumberland et al., 2016).

The elements Ni and Cu are associated with organic matter. If Ni and Cu are not scavenged by organic matter, then they are not enriched in sediments. High Ni and Cu indicate either high organic matter flux brought these elements to the sediment or the occurrence of reducing conditions that allowed them to be fixed within the sediment (Tribovillard et al., 2006). Importantly, high Ni and Cu persist even if the organic matter is later oxidized as long as reducing conditions allowed the conversion of Ni and Cu to sulphides and/or the incorporation into pyrite (Tribovillard et al., 2006). Guryul Ni values are comparable with reference sediments and tend to be higher in the sandstones and carbonates; in contrast, Cu is mostly lower than reference sediments and lowest in the carbonates (Table S3). The Ni/Al values are highest in the limestones, with the sandstone and argillite values straddling the equivalent reference sediment values (Figure S2). The Cu/Al values are much more variable, independent of lithology, and tend to plot around reference values even for limestone lithologies, although with a few high peaks above 10 in E3 of the Khunamuh Formation which do not correspond with Ni/Al peaks (Figure S2). The Ni/Al and Cu/Al patterns suggest low organic matter accumulation rates and/or a lack of reducing conditions to fix the Ni and Cu in the Guryul sediments. The Guryul TOC values are all less than 1% (Table S2).

The Ni/Co ratios are also used to infer palaeoxygenation; Ni/Co <5 indicate oxic, 5–7 indicate dysoxic, and >7 suboxic to anoxic conditions in the sediment (Jones and Manning, 1994). Guryul Ni/Co ratios are generally much higher than the reference material (Figure 23). The values are in the oxic field except in the fluctuating zone above +4 m when they rapidly switch between oxic-dysoxic and suboxic-anoxic values suggesting multiple rapid and transient changes in oxygen availability. Such rapid oxic/dysoxic/anoxic patterns are recognized elsewhere (Algeo et al., 2011).

The Cu/Zn ratios have been used as a redox proxy in the Baltic Sea where ratios increase to above 1 from oxic to anoxic conditions (Hallberg, 1976). Guryul Cu/Zn ratios are mostly below 1, and comparable with the reference sediments, but several are above 1 in argillites at +4.6, +11, +13.4 m (Figure 19). There are too few Mo values to calculate (Cu + Mo)/Zn ratios.

Similar to Ni/Co and Cu/Zn, Cr/V and V/(V + Ni) ratios can also be used to reconstruct palaeo-oxygenation; Cr/V values more than 0.5 indicate oxic conditions, values between 0.24 and 0.5 dysoxic, and less than 0.24 are suboxic/anoxic (Jones and Manning, 1994). All Guryul Cr/V ratios are in the oxic...
field according to this, and comparable with the reference sediments, except for GU −7.8 and GU −16.5 m that plot just in the dysoxic region (Figure 18). V/(V + Ni) ratios less than 0.46 indicate oxic conditions, 0.46–0.6 indicate dysoxic, 0.6–0.85 indicate anoxic, and >0.85 indicate euxinic (Jones and Manning, 1994; Rimmer, 2004). Based on these studies, nearly all of the Guryul V/(V + Ni) ratios plot in the anoxic or euxinic regions (Table S3), which is at odds with the other trace element redox proxies, which suggest mostly oxic conditions. A major difference between Ni and V deposition is that Ni relies on organic matter while V does not (Tribovillard et al., 2006). The low TOC contents of Guryul could be influencing Ni deposition relative to V, causing, in this case, inflated V/(V + Ni) that are unsuitable for reconstructing redox conditions at Guryul.

Cerium (Ce) varies with oxidation-reduction conditions (Elderfield and Greaves, 1982). The anomaly is based on the assumption of a “linear” decline in rare earth concentrations with an increase of atomic number when the elements are normalized to some standard, for example, chondrites. Cerium can deviate significantly from a line that declines toward the heavier rare earths creating positive or negative anomalies. The utility of the Ce anomaly (Ce/Ce*) as a palaeoxygenation proxy depends on whether the values are controlled by changes in lithology or seawater chemistry. The Guryul Ce anomaly contains a significant Pearson’s correlation coefficient with Al ($r = -0.421; p < .001$) but only accounts for a portion of the total variance ($R^2 = .190$) (Table S4). Assuming Ce/Ce* reflects changes in seawater chemistry for Guryul, Ce/Ce* fluctuates around 1.0 suggesting oxic conditions with no large negative or positive values or systematic changes up section (Figure 23).

Europium also varies with oxidation-reduction conditions. Negative Eu/Eu* values (i.e., depletion in Eu) reflect oxidation of Eu$^{2+}$ to Eu$^{3+}$ and loss of Eu from the sediments and thus oxic conditions (Hannigan et al., 2010). On this basis, all the Guryul sediments are oxic, which fits with the Ce anomalies (Figure 23). However, neither in the Quaternary anoxic Black Sea sediments, nor in the Devonian Marcellus Shale...
is there any consistent Eu anomaly. There are no significant depletions or enrichments of Eu\(^{2+}\) or Eu\(^{3+}\) within the oxic, suboxic and anoxic layers within the Black Sea water column (Schijf et al., 1994). The Marcellus Shale, while having some Ce anomalies consistent with anoxic conditions, has no Eu anomalies consistent with anoxic conditions (Abdalla, 2012). The Guryul Eu anomalies must therefore be disregarded as indicators of oxic-anoxic conditions.

Overall, the majority of redox sensitive trace elements and ratios suggest that the Guryul sediments were deposited under oxic conditions with little systematic change, at least at the sample spacing used here. At the most, a few ratios (Cu/Zn and Ni/Co) suggest only transient intervals of lowered oxygen availability in Khunamuh E2 and 3, but no long-term trend of reducing conditions. This agrees with the previous interpretation of Guryul Ravine by Algeo et al. (2007) which focused sampling around the Late Permian extinction interval. However, the oxic interpretation based on trace elements is at odds with work based on pyrite framboolid size distributions, which suggested anoxic conditions at +1 m and above (Wignall et al., 2005).

### 11 | COMPARISON WITH BARUS, SALT RANGE, SPITI AND SOUTHERN TIBET

There are sections close to the Guryul ravine section at Barus and Mandakhpal, and in the same structural unit (Figure 2). Barus, while similar to Guryul, has some interesting differences (Nakazawa et al., 1975). Neither Barus nor Mandakhpal have other published studies. The only other marine sections on the north Indian Palaeozoic passive margin outside Kashmir are in the Surghar Range, Salt Range, Ladakh, Spiti, Nepal and southern Tibet (Figure 24A). The Surghar and Salt Range are in two separate probably parutochthonous nappes on the western side of the Indian continent. The rest are on the allochthonous High Himalaya and Tethyan units, juxtaposed with non-marine Permian–Triassic sediments in the lesser Himalaya and Indian shield along the Main Central Thrust and associated faults. Since the Miocene, significant underthrusting and frontal erosion (of up to 1,000 km and more) has occurred along these faults (Brookfield, 1993; Hinsbergen et al., 2018). This means that the High Himalaya and Tethyan sections were deposited far from their current positions in outer shelf and deeper tectonic environments and that the entire inner shelf of the Himalayan passive margin has been removed, except in the Salt Range (Figure 24B). Plausible palaeoenvironmental reconstructions need to consider tectonic displacements.

The Barus section has a somewhat thicker Zewan Formation compared with Guryul. The main differences, however, are; the thicker E1 unit (6 m compared with 2 m) between the LPEH and the FAD of *H. parvus*, and the absence of hummocky cross-stratified bioclastic storm/tsunami beds (Figure 25). This may be due to the Guryul E1 unit being deposited in a broad deep shelf channel cut into the Zewan Formation during a latest Permian sea-level fall; such channels are common on all Recent continental shelves (Mellet et al., 2013). The reduced number of limestones above may be an artifact as the Barus section was studied in less detail than the Guryul section and so far has no geochemical analyses (Nakazawa et al., 1975).

The Salt Range has many studies (Kummel and Teichert, 1970; Pakistani-Japanese Research Group, 1985; Wignall and Hallam, 1993; Hermann et al., 2011, 2012) and one of the most thoroughly studied sections was used, the Nammal section in its western part, as an example of the stratigraphy (Figure 25). The Chhidru sandstones are very immature, tabular and festoon cross-bedded, and contain
a diverse and prolific marine fauna of bellerophontid gastropods, bryozoans, foraminifera, crinoids, echinoids, brachiopods and algae (Pakistan-Japanese Research Group, 1985; Wignall and Hallam, 1993). They are typical shallow shelf deposits. The topmost White Sandstone, however, is a very poorly fossiliferous, thinly laminated fine-grained felsparic sandstone (Kummel and Teichert, 1970), deposited in shallow subtidal to intertidal environments with freshwater influences (Mertmann, 2003): it is similar to the topmost Zewan sandstone at Guryul.

The upper Chhidru has some typical cool-water conodonts such as *Vjalovognathus* sp. and *Merrillina* sp., as in the topmost Zewan at Guryul (Shen et al., 2006; Brookfield and Sun, 2015). The overlying Katwai Member is a remarkably uniform dolomite and limestone unit with poorly preserved Induan brachiopods and ammonoids (Kummel and Teichert, 1970) resting on an erosion surface cut into the White Sandstone, with up to 0.15 m relief (Wignall and Hallam, 1993). At Nammal, the Katwai Member consists of a fining upwards 3 m thick calcareous sandstone unit that begins with a thin imper- sistent tabular cross-bedded grainstone with possibly re-worked Permian brachiopods and mudcracks. Above is a widespread intraclastic echinoderm grainstone with large rip-up clasts of micrite at discrete horizons, abraded, re-worked Permian brachiopods and foraminifera, and low-angle cross-beds and glauconite towards its top. The top unit is thin-bedded packstones and grainstones, sometimes crinoidal, with fish bones and phosphatic pebbles indicating condensation (Wignall and Hallam, 1993). The shallow water marine Katwai Member grades up into the Mittiwal Member, which consists of silty shales and interbedded limestones with ammonoids (Figure 25). The shales contain the pelagic bivalve, *Claraia*, and are finely laminated in thin section, with 1 mm thick couplets of marl and very
thin quartz silt/pyrite/kerogen-rich laminae. The thicker limestones are graded with large ammonoids concentrated toward their bases, probably deposited by turbidity currents (Wignall and Hallam, 1993). The Mittiwali Member is a deep-water oxygen-restricted shelf deposit similar to those accumulating in oxygen-minimum levels in modern oceans from 200 to 1,500 m depth (Paulmier and Ruiz-Pino, 2009).

The Upper Permian Wargal and Chhidru Formations are comparable with the Zewan Formation at Guryul, and correlate well on both sediment and biota. The Induan part of the Mianwali Formation is less than half as thick as at Guryul (Figure 25). It is tempting to correlate the Katwai Member with Guryul E1 bioclastic beds, as a nearer shore equivalent of these tsunami deposits, especially in view of the $\delta^{13}C_{\text{carb}}$ values of around 0 in both (Figure 25). But the Katwai Member contains the basal Triassic conodont, *H. parvus*, which makes it equivalent of basal E2 unit at Guryul, unless E1 is basal Triassic, which is possible as few conodonts have been found in E1, although it has not been thoroughly investigated (Brookfield and Sun, 2015; Brosse *et al*., 2017). The limited geochemical analyses of shales by Hassan *et al*., which did not include the Katwai Member, indicate, like Guryul, little change in CIA (72–76) up section from the Chhidru into the Mianwali, except for a slight decrease to 70 in the lowest Mittiwali Member (Hassan *et al*., 1999). Calculating ratios from Hassan *et al*.’s data show that Ti/Al ratios are much higher than at Guryul, a function of heavy mineral concentration in sandstones, but increase only slightly from 500 to 700 up section. Zr/Al values range sporadically from 11 to 52, tend to be higher than Guryul values, are in the range of reference sediments, and have the highest values at the top of the Mittiwali Member. The Ti/Zr values vary sporadically from 13 to 43, which are lower than most Guryul values, with the lowest values at the top of the Mittiwali Member. Both higher Zr/Al and lower Ti/Zr at Nammal show the greater importance of water transport in Salt Range Permian–Triassic sediments compared with Guryul, in keeping with their shallower water environment.
The Ladakh sections have not been studied much, and not at all geochemically.

The Spiti sections are perhaps the most thoroughly studied Permian–Triassic sections in the Himalaya (see Ghosh et al., 2016 for bibliography), but published geochemical studies are limited to a few metres above and below the Permian–Triassic boundary (Williams et al., 2012; Ghosh et al., 2016). The most thoroughly studied Mud section was used as an example (Figure 26). The late Permian Gungri Shale consists of black phosphatic silty shales with siltstone horizons containing sponge spicles, spiriferid brachiopods and crinoid ossicles and was deposited in a normal marine environment with high productivity, indicated by the high phosphorous content (Garzanti et al., 1996). The thin limonite layer at the Permian–Triassic boundary consists of small pebbles of micrite limestone and quartz sand in a limonite matrix—although the original was pyrite (Shukla et al., 2002). It marks an episode of reworking of the Gungri Shale under anoxic conditions and resembles other pyritic Permian–Triassic boundary layers elsewhere, for example, western Canada (Schöpfer et al., 2012). The overlying earliest Triassic Mikin Formation consists of grey limestone, marl and subordinate thin shale bands in various proportions with ammonoids and pelagic bivalves (Bhargava et al., 2004). These characters indicate an overall deeper shelf or continental slope depositional environment compared with the Guryul succession. Element ratios like Th/Sc show a continental crust source with no change in CIW indices up section (Williams et al., 2012). The Ti/Al ratios (−0.05 to 0.07) do not change from the Gungri Shale into the Mikin Formation, do not show the slight increase as at Guryul, and are at the lower end of the reference sediment values; Ti/Zr values (25–38) also vary little and unsystematically, with the variation due to relatively low Zr values as at Guryul (unpublished data of JW and AS). There is a sharp negative shift of −4‰, from +1 to −2‰ in the δ13Ccarb from the limonite layer into the Otoceras limestone above (Ghosh et al., 2002), comparable to the more gradual shift across E1 at Guryul, which fits the erosional reworking and lag concentration inferred for the limonite layer (Figure 26). The Muth section Griesbachian is highly condensed compared with Guryul.

The Nepal sections have no good published sedimentological or geochemical information (Waterhouse, 2004).

The southern Tibet sections are just north of the Nepal sections and in the same structural block as Nepal and Spiti (Figure 24). The Selong section, unlike some others, spans the Permian–Triassic boundary, and is reasonably well studied (Garzanti et al., 1998; Wignall and Newton, 2003; Yuan et al., 2018). It is unlike, and highly condensed compared with, the Guryul and Mud sections, but is similar to some of the Nepal sections (Waterhouse, 2004). The Upper Permian Qubueraga Formation consists of interbedded bioclastic limestones and calcareous shales of Changhsingian age with a coral bed at its top, capped by a caliche soil layer indicating emersion, overlain by a stromatolite layer (Wignall and Newton, 2003). The overlying Kangshare Formation begins with a crinoidal wackestone to packstone containing Permian brachiopods, overlain by well-bedded limestones with ammonoids, spanning the Griesbachian and Dienerian. Above are Smithonian to Spathian deep-water ammonoid-bearing nodular limestones and calcareous shale (Garzanti et al., 1998). The bioclastic limestones and caliche soil layer suggest shallow water and emersion at the end of the Permian, followed by subsidence and accumulation of deep-water limestones.

The time lost at the emersion surface seems relatively short as the Permian brachiopods above and below the surface are the same (Yuan et al., 2018). There is no published geochemistry on the Selong section. It possibly accumulated on an uplifted shelf-edge horst, and shows the same deepening trend across the Permian–Triassic boundary, with very shallow water environments at the boundary, as seen from the Salt Range through Kashmir to Spiti.

Oceanic Permian–Triassic sections occur in the Indus Suture zone to the north where a limestone exotic near Lamayuru, Ladakh, shows late Permian deposition on a shallow-water carbonate platform, followed by late Permian–early Triassic break-up, tilting, fissuring and subsidence, early Triassic mantling by ammonite-bearing, pink pelagic carbonate and chloride-rich volcanioclastic sediment, and finally deposition of FeO-rich and MnO-rich sediments, as crusts and fissure fills (Robertson, 1998).

## 12 | CONCLUSIONS

The Guryul Permian-Triassic sediments were deposited on a passive margin at a latitudes of about 40°S along the southern margin of the Neothyys Ocean. In such a palaeotectonic situation, much of the clastic sediment comes from continental basement and cover. The closest modern analogy for the Guryul section is the Japan Sea, which is in a similar tectonic and environmental situation to that inferred for the northern Indian Permio-Triassic sections (Sirocko et al., 2000; Cha et al., 2007; Chang et al., 2016). Both are “bathtub oceans” in which free connection with the larger ocean is limited by shallow water sills.

Many of the geochemical characters of the Guryul sediments can be explained if, like the Japan Sea, much of the clastic sediments were mainly supplied by the wind rather than by streams. For example, the common titanium-bearing minerals, titanomagnetite and ilmenite are softer than zircon and can be degraded by wind action so that eolian dust may have higher concentrations of Ti than Zr. In desert basins dust is influenced by a long history of transport, with physical
winnowing removing Zr, Hf and Th minerals in lags, and chemical leaching removing Na, K and Rb.

Major element variations are mostly subdued and take place across the change from the dominantly sandy shallow-water Zewan Formation into the dominantly shaly deeper-water Khunamuh Formation (where species diversity starts to decline) and not at the palaeontological Permo-Triassic boundary. A major change in inferred original clay mineralogy from sericite-dominated to kaolinite-dominated takes place at the same level. In the first 3 m of the Khunamuh Formation (with the Permo-Triassic boundary at 2 m), Ti, Ti/Al, Fe, Fe/Al, Mg, Mg/Al, K (but not K/Al) increase, and Na, Na/Al and Na/K decrease. Similar subdued changes take place in the trace element ratios, with small increases in Sc/Al, Cr/Ni, Co/Th, Cr/Th, Cu/Zn, Rb/Sr and decreases in As/Al, Th/Al, Th/Sc. Although there are some sporadic high values of element ratios, the overall geochemistry indicates that the Guryul sediments were derived from dominantly silica-rich continental rather than silica-poor sources although with some more silica-poor inputs in the Khunamuh Formation, possibly from the Siberian Traps, as the Panjal Traps were already buried by this time. The low High Field Strength Element (Zr, Nb, REE, Th, U, Ta) values and ratios to other elements indicate that the resistant minerals containing these elements were somehow removed prior to deposition, possibly by the heavy minerals containing these elements being left behind as eolian lags in source sediments—although the high Nb/Ta values in places are an enigma. The Ti/Al, Ti/K and Ti/Sc ratios increase from the Zewan into the Khunamuh Formation and then fluctuate somewhat at higher than reference sediment values, suggesting increasing aridity across the Permo-Triassic boundary with increased wind abrasion of suitable Ti-bearing heavy minerals accounting for both the high Ti/Al and Ti/Zr ratios. Various geochemical redox proxies suggest mainly oxic depositional conditions, with episodes of anoxia, but with little systematic variation across the Permo-Triassic extinction boundary at the sample spacing used. Geochemical proxies for productivity suggest a slight decrease across the Permo-Triassic boundary, and one or more decreased intervals in the Early Triassic.

The lack of consistent element geochemical changes across the Permo-Triassic boundary occur not only in adjacent shallower-water marine sections, such as the Salt Range, Pakistan (Hassan et al., 1999), but also in deeper shelf sections, such as at Spiti, India (Williams et al., 2012) and deep-water oceanic Panthalassic sections (Hori et al., 2007) (though short-lived excursions are common and still to be interpreted and correlated), and in South African and European continental sediments (Hiete et al., 2006; Coney et al., 2007). The environmental changes that caused the latest Permian extinction are not recorded in the elemental geochemistry of the sediments, but only in changes in the various isotopic systems, which reflect predominantly atmospheric and oceanic chemical and biological changes (Payne and Clapham, 2012). Some of the primary signatures of these come from carbon, nitrogen and sulphur isotopic changes (Baud et al., 1996; Kaiho et al., 2006; Riccardi et al., 2006; Korte and Kozur, 2010; Algeo et al., 2013; Saitoh et al., 2014; Zhang et al., 2017) which have yet to be integrated with the palaeontological, sedimentological and elemental geochemical changes already recorded from both marine and continental environments.

Future studies will involve more detailed comparisons of palaeoenvironmental changes during the Quaternary in the NW Pacific marginal seas with those of the Permian-Triassic boundary of the northeastern Gondwana margin, from the African and Indian non-marine basins across the Arabian, Indian and Australian passive margins into the deep-water oceanic section of the Neotethys along the Indus Suture Zone.

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CONFLICT OF INTEREST

The authors certify that they have no affiliations with or involvement in any organization or entity with any financial interest (such as honoraria; educational grants; participation in speakers’ bureaus; membership, employment, consultancies, stock ownership, or other equity interest; and expert testimony or patent-licensing arrangements), or non-financial interest (such as personal or professional relationships, affiliations, knowledge or beliefs) in the subject matter or materials discussed in this manuscript.

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REFERENCES

Abdalla, N.R. (2012) Rare earth elements as a Paleo-Ocean redox proxy within the union springs member of the marcellus formation.
BSc Thesis. Department of Geosciences, The Pennsylvania State University College of Earth and Mineral Sciences, 41 pp.

Akhtar, R.M., Balaram, V., Ganai, J.A., Dar, S.A. and Krishna, A.K. (2016) Geochemistry of sedimentary rocks from Permian-Triassic boundary sections of Tethys Himalaya: implications for paleo-weathering, provenance, and tectonic setting. Acta Geochimica, 55, 428–436.

Algeo, T.J., Hannigan, R., Rowe, H., Brookfield, M.E., Baud, A., Krystyn, L. et al. (2007) Sequencing events across the Permian-Triassic boundary, Guryul ravine (Kashmir, India). Palaeogeography, Palaeoclimatology, Palaeoecology, 252, 328–346.

Algeo, T.J., Kuwahara, K., Sano, H., Bates, S., Lyons, T., Elswick, E. et al. (2011) Spatial variation in sediment fluxes, redox conditions, and productivity in the Permian-Triassic Panthalassic Ocean. Palaeogeography, Palaeoclimatology, Palaeoecology, 308, 65–83. https://doi.org/10.1016/j.palaeo.2010.07.007.

Algeo, T.J., Henderson, C.M., Tong, J.N., Feng, Q.L., Yin, H.F. and Tyson, R.V. (2013) Plankton and productivity during the Permian-Triassic boundary crisis: an analysis of organic carbon fluxes. Global and Planetary Change, 105, 52–67.

Anand, R.R. and Gilkes, R.J. (1984) Weathering of ilmenite in a lateritic Pallid zone. Clays and Clay Minerals, 32, 363–374.

Anderson, R.F. and Winckler, G. (2005) Problems with paleoproductivity proxies. Palaeoceanography, 20, PA3012. doi:https://doi.org/10.1029/2004PA001107.

Anderson, C., Murray, R., Dunlea, A., Giosan, L., Kinsley, C., McGee, D. et al. (2019) Aeolian delivery to Ulleung Basin, Korea (Japan Sea), during development of the East Asian Monsoon through the last 12 Ma. Geological Magazine, 156, 1–12. https://doi.org/10.1017/S001675681900013X.

Armstrong-Altrin, J.S. (2009) Provenance of sands from Cazones, Acapulco, and Bahia Kino beaches, Mexico. Revista Mexicana de Ciencias Geológicas, 26, 764–782.

Armstrong-Altrin, J.S. and Verma, S.P. (2005) Critical evaluation of six tectonic setting discrimination diagrams using geochemical data of Neogene sediments from known tectonic settings. Sedimentary Geology, 177, 115–129. https://doi.org/10.1016/j.sedgeo.2005.02.004.

Bahk, J.J., Chough, S.K. and Han, S.J. (2000) Origin and paleoceanographic significance of laminated muds from the Ulleung Basin, East Sea (Sea of Japan). Marine Geology, 16, 459–477.

Bahlburg, H. and Dobrzinski, N. (2011) A review of the Chemical Index of Alteration (CIA) and its application to the study of Neoproterozoic glacial deposits and climate transitions. Geological Society of London Memoir, 36, 81–92.

Bains, S., Norris, R., Corfield, R. and Faul, K. (2000) Termination of global warmth at the Palaeocene/Eocene boundary through productivity feedback. Nature, 407, 171–174.

Baresel, B., Bucher, H., Brosse, M., Cordey, F., Guodun, K. and Schaltegger, U. (2017) Precise age for the Permian-Triassic boundary in South China from high-precision U-Pb geochronology and Bayesian age-depth modeling. Solid Earth, 8, 361–378.

Baud, A., Atudorei, V. and Sharp, Z. (1996) Late Permian and Early Triassic evolution of the northern Indian margin: carbon isotopes and sequence stratigraphy. Geodynamica Acta (Paris), 9, 57–77.

Bhargava, O.N., Krystyn, L., Balini, M., Lein, R. and Nicora, A. (2004) Revised litho- and sequence stratigraphy of the Spiti Triassic. Albertiana, 30, 21–39.

Bhatia, M.R. and Crook, K.A.W. (1986) Trace element characteristics of greywacke and tectonic setting discrimination of sedimentary basins. Contributions to Mineralogy and Petrology, 92, 181–193.

Bone, S.E., Dynes, J.J., Cliff, J. and Bargar, J.R. (2017) Uranium(IV) adsorption by natural organic matter in anoxic sediments. Proceedings of the National Academy of Sciences of the USA, 114(4), 711–716. https://doi.org/10.1073/pnas.1611918114.

Borges, J.B., Huh, Y., Moon, S. and Noh, H. (2008) Provenance and weathering control on river bed sediments of the eastern Tibetan Plateau and the Russian Far East. Chemical Geology, 254, 52–72.

Bracciali, L., Marroni, M., Pandolfi, L. and Rocchi, S. (2007) Geochemistry and petrography of Western Tethys Cretaceous sedimentary covers (Corsica and Northern Apennines): from source area to configuration of margins. Geological Society of America Special Paper, 420, 73–93.

Breeneck, G.A., Herrmann, A.D., Algeo, T.J. and Anbar, A.D. (2011) Rapid expansion of oceanic anoxia immediately before the end-Permian mass extinction. Proceedings of the National Academy of Sciences of the USA, 108(43), 17631–17634. https://doi.org/10.1073/pnas.1106039108.

Brookfield, M.E. (1993) The Himalayan passive margin from Precambrian to Cretaceous times. Sedimentary Geology, 84, 1–35.

Brookfield, M.E. (2004) The enigma of fine-grained alluvial basin fills: the Permo-Triassic (Cumbrian Coastal and Sherwood Sandstone Groups) of the Solway Basin, NW England and SW Scotland. International Journal of Earth Sciences, 93, 282–296.

Brookfield, M.E. and Sun, Y. (2015) Preliminary report of new conodont records from the Permian-Triassic boundary section at Guryul ravine, Kashmir, India. Permophiles, 61, 24–25.

Brookfield, M.E., Shellnutt, J.G., Yui, T.F. and Bhat, G.M. (in press). A 200 kyr eccentricity cycle across the Permo-Triassic boundary, Guryul Ravine, Kashmir, India. GeoScience Frontiers.

Brookfield, M.E., Twitchett, R.J. and Goodings, C. (2003) Palaeoenvironments of the Permian-Triassic transition sections in Kashmir, India. Palaeogeography, Palaeoclimatology, Palaeoecology, 198, 353–371.

Brookfield, M.E., Shellnutt, G., Qi, L., Hannigan, R., Bhat, M. and Wignall, P.B. (2010) Platinum element group variations at the Permo-Triassic boundary in Kashmir and British Columbia and their significance. Chemical Geology, 272, 12–19.

Brookfield, M.E., Algeo, T.J., Hannigan, R., Williams, J. and Bhat, G.M. (2013) Shaken and stirred: seismites and tsunamites at the Permian-Triassic boundary, Guryul ravine, Kashmir, India. Palaios, 28, 566–582.

Brosse, M., Baud, A., Bhat, G.M., Bucher, H., Leu, M., Vennemann, T. et al. (2017) Conodont-based Griesbachian biochronology of the Guryul Ravine section (basal Triassic, Kashmir, India). Geobios, 50, 359–387. https://doi.org/10.1016/j.geobios.2017.10.001.

Buoitso, L.A., Gningras, M.K., Macceachern, J., Mångano, M.G., Zonneveld, J.P., Pemberton, S.G. et al. (2005) Colonization of brackish-water systems through time: evidence from the trace-fossil record. Palaios, 20, 321–347.

Calvert, S. and Pedersen, T. (1993) Geochemistry of recent oxic and anoxic marine sediments: implications for the geological records. Marine Geology, 11, 67–88.

Calvert, S.E. and Pedersen, T.F. (2007) Element proxies for palaeoclimatic and palaeoceanographic variability. Developments in Marine Geology, 1, 567–644.

Castillo, S., Moreno, T., Querol, X., Alastuey, A., Cuevas, E., Herrmann, L. et al. (2008) Trace element variation in size-fractionated African desert dusts. Journal of Arid Environments, 72, 1034–1045.

Catuneanu, O., Wopfner, H., Eriksson, P.G., Cairncross, B., Rubidge, B.S., Smith, R.M.H. et al. (2005) The Karoo basins of south-central Africa. Journal of African Earth Sciences, 43, 211–253.
Cauwet, G. (1987) Influence of sedimentological features on the distribution of trace metals in marine sediments. *Marine Chemistry*, 22, 221–234.

Cha, H.J., Choi, M.S., Lee, C.B. and Shin, D.H. (2007) Geochemistry of surface sediments in the southwestern East/Japan Sea. *Journal of Asian Earth Sciences*, 29, 685–697.

Chang, K.I., Zhang, C.I., Park, C., Kang, D.J., Ju, S.J., Lee, S.H. *et al.* (2016) *Oceanography of the East Sea (Japan Sea)*. New York: Springer, 460 pp.

Chen, J., An, Z. and Head, J. (1999) Variation of Rb/Sr ratios in the loess-paleosol sequences of Central China during the last 130,000 years and their implications for monsoon paleoclimatology. *Quaternary Research*, 31, 215–219.

Chen, J., An, Z.S., Liu, L.W., Ji, J.F., Yang, J.D. and Chen, Y. (2001) Variations in chemical compositions of the eolian dust in Chinese Loess Plateau over the past 2.5 Ma and chemical weathering in the Asian inland. *Science in China (Series D)*, 44, 403–413.

Chen, B., Joachimski, M.M., Shen, S.Z., Lambert, L.L., Lai, X.L., Wang, X.D. *et al.* (2013) Permian ice volume and palaeoclimate history: oxygen isotope proxies revisited. *Gondwana Research*, 24, 77–99.

Choi, J., Lee, K.E. and Lee, H.J. (2012) Spatial and temporal changes in sea surface temperature, circulation and subpolar front of the East Sea (Japan Sea) during the last 130,000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 313–314, 225–233.

Chumakov, N.M. and Zharkov, M.A. (2003) Climate during the Permian-Triassic biosphere reorganization. Article 2. Climate of the Late Permian and Early Triassic: general inferences. *Stratigraphy and Geological Correlation*, 11, 361–375.

Clarkson, M.O., Kasemann, S.A., Wood, R., Lenton, T.M., Daines, S.J., Richoz, S. *et al.* (2015) Ocean acidification and the Permo-Triassic mass extinction. *Science*, 348, 229–232.

Condic, K.C. (1993) Chemical composition and evolution of the upper continental crust: contrasting results from surface samples and shales. *Chemical Geology*, 104, 1–37.

Condic, K.C. and Wronkiewicz, D.J. (1990) The Cr/Th ratio in Precambrian pelites from the Kaapvaal Craton as an index of craton evolution. *Earth & Planetary Science Letters*, 90, 256–267.

Coney, L., Reimold, W.U., Hancock, J.P., Mader, D., Koeberl, C., McDonald, I. *et al.* (2007) Geochemical and mineralogical investigation of the Permian-Triassic boundary in the continental realm of the southern Karoo Basin, South Africa. *Palaeoworld*, 16, 67–104.

Cox, R. and Lowe, D.R. (1996) Quantification of the effects of secondary matrix on the analysis of sandstone compositions, and a petrographic-chemical technique for retrieving original framework grain modes of altered sandstones. *Journal of Sedimentary Research*, 66, 548–558.

Cox, R., Lowe, D.R. and Bullers, R.L. (1995) The influence of sediment recycling and basement composition on evolution of mudrock chemistry in the southwestern United States. *Geochimica et Cosmochimica Acta*, 59, 2919–2940.

Cumberland, S.A., Douglas, G., Grice, K. and Moreau, J.W. (2016) Uranium mobility in organic matter-rich sediments: A review of geological and geochemical processes. *Earth-Science Reviews*, 159, 160–185. https://doi.org/10.1016/j.earscirev.2016.05.010.

Cutter, G.A. and Bruland, K.W. (1984) The marine biogeochemistry of selenium: a re-evaluation. *Limnology and Oceanography*, 29(6), 1179–1192.

Das, H.A., Zonderhuis, J. and van der Marel, H.W. (1971) Scandium in rocks, minerals and sediments and its relations to iron and aluminium. *Contributions to Mineralogy and Petrology*, 32, 231–244.

Dasch, E.J. (1969) Strontium isotopes in weathering profiles, deep-sea sediments, and sedimentary rocks. *Geochimica et Cosmochimica Acta*, 33, 1521–1552.

Dennison, J.M. (1972) Statistical meaning in geologic field work. *Geological Society of America Special Paper*, 146, 25–38.

Dickey, J.S. Jr (1975) A hypothesis of origin for podiform chromite deposits. *Geochimica et Cosmochimica Acta*, 39, 1061–1074.

DiPietro, J.A. and Pogue, D.R. (2004). Tectonostratigraphic subvisions of the Himalaya: a view from the west. *Tectonics*, 23, 20. doi:https://doi.org/10.1029/2003TC001554.

Dominik, J. and Stanley, D.J. (1993) Boron, beryllium and sulfur in Holocene sediments and peats of the Nile delta, Egypt: their use as indicators of salinity and climate. *Chemical Geology*, 104, 203–216.

Driscoll, F.C. (2016) Role of sorting on the composition of siliciclastic sediment: implications for interpreting provenance after limited transport in an arid climate. Master's Thesis, University of Tennessee, 107 pp. https://trace.tennessee.edu/utk_graddthes/4032

Dymond, J., Suess, E. and Lyle, M. (1992) Barium in deep-sea sediment: a geochemical proxy for paleoproductivity. *Paleoceanography*, 7, 163–181.

Eden, D.N., Wen, Q.Z., Hunt, J.L. and Whitten, J.S. (1994) Mineralogical and geochemical trends across the Loess Plateau, North China. *Catena*, 21, 73–90.

Ehrmann, W., Schmiedl, G., Beuscher, S. and Krüger, S. (2017) Intensity of African humid periods estimated from Saharan dust fluxes. *PLoS ONE*, 12(1), 18. https://doi.org/10.1371/journal.pone.0170989.

Elderfield, H. and Greaves, M.J. (1982) The rare earth elements in seawater. *Nature*, 296, 214–219.

Epstein, A.G., Epstein, J.B. and Harris, L.D. (1977) Conodont color alteration: an index to organic metamorphism. *U.S. Geological Survey Professional Paper* 995, 31 pp.

Erwin, D. (2006) *Extinction: How Life on Earth Nearly Ended 250 Million Years Ago*. Princeton, CA: Princeton University Press, 320 pp.

Fedo, C.M., Nesbitt, H.W. and Young, G.M. (1995) Unraveling the effects of potassium metasomatism in sedimentary rocks and paleosols, with implications for paleoweathering conditions and provenance. *Geology*, 23, 921–924.

Fluteau, F., Besse, J., Broutin, B. and Ramstein, G. (2001) The Late Permian climate. What can be inferred from climate modeling concerning Pangaea scenarios and Hercynian range altitude? *Palaeogeography, Palaeoclimatology, Palaeoecology*, 167, 39–71.

Frederickson, A.F. and Reynolds, R.C. (1960) Geochemical method of determining paleosalinity. *Clays and Clay Minerals*, 8, 203–213.

Furst, M.J. (1981) Boron in siliceous materials as a paleosalinity indicator. *Geochimica et Cosmochimica Acta*, 45, 1–13.

Gallego-Torres, D., Martinez-Ruis, F., De Lange, G.J., Jiminez-Espejo, F.J. and Ortega-Huertas, M. (2010) Trace-elemental derived paleoceanographic and paleoclimatic conditions for Pleistocene Mediterranean sapropels. *Paleoceanography, Palaeoclimatology, Palaeoecology*, 293, 76–89.

Garver, J.I., Royce, P.R. and Smick, T.A. (1996) Chromium and nickel in sea surface temperature, circulation and subpolar front of the East Asian inland. *Palaeoworld*, 6, 215–219.

Garzanti, E. and Alberto, R. (2016) Provenance control on chemical indices of weathering (Taiwan river sands). *Sedimentary Geology*, 336, 81–95.
Garzanti, E., Angiolini, L. and Sciunnach, D. (1996) The Permain Kulung Group (Spitit, Lahaul, and Zanskar; NW Himalaya): sedimentary evolution during rifting transition and initial opening of Neotethys. Rivista Italiana di Paleontologia e Stratigrafia, 102, 175–200.

Garzanti, E., Nicora, A. and Rettori, R. (1998) Permo‐Triassic boundary and Lower to Middle Triassic in South Tibet. Journal of Asian Earth Sciences, 16, 143–157.

Garzanti, E., Ando, S., Vezzoli, G., Lustrino, M., Boni, M. and Vermeesch, P. (2012) Petrology of the Namib Sand Sea: long-distance transport and compositional variability in the wind‐dispersed Orange Delta. Earth‐Science Reviews, 112, 171–189.

van Geen, A., Luoma, S.N., Fuller, C.C., Anima, R., Clifton, H.E. and Trumbore, S. (1992) Evidence from Cd/Ca ratios in foraminifera for greater upwelling off California 4,000 years ago. Nature, 358, 54–56.

van Insbergen, D.J.J., Lippert, P.C., Lie, S., Huang, W., Advokaata, E.L. and Spakman, W. (2018) Reconstructing Greater India: Paleogeographic, kinematic, and geodynamic perspectives. Tectonophysics, 760, 69–94. https://doi.org/10.1016/j.tecto.2018.04.006.

Ghosh, P., Bhattacharyya, S.K., Shukla, A.D., Shukla, P.M., Bhandari, N., Parthasarathy, G. et al. (2002) Negative δ13C excursion and anoxia at the Permo‐Triassic boundary in the Tethys Sea. Current Science, 83, 498–502.

Ghosh, N., Basu, A.R., Bhargava, O.N., Shukla, U.K., Ghatak, A., Garzione, C.N. et al. (2016) Catastrophic environmental transition at the Permain‐Triassic Neo‐Tethyan margin of Gondwanaland: Geochemical, isotopic and sedimentological evidence in the Sivp Valley, India. Gondwana Research, 34, 324–345. https://doi.org/10.1016/j.gr.2015.04.006.

Govin, A., Holzwarth, U., Heslop, D., Keeling, L.F., Zabel, M., Mulitza, S. et al. (2012) Distribution of major elements in Atlantic surface sediments (36°N–49°S): imprint of terrigenous input and continental weathering. Geochemistry, Geophysics, Geosystems, 13, Q01013. https://doi.org/10.1029/2011GC003785.

Govindaraju, K. (1994) 1994 compilation of working values and sample descriptions for 383 geostandards. Geostandards Newsletter, 18 (Suppl. S1), 1–158.

Green, T.H. (1995) Significance of Nb/Ta as an indicator of geochemical processes in the crust‐mantle system. Chemical Geology, 120, 347–359.

Griffin, J.J., Windom, H. and Goldberg, E.D. (1968) The distribution of clay minerals in the World Ocean. Deep Sea Research, 15, 433–459.

Gu, S.Z., Zhang, M.H., Gui, B.W. and Lu, X.D. (2007) An attempt to quantitatively reconstruct the paleo‐primary productivity by counting the radiolaria fossil in cherts from the latest Permian Dalong Valley, India. Acta Geochimica, 36(3), 412–416.

Guo, Y.L., Yang, S.Y., Li, C., Bi, L. and Zhao, Y. (2017) Sediment recycling and indication of weathering proxies. Acta Geochimica, 36(3), 498–501.

Hallberg, R.O. (1976) A geochemical method for investigation of paleoredox conditions in sediments. Ambio Special Report, 4, 139–147.

Hannigan, R.E. and Basu, A.R. (1998) Late diagenetic trace element remobilization in organic‐rich black shales of the Taconic foreland basin of Quebec, Ontario and New York. In: Schieber, J.W.Z. and Sethi, P.S. (Eds.) Sahel and Mudstone: Volume II, Petrography, Petrophysics, Geochemistry, and Economic Geology, Stuttgart: E. Schweizerbart'sche Verlagsbuchhandlung, pp. 209–233.

Hannigan, R., Dorval, E. and Jones, C. (2010) The rare earth element chemistry of estuarine surface sediments in the Chesapeake Bay. Chemical Geology, 272, 20–30. https://doi.org/10.1016/j.chemgeo.2010.01.009.

Harnois, L. (1988) The CIW index: a new chemical index of weathering. Sedimentary Geology, 55, 319–322.

Hassan, S., Ishiga, H., Roser, B.P., Dozen, K. and Naka, T. (1999) Geochemistry of Permain‐Triassic shales in the Salt Range, Pakistan: implications for provenance and tectonism at the Gondwana margin. Chemical Geology, 158, 293–314.

Hayashi, K.L., Fuyisawa, H., Holland, H.D. and Ohmoto, H. (1997) Geochemistry of approximately 1.9 Ga sedimentary rocks from northeastern Labrador, Canada. Geochimica et Cosmochimica Acta, 61, 4115–4137.

Hein, J.R. and Koschinsky, A. (2014). Deep‐ocean ferromanganese crusts and nodules. In: Holland, H.D. and Turekian, K.K. (Eds.), Treatise on Geochemistry, 2nd edition. Amsterdam: Elsevier, vol. 13, ch. 11. pp. 273–291.

Henderson, G.M. (2002) New oceanic proxies for paleoclimate. Earth and Planetary Science Letters, 203, 1–13.

Herrmann, E., Hochuli, P.A., Mêhay, S., Bucher, H., Brühwiler, T., Ware, D. et al. (2011) Organic matter and palaeoenvironmental signals during the Early Triassic biotic recovery: The Salt Range and Surghar Range records. Sedimentary Geology, 234, 19–41. https://doi.org/10.1016/j.sedgeo.2010.11.003.

Herrmann, E., Hochuli, P.A., Bucher, H., Brühwiler, T., Hautmann, M., Ware, D. et al. (2012) Climatic oscillations at the onset of the Mesozoic inferred from palynological records from the North Indian Margin. Journal of the Geological Society of London, 169, 227–237. https://doi.org/10.1144/0016‐74922010‐130.

Hiete, M., Berner, U., Heunisch, C. and Röhling, H.-G. (2006) A high‐resolution inorganic geochemical profile across the Zeichstein‐Buntsandstein boundary in the North German Basin. Zeitschrift der Deutschen Gesellschaft für Geowissenschaften, 157(1), 77–106.

Hingston, F.J. (1964) Reactions between boron and clays. Australian Journal of Soil Research, 2, 83–95.

Holland, H. (1984) The Chemical Evolution of the Ocean and Atmosphere. Princeton, CA: Princeton University Press, 582 pp.

Hori, R.S., Higuchi, Y., Fujiki, T., Maeda, T. and Ikehara, M. (2007) Geochemistry of the Orutamenu Formation, Arrow Rocks, Northland, New Zealand. In: Spörli, K.B., Takemura, A. and Hori, R.S. (Eds.) The Oceanic Permain/Triassic boundary sequence at Arrow Rocks (Orutamenu), Northland, New Zealand. GNS Science Monograph 24, 123–156.

Huang, Y., Chen, Z.-Q., Algeo, T.J., Zhao, L., Baud, A., Bhat, G.M. et al. (2019) Two-stage marine anoxia and biotic response during the Permain‐Triassic transition in Kashmir, northern India: pyrite framboid evidence. Global and Planetary Change, 172, 124–139.

Irino, T. and Tada, R. (2000) Quantification of aeolian dust (Kosa) contribution to the Japan Sea sediments and its variation during the last 200 ky. Chemical Geology, 34, 59–93.

Jha, N., Aggarwal, N. and Misra, S. (2018) A review of the palynosratigraphy of Gondwana sediments from the Godavari Graben, India: Global comparison and correlation of the Permain‐Triassic palynosfora. Journal of Asian Earth Sciences, 163, 1–21. https://doi.org/10.1016/j.jseaes.2018.05.014.

Johnsson, M.J. (1993) The system controlling the composition of clastic sediments. Geological Society of America Special Paper, 285, 1–19.
Jolivet, L., Tamaki, K. and Fournier, M. (1994) Japan Sea, opening history and mechanism: A synthesis. *Journal of Geophysical Research: Solid Earth*, 99, 22237–22259. https://doi.org/10.1029/93jb03463.

Jones, B. and Manning, D.A.C. (1994) Comparison of geochemical indices used for the interpretation of palaeoredox conditions in ancient mudstones. *Chemical Geology, 111*, 111–129.

Kackstaetter, U.W. (2014) SEDMIN — Microsoft Excel™ spreadsheet for calculating fine-grained sedimentary rock mineralogy from bulk geochemical analysis. *Central European Journal of Geosciences*, 6, 170–181.

Kaiho, K., Chen, Z.Q., Kawahata, H., Kajiwara, Y. and Sato, H. (2006) Conodont deformation patterns and textural alteration in the Ulleung Basin, East Sea (Sea of Japan). *Journal of Asian Earth Sciences*, 39, 396–405.

Kaneoka, I. (1986) Constraints on the time of the evolution of the Japan Sea Floor based on radiometric ages. *Journal of Geomagnetism and Geoelectricity*, 38, 475–485.

Kiehl, J.T. and Shields, C.A. (2005) Climate simulation of the latest Permian: implications for mass extinction. *Geology*, 33, 757–760.

Kim, H.Y. and Kim, D.C. (2001) Comparison and correlation of physical properties from the plain and slope sediments in the Ulleung Basin, East Sea (Sea of Japan). *Journal of Asian Earth Sciences*, 19, 669–681.

Königshof, P. (2003) Conodont deformation patterns and textural alteration in Paleozoic conodonts: examples from Germany and France. *Senckenbergiana Lethaea*, 83, 149–156.

Korte, C. and Kozur, H.W. (2010) Carbon-isotope stratigraphy across the Permian-Triassic boundary: a review. *Journal of Asian Earth Sciences*, 39, 215–235.

Kramm, U. (1973) Chloritoid stability in manganic rich low-grade metamorphic rocks, Venn-Stavelot Massif, Ardennes. *Contributions to Mineralogy and Petrology*, 41, 179–196.

Krystyn, L., Balini, M. and Nicora, A. (2004) Lower and Middle Triassic stage and substage boundaries in Spiti. *Albertiana*, 30, 40–53.

Kumar, N., Anderson, R.F., Mortlock, R.A., Froehlich, P.N., Kubik, P., Ditrich-Hannen, B. et al. (1995) Increased biological productivity and export production in the glacial Southern Ocean. *Nature*, 378, 675–680.

Kumar, K., Tewari, R., Agnihotri, D., Sharma, A., Pandita, S.K., Pillai, S.S.K. et al. (2017) Geochemistry of the Permian-Triassic sequences of the Guryul Ravine section, Jammu and Kashmir, India: implications for oceanic redox conditions. *GeoResearch Journal*, 13, 114–125.

Kummel, B. and Teichert, C. (1970) Stratigraphy and paleontology of the Wapadsberg Pass Area and implications for Late Permian climate. *Geological Society of America Abstracts with Programs*, 69(4), 111–127.

Liu, K.-K., Kang, C.K., Kobari, T., Liu, H., Rabouille, C. and Fennel, K. (2014) Biogeochemistry and ecosystems of continental margins in the western North Pacific Ocean and their interactions and responses to external forcing – an overview and synthesis. *Biogeosciences*, 11, 7061–7075. https://doi.org/10.5194/bg-11-7061-2014.

Liu, J., Li, J.J., Song, C.H., Yu, H., Peng, T.J., Hui, Z.C. et al. (2016) Palynological evidence for late Miocene stepwise aridification on the northeastern Tibetan Plateau. *Climate of the Past*, 12, 1473–1484. https://doi.org/10.5194/cp-12-1473-2016.

Lutaenko, K.A. (2012) Zonal-biogeographical analysis of bivalve moluscan faunas in three major areas of the western Sea of Japan. *Proceedings of the Russia-China Bilateral Symposium on Marine Ecosystems under the Global Change in the Northwestern Pacific*, Vladivostok, Russia, October 8–9, pp. 89–94.

Madhavaraju, J. (2015) Geochemistry of Late Cretaceous sedimentary rocks of the Cauvery Basin, South India: Constraints on palaeoweathering, provenance, and end Cretaceous environments. In: Ramkumar, M. (Ed.) *Geostratigraphy: Concepts, Techniques, and Applications*. Amsterdam: Elsevier, pp. 185–214.

Marriott, C.S., Henderson, G.M., Crompton, R., Staubwasser, M. and Shaw, S. (2004) Effect of mineralogy, salinity, and temperature on Li/Ca and Li isotope composition of calcium carbonate. *Chemical Geology*, 212, 5–15.

Martin, K.D. (2004) A re-evaluation of the relationship between trace fossils and anoxia. *Geological Society of London Special Publication*, 228, 141–156.

Mason, I.L. (1939) Studies on the fauna of an Algerian hot spring. In: *Interim Report of Sedimentary Boundary Problems: Permian and Triassic of West Pakistan*. In: Kummel, B. and Teichert, C. (Eds.) *Regional Oceanography. Part 1*. Leningrad: Gydrometeoizdat, pp. 291–463 (in Russian).

Lev, S.M., McLennan, S.M. and Hanson, G.N. (1999) Mineralogic controls on REE mobility during black-shale diagenesis. *Journal of Sedimentary Petrology*, 69, 1071–1082.

Li, Y.H. (2000) A Compendium of Geochemistry: From the Solar Nebula to the Human Brain. Princeton, CA: Princeton University Press, 475 pp.

Li, C. and Yang, S. (2010) Is chemical index of alteration (ClA) a reliable proxy for chemical weathering in global drainage basins. *American Journal of Science*, 310, 111–127.

Liguori, B.T.P., De Almeida, M.G. and De Rezende, C.E. (2016) Barium and its importance as an indicator of (paleo)productivity. *Annals of the Brazilian Academy of Sciences*, 88(4), 2093–2103.

Mazullo, J., Alexander, A., Tieh, T. and Menglin, D. (1992) The effects of wind transport on the shapes of quartz silt grains. *Journal of Sedimentary Research*, 62(6), 961–971. https://doi.org/10.1306/D4267A2A-2B26-11D7-8648000102C1865D.

McLennan, S.M., Hemmings, S., McDaniel, D.K. and Hanson, G.N. (1993) Geochemical approaches to sedimentation, provenance and tectonics. *Geological Society of America Special Paper*, 285, 21–40.

McManus, J., Berelson, W.M., Klinkhammer, G.P., Johnson, K.S., Coale, K.H., Anderson, R.F. et al. (1998) Geochemistry of barium in marine sediments: implications for its use as a paleoproxy. *Chemochimica et Cosmochimica Acta*, 62, 3453–3473.
Meltell, C.L., Hodgson, D.M., Plater, A.J., Mauz, B., Selby, I. and Lang, A. (2013) Denudation of the continental shelf between Britain and France at the glacial–interglacial timescale. Geomorphology, 203, 79–96. https://doi.org/10.1016/j.geomorph.2013.03.030.

Mertmann, D. (2003) Evolution of the marine Permian carbonate platform in the Salt Range (Pakistan). Palaeogeography, Palaeoclimatology, Palaeoecology, 191, 373–384. https://doi.org/10.1016/S0031-0182(02)00672-7.

Mir, A.R., Balaram, V., Ganai, J.V., Dar, S.A. and Krishna, A.K. (2016) Geochemistry of sedimentary rocks from Permian-Triassic boundary areas of Tethys Himalaya: implications for paleo-weathering, provenance, and tectonic setting. Acta Geochimica, 35(4), 428–436.

Mitchell, K. (2011) Selenium as Paleo-Oceanographic proxy: a first assessment. MSc thesis. Georgia Institute of Technology, School of Earth and Atmospheric Sciences. 58 p.

Mongelli, G. (2004) Rare-earth elements in Oligo-Miocenic pelitic sediments from Lagonegro basin, southern Apennines, Italy: implications for provenance and source-area weathering. International Journal of Earth Sciences, 93, 612–620.

Moore, B.R. and Dennen, W.H. (1970) A geochemical trend in siliconaluminum-iron ratios and the classification of clastic sediments. Journal of Sedimentary Petrology, 40, 1147–1162.

Moreno, T., Querol, X., Castillo, S., Alastuey, A., Cuevas, E., Herrmann, L. et al. (2006) Geochemical variations in aeolian mineral particles from the Sahara-Sahel dust corridor. Chemosphere, 65, 261–270.

Morford, J.L. and Emerson, S. (1999) The geochemistry of redox sensitive trace metals in sediments. Geochimica et Cosmochimica Acta, 63, 1735–1750.

Mundil, R., Ludwig, K.R., Metcalfe, I. and Renne, P.R. (2004) Age and timing of the Permian mass extinctions: U/Pb dating of closed-system zircons. Science, 305, 1760–1763. https://doi.org/10.1126/science.1101012.

Murray, J.W. and Alve, E. (2011) The distribution of agglutinated foraminifera in NW European seas: Baseline data for the interpretation of fossil assemblages. Palaeontologia Electronica, 14, 14.2.14A. https://palaeo-electronica.org/2011_2/248/index.html

Murray, J.W., Lee, B.S., Bullister, J. and Luther, G.W. (1999). The sulfide zone of the Black Sea. In: Beşiktepe, S.T., Uluatwa, U. and Bologa, A.S. (Eds.), Environmental Degradation of the Black Sea: Challenges and Remedies. NATO Science Series (2. Environmental Security), vol. 56. Dordrecht: Springer. pp. 75–91.

Nadeau, P.H. (2011) Earth’s energy “Golden Zone”: a synthesis from mineralogical research. Clay Minerals, 46, 1–24.

Nakazawa, K., Kapoor, H.M., Ishii, K., Bando, Y., Okimura, Y. and Tokuoka, T. (1975) The Upper Permian and Lower Triassic in Mandakpal Valley, Kashmir, India. Memoirs of the Faculty of Science, Kyoto University, Series Geology and Mineralogy, 42, 1–106.

Nesbitt, H.W. and Young, G.M. (1982) Early Proterozoic climates and plate motions inferred from major element chemistry of lutes. Nature, 299, 715–717.

Nesbitt, H.W., Fedo, C.M. and Young, G.M. (1997) Quartz and feldspar stability, steady and non-steady state weathering, and petrogenesis of siliciclastic sands and muds. Journal of Geology, 105, 173–191.

Nicholls, G.D. and Loring, D.H. (1960) Some chemical data on British Carboniferous sediments and their relationship to the clay mineralogy of these rocks. Mineralogical Magazine, 4, 196–207.

Nicholls, G.J. and Daly, M.C. (1989) Sedimentation in an intracratonic extensional basin: the Karoo of the Central Morondova Basin, Madagascar. Geological Magazine, 126, 339–354.

Nye, S., Brookfield, M.E. and Hannigan, R.E. (2016) Paleo-environmental changes leading to the Permian-Triassic transition in Mandakpal Valley, Kashmir, India. Geological Society of America, Abstracts with Programs, 48(7), 225–231.

Ourbak, T., Corrège, T., Malaize, T., Le Corneec, F., Charlier, K. and Peypouquet, J.P. (2006) A high-resolution investigation of temperature, salinity, and upwelling activity proxies in corals. Geochemistry, Geophysics, Geosystems, 7, Q03013. https://doi.org/10.1029/2005GC001064.

Pace, D.W., Gastaldo, R.A. and Neveling, J. (2009) Early Triassic aggradational and degradational landscapes of the Karoo Basin and evidence for climate oscillation following the P-Tr event. Journal of Sedimentary Research, 79, 316–331.

Pakistani-Japanese Research Group (1985) Permian and Triassic systems in the Salt Range and Surgar Range, Pakistan. In: Nakazawa, K. and Dickins, J.M. (Eds.) The Tethys: Her Paleogeography and Paleobiogeography from Paleozoic to Mesozoic. Tokyo: Tokai University Press, pp. 221–312.

Park, K.A., Lee, E.Y., Li, X., Chung, S.R., Sohn, E.H. and Hong, S. (2015) NOAA/AVHRR sea surface temperature accuracy in the East/Japan Sea. International Journal of Digital Earth, 8, 784–804.

Parker, R.L. and Fleischer, M. (1968) Geochemistry of Niobium and Tantalum. U.S. Geological Survey Professional Paper 612, 43 pp.

Parrish, J.T. (1993) Climate of the supercontinent Pangea. Journal of Geology, 101, 215–233.

Paulmier, A. and Ruiz-Pino, D. (2009) Oxygen minimum zones (OMZs) in the modern ocean. Progress in Oceanography, 80, 113–128. https://doi.org/10.1016/j.pocean.2008.08.001.

Payne, J.L. and Clapham, M.E. (2012) End-Permian mass extinction in the oceans: an ancient analog for the twenty-first century? Annual Review of Earth and Planetary Science, 40, 89–111.

Paytan, A. and Griffith, E.M. (2007) Marine barite: Recorder of variations in ocean export productivity. Deep-Sea Research II, 54, 687–705.

Peel, M.C., Finlayson, B.L. and McMahon, T.A. (2007). Updated world map of the Köppen-Geiger climate classification. Hydrology and Earth System Sciences Discussions, European Geosciences Union, 4(2), 439–473.

Pfänder, J.A., Jung, S., Münker, C., Stracke, A. and Mezger, K. (2012) A possible high Nb/Ta reservoir in the continental lithospheric mantle and consequences on the global Nb budget — evidence from continental basalts from Central Germany. Geochimica et Cosmochimica Acta, 77, 232–251.

Piper, D.Z. and Calvert, S.E. (2011) Holocene and late glacial palaeoceanography and palaeolimnology of the Black Sea: changing provenance and basin hydrography over the past 25,000 years. Geochimica et Cosmochimica Acta, 75, 5597–5624.

Piper, D.Z. and Isacs, C.M. (1996) Instability of bottom-water redox conditions during accumulation of Quaternary sediment in the Japan Sea. Geochimica et Cosmochimica Acta, 59, 1147–1162.

Pirring, M., Illner, P. and Matthiessen, J. (2008) Biogenic barium in the modern ocean. Geology, 36(2), 439–473.

Pfänder, J.A., Jung, S., Münker, C., Stracke, A. and Mezger, K. (2012) A possible high Nb/Ta reservoir in the continental lithospheric mantle and consequences on the global Nb budget — evidence from continental basalts from Central Germany. Geochimica et Cosmochimica Acta, 77, 232–251.

Piper, D.Z. and Isacs, C.M. (1996) Instability of bottom-water redox conditions during accumulation of Quaternary sediment in the Japan Sea. Geochimica et Cosmochimica Acta, 59, 1147–1162.

Pirrung, M., Illner, P. and Matthiessen, J. (2008) Biogenic barium in the modern ocean. Geology, 36(2), 439–473.

Preto, N., Kustatscher, E. and Wignall, P.B. (2010) Triassic climates—State of the art and perspectives. Palaeogeography, Palaeoclimatology, Palaeoecology, 290, 1–10.

Reinhardt, E.G., Cavazza, W., Patterson, R.T. and Blenkinsop, J. (2000) Differential diagenesis of sedimentary components and
the implication for strontium isotope analysis of carbonate rocks. Chemical Geology, 164, 331–343.
Reitz, A., Pfeifer, K., de Lange, G.J. and Klump, J. (2004) Biogenic barium and the detrital Ba/Al ratio: a comparison of their direct and indirect determination. Marine Geology, 204, 289–300.
Riccardi, A.L., Arthur, M.A. and Kump, L.R. (2006) Sulfor isotope evidence for chemocline upward excursions during the end-Permian mass extinction. Geochimica et Cosmochimica Acta, 70, 5740–5752.
Rimmer, S.M. (2004) Geochemical paleoredox indicators in Devonian-Mississippian black shales, Central Appalachian Basin (USA). Chemical Geology, 206, 373–391.
Riquier, L., Tribovillard, N., Averbuch, O., Devleeschouwer, X. and Riboulleau, A. (2006) Late Frasnian Kellwasser horizons of the Harz Mountains (Germany): two oxygen-deficient periods resulting from different mechanisms. Chemical Geology, 233, 137–155.
Robertson, A. (1998) Rift-related sedimentation and volcanism of the north-Indian margin inferred from a Permian-Triassic exotic block at Lamayuru, Indus suture zone (Ladakh Himalaya) and regional comparisons. Journal of Asian Earth Sciences, 16(2–3), 159–172. https://doi.org/10.1016/S0743-9547(98),00009-9.
Rollinson, H.R. (1993) Using Geochemical Data: Evaluation Presentation and Interpretation. Harlow: Longman, 352 pp.
Saitoh, M., Ueno, Y., Nishizawa, M., Isozaki, Y., Takai, K., Yao, J.X., et al. (2014) Nitrogen isotope chemostratigraphy across the Permian-Triassic boundary at Chaotian, Sichuan, South China. Journal of Asian Earth Sciences, 93, 113–128.
Sarnthein, M., Thiede, J., Pflaummann, U., Erlenkeuser, H., Fiitterer, D., Koopmann, B. et al. (1982) Atmospheric and oceanic circulation patterns off northwest Africa during the past 25 million years. In: von Rad, U., Hinz, K., Sarnthein, M. and Seibold, E. (Eds.) Geology of the Northwest African Continental Margin. Berlin, Heidelberg, New York: Springer-Verlag, pp. 545–603.
Savage, N.M. and Bassett, M.G. (1985) Caradoc-Ashgill conodont faunas from Wales and the Welsh borderland. Palaeontology, 28, 679–713.
Schaller, T., Moore, H.C. and Wehrli, B. (1997) Sedimentary profiles of Fe, Mn, V, Cr, As and Mo as indicators of benthic redox conditions in Baldeggersee. Aquatic Science, 59, 345–361.
Schiff, J., de Baar, H.J. and Millero, F.J. (1994) Biogenic barium and the detrital Ba/Al ratio: a comparison of their direct and indirect determination. Marine Geology, 204, 289–300.
Reitz, A., Pfeifer, K., de Lange, G.J. and Klump, J. (2004) Biogenic barium and the detrital Ba/Al ratio: a comparison of their direct and indirect determination. Marine Geology, 204, 289–300.
Riccardi, A.L., Arthur, M.A. and Kump, L.R. (2006) Sulfor isotope evidence for chemocline upward excursions during the end-Permian mass extinction. Geochimica et Cosmochimica Acta, 70, 5740–5752.
Rimmer, S.M. (2004) Geochemical paleoredox indicators in Devonian-Mississippian black shales, Central Appalachian Basin (USA). Chemical Geology, 206, 373–391.
Riquier, L., Tribovillard, N., Averbuch, O., Devleeschouwer, X. and Riboulleau, A. (2006) Late Frasnian Kellwasser horizons of the Harz Mountains (Germany): two oxygen-deficient periods resulting from different mechanisms. Chemical Geology, 233, 137–155.
Robertson, A. (1998) Rift-related sedimentation and volcanism of the north-Indian margin inferred from a Permian-Triassic exotic block at Lamayuru, Indus suture zone (Ladakh Himalaya) and regional comparisons. Journal of Asian Earth Sciences, 16(2–3), 159–172. https://doi.org/10.1016/S0743-9547(98),00009-9.
Rollinson, H.R. (1993) Using Geochemical Data: Evaluation Presentation and Interpretation. Harlow: Longman, 352 pp.
Saitoh, M., Ueno, Y., Nishizawa, M., Isozaki, Y., Takai, K., Yao, J.X., et al. (2014) Nitrogen isotope chemostratigraphy across the Permian-Triassic boundary at Chaotian, Sichuan, South China. Journal of Asian Earth Sciences, 93, 113–128.
Sarnthein, M., Thiede, J., Pflaummann, U., Erlenkeuser, H., Fiitterer, D., Koopmann, B. et al. (1982) Atmospheric and oceanic circulation patterns off northwest Africa during the past 25 million years. In: von Rad, U., Hinz, K., Sarnthein, M. and Seibold, E. (Eds.) Geology of the Northwest African Continental Margin. Berlin, Heidelberg, New York: Springer-Verlag, pp. 545–603.
Savage, N.M. and Bassett, M.G. (1985) Caradoc-Ashgill conodont faunas from Wales and the Welsh borderland. Palaeontology, 28, 679–713.
Schaller, T., Moore, H.C. and Wehrli, B. (1997) Sedimentary profiles of Fe, Mn, V, Cr, As and Mo as indicators of benthic redox conditions in Baldeggersee. Aquatic Science, 59, 345–361.
Schijf, J., de Baar, H.J. and Millero, F.J. (1994) Kinetics of Ce and Nd scavenging in Black Sea waters. Marine Chemistry, 46, 345–359.
Schock, H.H. (1975) Geochemistry and mineralogy. In: Horovitz, C.T. (Ed.) Scandium: Its Occurrence, Chemistry, Physics, Metallurgy, Biology and Technology. London: Academic Press, pp. 50–65.
Schoepfer, S.D., Henderson, C.M., Garrison, G.H. and Ward, P.D. (2012) Cessation of a productive coastal upwelling system in the Panthalassic Ocean at the Permian-Triassic Boundary. Palaeogeography, Palaeoclimatology, Palaeoecology, 313-314, 181–188. https://doi.org/10.1016/j.palaeo.2011.10.019.
Schoepfer, S.D., Shen, J., Wei, H.Y., Tyson, R.V., Ingall, E. and Algeo, T.J. (2015) TOC, organic P, and biogenic Ba accumulation rates as proxies for marine primary productivity and export flux. Earth-Science Review, 149, 23–52.
Sheldon, N.D. and Tabor, N.J. (2009) Quantitative paleoenvironmetal and paleoclimatic reconstruction using paleosoils. Earth-Science Reviews, 95, 1–52.
Shellnutt, J.G., Bhat, G.M., Brookfield, M.E. and Jahn, B.M. (2011) No link between the Panjal Traps (Kashmir) and the Late Permian mass extinctions. Geophysical Research Letters, 38, L19308. https://doi.org/10.1029/2011GL049032.
Zhao, Y.-Y. and Yan, M.-C. (1992) Abundance of chemical elements in sediments from the Huanghe River, the Changjiang River and the continental shelf of China. *Chinese Science Bulletin, 37*(23), 1991–1994. https://doi.org/10.1360/sb1992-37-23-1991.

**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section.