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Controls on sediment transfer dynamics at a tributary–trunk transition in the Little Karoo, with implications for interpreting the landscape response to environmental change

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ABSTRACT

Zones of alluviation at tributary – trunk confluences can act as sediment storage/transfer switches. Evaluating the temporal variation in tributary – trunk connectivity is key to understanding the origin, dynamics and residence time of tributary alluvial fill sequences, and determining relative and interacting effects of different drivers of landscape development. This paper evaluates processes and timescales of tributary (Prins River) – trunk (Touws River) connectivity at a site in the Little Karoo, as context for discussing sediment dispersal dynamics and implications for interpreting the landscape response to environmental change. An alluvial terrace in the tributary valley preserves a chronology (optically stimulated luminescence) of tributary valley alluviation that is regionally synchronous with valley alluviation in the upper Huis River and floodplain alluviation in the lower Touws and Groot rivers. A climatic shift within the Little Karoo at ~1000 years BP from relative aridity to relative humidity (and higher-energy rain-bearing circulation types) may have initiated widespread reworking of alluvial fills and the breaching of geomorphological buffers. Alternatively, there may be an intrinsic limit to sediment preservation potential associated with a regional floodplain cycling time of one to two thousand years. Longer archives are needed to contextualize fluvial responses to climatic variability in the region.

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Tributary-trunk relations; drylands; non-perennial rivers; fluvial sedimentology; luminescence dating

1. Introduction

Sediment connectivity (or (dis)connectivity (K. A. Fryirs, Brierley, Preston, & Spencer, 2007); describes the efficiency of sediment transfer between components of a geomorphic system (Wohl et al., 2019). In river catchments, this includes all structural linkages and functional transfer/exchange/storage processes between the point of sediment generation by detachment, and the point of sediment delivery to intermediary stores or more

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permanent intra-continental or oceanic sinks. Spatial and temporal variation in connectivity determines sediment supply- and transport-limitations that can manifest in complex ways to influence sediment delivery patterns at the catchment outlet (De Vente et al., 2007; K. A. Fryirs, 2013; Walling, 1983). Moreover, the variation in connectivity can shape the structure and function of catchment riverscapes (*sensu* Skidmore & Wheaton, 2022), thereby defining the diversity and configuration of fluvial styles (K. A. Fryirs & Brierley, 2013). Given its importance in understanding sediment dispersal, and logical congruence with ideas from hydrology and ecology (M. C. Grenfell et al., 2022; Turnbull & Wainwright, 2019), connectivity has become one of the most dominant conceptual frameworks for understanding earth surface processes in catchment systems (Parsons et al., 2015; Poepl et al., 2023; Wohl, 2017; Wohl et al., 2017).

Connectivity may be conceptualized in both structural and functional terms (see Wainwright et al., 2011; Wohl, 2017 for definitions). Structural connectivity refers to the spatial arrangement of landscape units, such as the degree of coupling between a hillslope and a channel (Harvey, 2001) or the catchment-scale pattern of hillslope coupling with the drainage network (Cavalli et al., 2013). The characterization and quantification of functional connectivity (dynamic linkages and fluxes between landscape units) is more complex (Poepl et al., 2023), as it demands insight into processes, rates and timescales of sediment generation, dispersal and storage. This is especially so for zones of confluence between tributary and trunk rivers, which are key structural elements in catchment-scale conceptualizations of sediment connectivity. From a functional perspective, these zones may act as sediment storage/transfer switches (*sensu* K. A. Fryirs, 2017; K. A. Fryirs, Brierley, Preston, & Kasai, 2007) that episodically activate ('turn on') and deactivate ('turn off'), such that the geomorphology of confluence zones of non-perennial rivers with highly variable flow and sediment transport regimes is quite distinct from that of perennial rivers with more steady flow and sediment transport (e.g., Best, 1988; Sambrook Smith et al., 2019).

Sediment transfer through non-perennial river confluence zones may be influenced by temporal variation in the relative dominance of tributary and trunk alluviation. Some systems are characterized by trunk valley impoundment through tributary fan deposition (e.g. McCarthy et al., 2011), while others are characterized by tributary valley impoundment through trunk floodplain alluviation that may occur progressively over centennial to millennial timescales (e.g., K. Fryirs & Gore, 2014; M. C. Grenfell et al., 2008; S. E. Grenfell et al., 2010) or rapidly in response to a single flood event (e.g., Baker et al., 1983). In cases where tributary impoundment occurs progressively over centennial and longer timescales, impoundment may not require wholesale damming of the tributary valley by a floodplain alluvial ridge (as in S. E. Grenfell et al., 2010) – a relatively minor sedimentary disconnect may be sufficient, which may be influenced by the position of the trunk channel within its floodplain in relation to the mouth of the tributary valley (Brierley & Fryirs, 1999; K. A. Fryirs, Brierley, Preston, & Spencer, 2007).; This is often the case in dryland non-perennial systems where tributaries lack the capacity to rework trunk buffers (S. E. Grenfell et al., 2014). The dynamics of impoundment may correlate with changes in trunk base level through the breaching of geological barriers (a lowering of local base level and increase in drainage network connectivity; Tooth et al., 2004) or valley alluviation with sea level rise (a rise in base level and decrease in network connectivity; S. E. Grenfell et al., 2010). Climatic forcing

may play a role in some environments (S. E. Grenfell et al., 2014; McCarthy et al., 2011), but this can be difficult to disentangle from intrinsic adjustments driven by the dynamics and persistence of alluviation at the confluence (K. A. Fryirs, Brierley, Preston, & Spencer, 2007; M. C. Grenfell et al., 2008), tributary responses to aggradation-driven steepening beyond a threshold of stability (Ellery et al., 2016), or an interplay of such drivers (Oldknow & Hooke, 2017).

Evaluating the temporal variation in tributary – trunk connectivity, the mechanisms of impoundment, and the underlying biophysical or human drivers (Poepl et al., 2017), is key to understanding the origin, dynamics and residence time of tributary alluvial fill sequences, which may host palustrine wetlands with stores of biogeochemically reactive elements such as carbon, nitrogen and phosphorous (S. Grenfell et al., 2019). The timescale of tributary – trunk connectivity is also the requisite context (Wohl, 2018) for interpretations of fluvial sedimentary archives of environmental change (Oldknow & Hooke, 2017; Walsh et al., 2023), while archive chronologies can, in turn, aid in the contextualization of catchment-scale connectivity over geomorphic timescales (K. Fryirs et al., 2022). Some recent progress has been made in this area and in the integration of analyses of structural and functional connectivity (Poepl et al., 2023). However, further work is needed in a broader array of catchment environments to advance understanding of the effects of environmental change on sediment cascades through catchment systems (K. A. Fryirs, Brierley, Preston, & Kasai, 2007), and to contextualize riverscape dynamics for improved catchment management in an uncertain future (Skidmore & Wheaton, 2022). Thus, the objectives of this paper are to i) evaluate tributary – trunk connectivity through geospatial modelling, empirical field and laboratory analyses of geomorphology, sedimentology and geochronology, ii) discuss the implications for understanding sediment dispersal dynamics in a dryland environment with highly variable flow and sediment dispersal patterns, and iii) discuss the implications for interpretation of the landscape response to environmental change.

2. Regional setting

2.1. *Catchment lithology and terrain*

The study site is located at the confluence of the Touws (trunk) and Prins (tributary) rivers (33.578874° S, 20.853710° E), ~80 km east-southeast of the town of Touwsrivier, in a region of the Western Cape Province of South Africa known as the Little Karoo (Figure 1). The Touws River rises near Touwsrivier at an elevation of 766 m amsl, is joined by the Prins River after ~91 km (down-valley distance) at an elevation of ~400 m amsl, and ends ~130 km down-valley to the east-southeast at its confluence with the Groot River at an elevation of 234 m amsl. The spatial distribution of lithology within the full catchment is illustrated in Figure 1 (main frame), which also shows the trunk and tributary sub-catchment boundaries upstream of the confluence. Mountains within the upper Touws sub-catchment are composed predominantly of Nardouw Subgroup (Table Mountain Group) quartz and feldspathic arenites, with lower relief dissected surfaces composed of Weltevrede Subgroup (Witteberg Group) quartz arenites, siltstones and finer-grained sedimentary rocks. Broad valley floors and flats are underlain by shales,

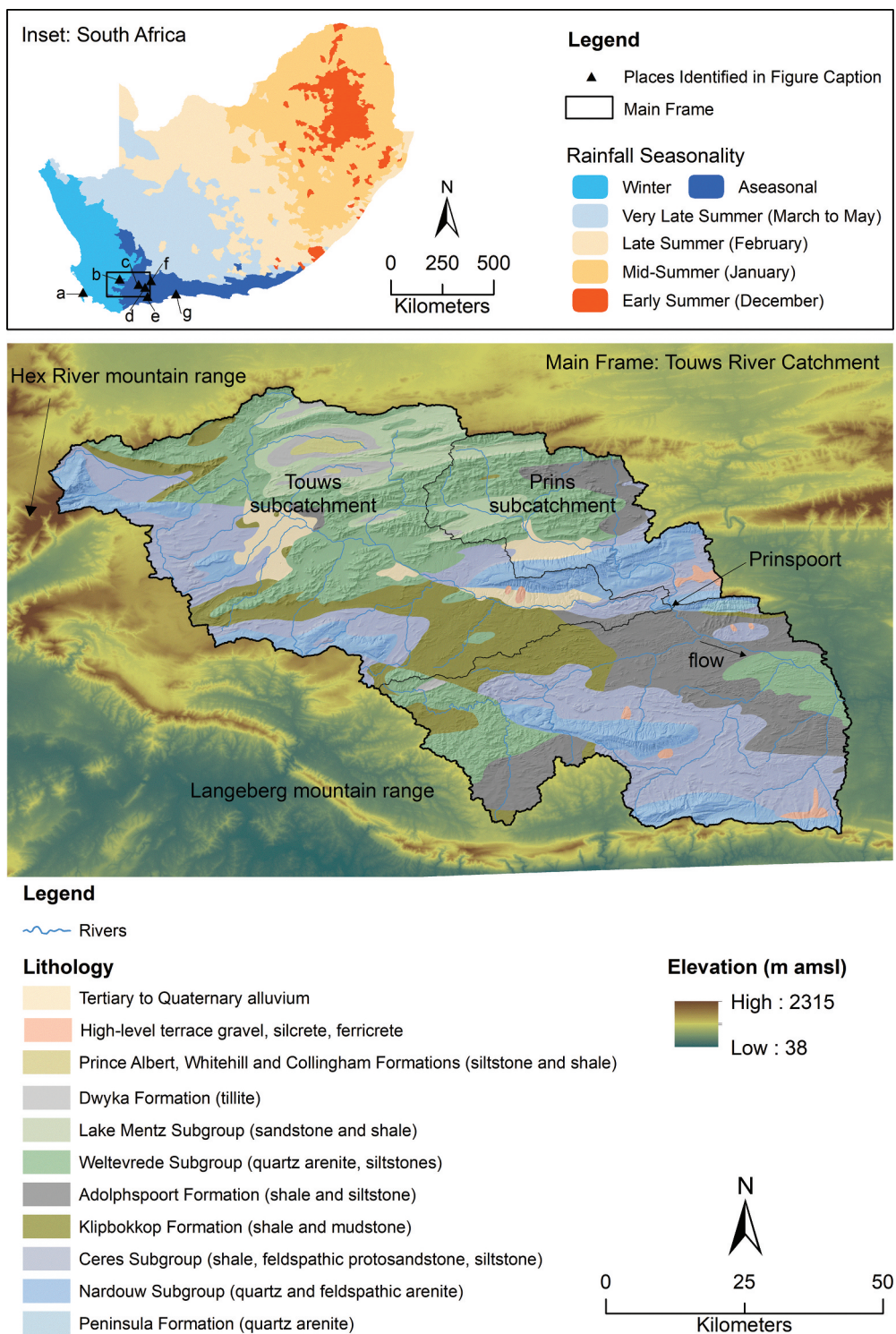


Figure 1. Regional setting of Prinspoort at the confluence of the Touws (trunk) and Prins (tributary) rivers in the Little Karoo, South Africa. The main frame shows the lithology ([Dataset] Council for

mudstones, and siltstones of the Ceres, Bidouw (Klipbökkop Formation) and Traka (Adolphspoor Formation) Subgroups of the Bokkeveld Group.

The Prins River rises at an elevation of 812 m amsl and follows a similar east-southeast alignment for 52 km down-valley to its confluence with the Touws, but cuts south in places through the quartz arenite Nardouw Subgroup relief of the Anysberg and Touwsberg Mountains. Much of the upper catchment of the Prins River is composed of dissected Weltevrede Subgroup quartz arenites, with broader valleys and flats underlain by Ceres, Lake Mentz and Traka Subgroup fine-grained sedimentary rocks. The Prins River is confined within a gorge (Prinspoort) as it traverses the toe of Touwsberg Mountain (Rietvlei Formation quartz arenite of the Nardouw Subgroup), and joins the Touws River as it exits this structure through a fault-bounded (low tectonic activity; Bordy et al., 2018) contact with Adolphspoor Formation shale.

2.2. Climate

The Touws River catchment lies at the transition from the present-day winter to aseasonal rainfall zones (Figure 1). The westernmost 25% [75%] of the Touws sub-catchment lies within the winter rainfall zone [aseasonal rainfall zone], while the entire Prins sub-catchment lies within the aseasonal rainfall zone (Figure 1, Table 1). The dominant rain-bearing circulation types of the winter rainfall zone are seasonal mid-latitude cyclones (cold fronts) and episodic cut-off lows, both associated with westerly waves. While cold fronts regularly make landfall over western coastal regions (e.g., Cape Town; Figure 1), their incursion to the Little Karoo is disrupted by the rain shadow effects of the Hex River and Langeberg mountain ranges flanking the western and southwestern margins of the Touws River catchment, respectively (Figure 1 main frame). Cut-off lows also affect the aseasonal rainfall zone, slowly traversing large areas over a period of two to three days, yielding a deluge of rainfall in the event that can account for most of the total rainfall in that year of record (Favre et al., 2013). The largest floods on record for the Touws River are associated with cut-off lows (M. C. Grenfell et al., 2021). High-intensity convective thunderstorms are also a feature of the Little Karoo, although their occurrence is both spatially and temporally variable (the circulation types driving these events are more characteristic of the higher-elevation Great Karoo and summer rainfall zones to the east; Figure 1).

The aseasonal rainfall zone in South Africa is something of a misnomer. South of the Langeberg-Outeniqua fold mountain ranges (e.g., George; Figure 1), ‘aseasonal’ means that year-round rainfall is *likely*, with frontal and post-frontal orographic rainfall a more regular seasonal occurrence than in the Little Karoo, more regular coastal and cut-off lows, and tropical-temperate troughs (Macron et al., 2014)

Geoscience CGS, 2019) and general terrain (hillshade model; [Dataset] Japanese Aerospace Exploration Agency (JAXA), 2021) of the Touws River catchment and the sub-catchment boundaries of the Touws and Prins rivers upstream of their confluence. The inset shows the location of the main frame in relation to the spatial distribution of rainfall seasonality across South Africa ([Dataset] Schulze & Maharaj, 2007b), with place names mentioned in the text (a = Cape Town, b = touwsrivier, c = Prinspoort, d = Touws River sites studied in Damm & Hagedorn, 2010, e = barrydale, f = seweweekspoor, g = george).

Table 1. Summary statistics of mean annual precipitation (MAP; [Dataset] Lynch & Schulze, 2007), mean annual potential A-Pan evaporation (PE; [Dataset] Schulze & Maharaj, 2007a), and (UNEP (1997) aridity index (AI = MAP/PE) for the sub-catchments upstream of the Touws – Prins confluence.

Climate descriptor	Touws sub-catchment	Prins sub-catchment
MAP min. (mm)	55	49
MAP max. (mm)	450	460
MAP mean (mm)	256	212
MAP std. dev. (mm)	54	75
Spatial CV (%)	21	35
PE min. (mm)	1637	1662
PE max. (mm)	2330	2295
PE mean (mm)	2111	2106
PE std. dev. (mm)	106	107
Spatial CV (%)	5	5
AI min.	0.02 (hyper-arid)	0.02 (hyper-arid)
AI max.	0.24 (semi-arid)	0.22 (semi-arid)
AI mean	0.12 (arid)	0.10 (arid)
AI std. dev.	0.03	0.04
Spatial CV (%)	25	40
% Winter Rainfall Zone	25	0
% Aseasonal Rainfall Zone	75	100

spreading the long-term annual rainfall distribution more evenly through the year than is the case in the winter rainfall-dominated west (Tyson & Preston-Whyte, 2000). This is part of the reason for the occurrence of large areas of forest vegetation in the general vicinity of George (with a preponderance of south-facing slopes also playing a role; Mucina & Geldenhuys, 2006). However, in the arid Little Karoo interior, ‘aseasonal’ means that rainfall is unlikely overall (Mucina & Rutherford, 2006) but can occur at any time of year, given the variety of winter- and summer-associated circulation types and the effects of orographic barriers on their variability of occurrence.

The result in the Touws River catchment is a pattern of high spatial and temporal variability in rainfall and river flows, with low-order drainage being ephemeral and larger rivers (Strahler 3rd order and higher) being ephemeral to intermittent (M. C. Grenfell et al., 2021). The mean annual temperature at the study site is 17.9°C, with summer temperatures sometimes exceeding 40°C (Mucina et al., 2006). Other climate descriptors for the Touws and Prins sub-catchments are summarized in Table 1 based on spatial extractions from datasets cited in the Table caption. Mean annual precipitation (MAP) for the Touws sub-catchment ranges from 55 to 450 mm, with a spatial mean of 256 mm and a spatial coefficient of variation (CV) of 21%. The MAP for the Prins sub-catchment ranges from 49 to 460 mm, with a spatial mean of 212 mm and a spatial CV of 35 %. The mean annual potential A-Pan evaporation (PE) for the Touws sub-catchment ranges from 1637 to 2330 mm, with a spatial mean of 2111 mm, and for the Prins sub-catchment from 1662 to 2295, with a spatial mean of 2106 mm. Both sub-catchments have a spatial CV in PE of 5%. United Nations Environment Programme (UNEP) (1997) aridity index (AI) values for the two sub-catchments are similar, with sub-catchment mean AI values of 0.10 (Prins) and 0.12 (Touws) falling within the arid class, although the spatial CV in MAP and AI is greater in the Prins sub-catchment.

2.3. Vegetation and land use

Catchment vegetation is diverse, and its spatial distribution is controlled largely by lithology and aspect (vegetation unit nomenclature after Mucina & Rutherford, 2006). The high elevation but low relief surfaces underlain by fine-grained sedimentary rocks are dominated by Renosterveld of the Units FRs 6 (Matjiesfontein Shale Renosterveld) and FRs 7 (Montagu Shale Renosterveld). Small areas of Fynbos such as Units FFs 23 (North Swartberg Sandstone Fynbos) and FFs 24 (South Swartberg Sandstone Fynbos), varying with aspect, occur on the quartz arenite Nardouw Subgroup relief of Anysberg and Touwsberg mountains, with more extensive areas of Unit FFq 3 (Matjiesfontein Quartzite Fynbos) on Weltevrede Subgroup rocks. The plains and low undulating hills of lower elevation areas are dominated by succulent and non-succulent shrubland of Unit SKv 8 (Western Little Karoo), with patches of Unit SKv 10 (Little Karoo Quartz Vygieveld) growing on soils weathered from quartz veins within mud-dominated lithologies (Mucina & Rutherford, 2006). Riparian vegetation at the confluence of the Prins and Touws rivers comprises riverine thicket of Unit AZi 8 (Muscadel Riviere), which is dominated by *Vachellia karroo* and *Caroxylon aphyllum*, and is variably invaded by *Tamarix* spp (M. C. Grenfell & Dube, 2022; Mucina et al., 2006).

The rugged terrain and relative water scarcity of the Little Karoo meant that European settlement and major agricultural activity in the region was initiated later (1850s to 1880s) than in other parts of the Western Cape. Before European settlement in the 1800s, the region was sparsely populated by nomadic Khoekhoen herders who drove small herds of fat-tailed sheep in response to the natural spatiotemporal variability in grazing resources (Morris, 2018). Present-day catchment land use comprises a mosaic of formally protected conservation areas administered by the provincial conservation authority or private consortia, private game farms established for hunting, vegetable seed cultivation (irrigated by borehole water or flow diversions and runoff harvesting schemes), fat-tailed sheep or goat farming (on irrigated pastures or in open rangeland), and Bonsmara cattle farming (in open rangeland and heavily reliant on seasonal wetland grazing; Makhonco, 2019).

3. Methods

3.1. Geospatial modelling of catchment connectivity

Two geospatial indices were derived from a 30 m resolution digital elevation model (JAXA, 2021) to contextualize sediment dispersal through the Touws River catchment, and at the confluence of the Touws and Prins rivers. First, SedInConnect was used to compute the index of connectivity (IC) of Borselli et al. (2008), with modifications and a general tool workflow following Crema and Cavalli (2018) and M. C. Grenfell et al. (2022). Given the relatively low resolution of the available DEM, the representation of topographic roughness is somewhat simplified, and the main landscape attributes driving a cell IC value are i) the average slope and size of the upslope contributing area (based on TauDEM D ∞ flow routing; Tarboton, 1997), and ii) the proximity and steepness of descent to the nearest target cell, where the target feature is a drainage line shapefile (Strahler 3rd order and higher) extracted from the DEM using TauDEM D8 flow routing. The output was

symbolized using a four-class quantile classification in ArcMap 10.8.2 to display the spatial distribution of areas with low, moderately low, moderately high and high target connectivity.

Second, the relatively potential stream power index (RPSPI) was computed for the drainage extracted previously using TauDEM Tools v5.3. RPSPI is the map algebra product of the upslope contributing area (D8, converted to catchment area in km² as a proxy for potential flood discharge) and slope in the D8 downslope direction averaged over a distance of 1 km. A flood discharge proxy was used due to the lack of widespread flow gauging across the catchment and the known role of infrequent large floods in shaping river morphology in this (M. C. Grenfell et al., 2021) and similar dryland river systems (Tooth, 2013). Measures typically used in stream power calculations, such as the mean annual discharge or the mean/median annual flood, besides being difficult to quantify with limited discharge data, do not represent geomorphologically effective flows in these environments (M. C. Grenfell et al., 2021). The output was symbolized using a three-class natural breaks classification in ArcMap 10.8.2, with line widths of 1, 2 and 4 for the low, medium and high stream power classes, respectively. Draping the RPSPI lines over the target IC raster provides a visual representation of the spatial variation in catchment- and confluence-scale lateral sediment supply (IC) and longitudinal transfer (RPSPI) potential, providing context for the site geomorphology, sedimentology and geochronology efforts.

3.2. Field survey and characterisation of sedimentology

An alluvial terrace was discovered at Prinspoort during reconnaissance fieldwork in 2017 (approximately bound by the 415 and 410 m contours at OSL-XS1 and XS2 in Figure 2, with the valley setting and terrace face shown in photographs in Figure 3). The stratigraphy of the terrace face was logged in the field and sampled both at regular 0.5 m intervals and at textural contacts for laboratory analyses to refine the field log. A differential GPS (dGPS) with relative sub-decimetre accuracy in x, y and z was used to survey a longitudinal profile of the thalweg and terrace/floodplain surfaces and to capture valley cross sections through the Prins – Touws confluence. Cross sections were located at the face sampling site (OSL-XS1) and downstream end of the terrace (XS2) where the Prins River is confined within the Prinspoort gorge, through a zone of gradual valley widening (XS3 and 4), to the transition to the broader Touws River valley and Prins – Touws confluence at XS7 through XS5 and 6 (Figure 2). In addition, a shallow cut (~0.5 m) into the Touws River floodplain upstream of the confluence with the Prins valley (TW1), and the modern Touws River channel bank (TW2) were logged and sampled to support luminescence dating of floodplain fill (Figures 2 and 3), in order to establish the relationship between floodplain development and tributary valley fill.

Sediment samples were pre-treated with hydrogen peroxide and hydrochloric acid to remove organic and carbonate cementing agents, respectively, dried and gently dispersed using a pestle and mortar. Samples were then passed through a vibrating stack of sieves with mesh sizes 2, 1, 0.5, 0.25, 0.125 and 0.063 mm, to quantify the sample particle size distribution (D_{50} and % gravel, sand, silt+clay) and uniformity coefficient (D_{60}/D_{10}) using HydrogeoSieveXL (Devlin, 2015). The organic carbon content of samples from the

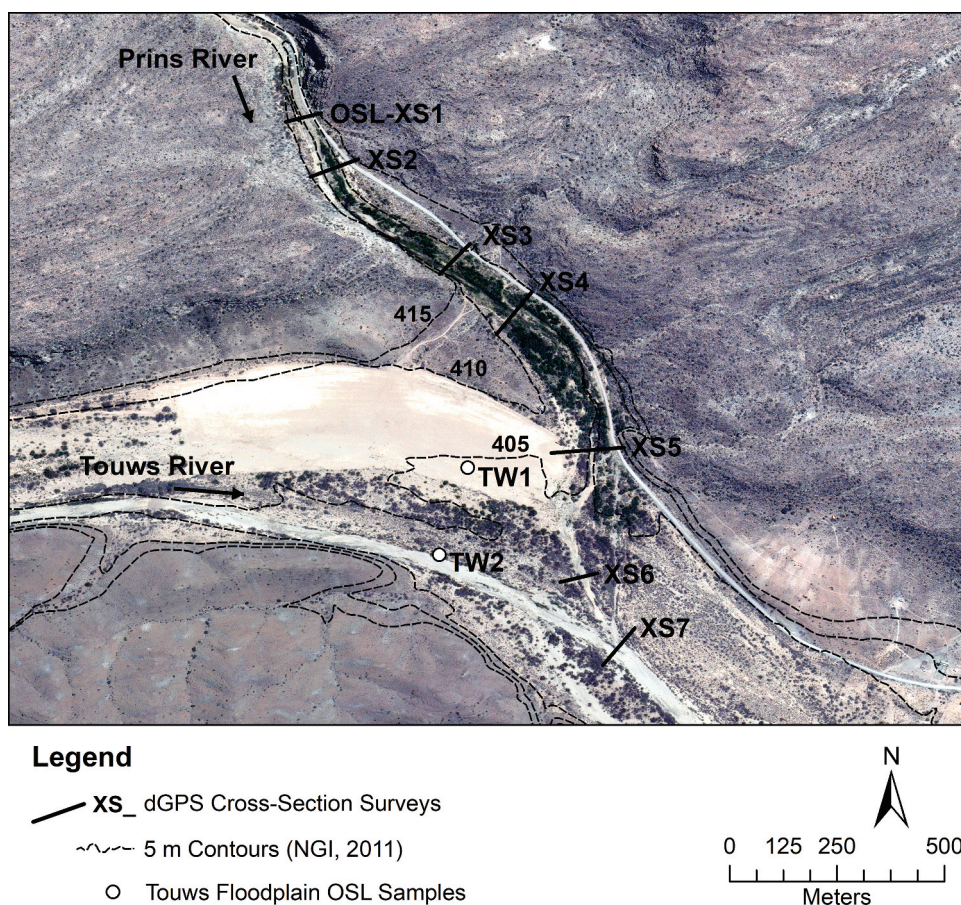


Figure 2. Location of field dGPS valley cross-section surveys and sediment sampling/OSL dating sites (prinspoort terrace – OSL-XS1, Touws River floodplain – TW1, TW2) in relation to the general terrain of the prins-Touws confluence (5 m contours; NGI, 2011). The base image is a 0.5 m resolution Worldview colour composite (23 Nov 2017).

terrace face was determined by loss on ignition using a blast furnace run at 550°C for 6 hours, to complement the particle size information in defining deposition and post-deposition environments.

3.3. Luminescence dating

Six clearly identifiable sand-dominated lenses from the Prinspoort terrace face were targeted for luminescence sampling, which involved driving 76 mm diameter (1.3 mm wall thickness) aluminium tubes (300 mm in length) into the face through the lens. Tubes were securely sealed in the field, and sample windows (the central 100 mm of the tube) were marked to ensure that the material selected for luminescence dating in the laboratory would not have been exposed to light. In addition, two sand-dominated lenses at the Touws floodplain sites described in Section 3.2 were sampled for luminescence dating, as described above.

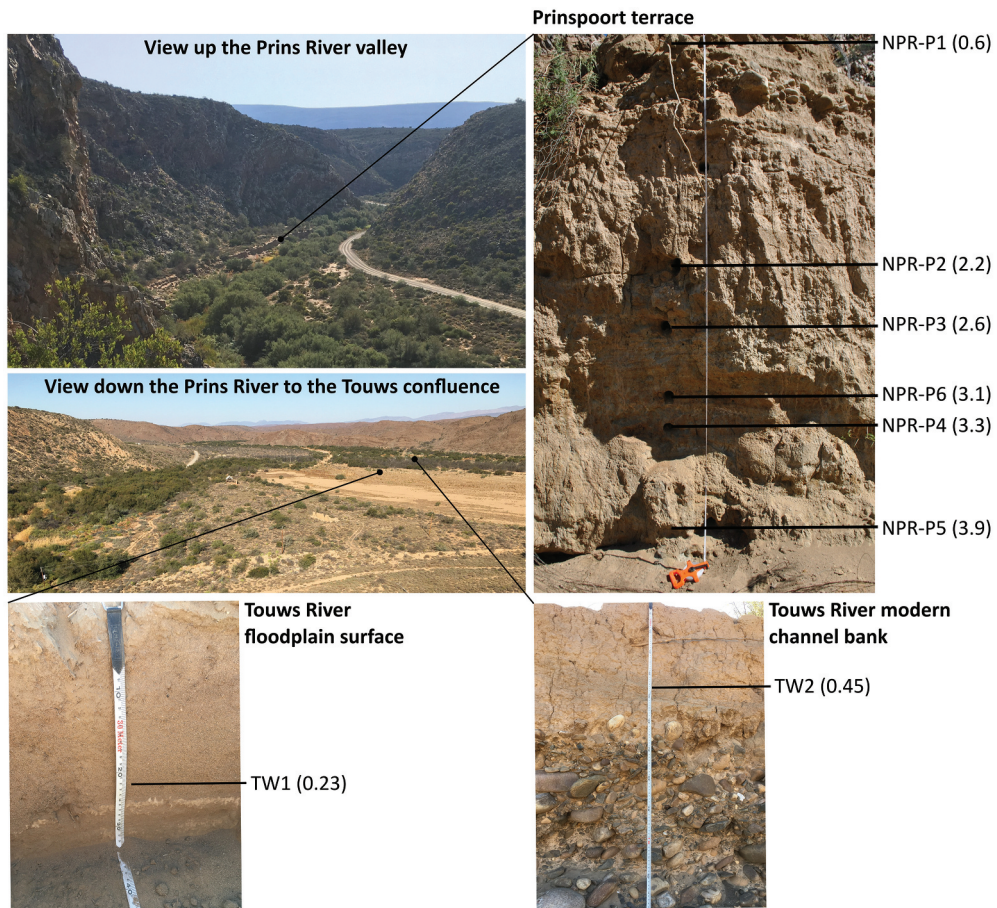


Figure 3. Field photographs of the general setting and stratigraphy of the sediment sampling/OSL dating sites (see Figure 2 for site locations). Profiles and locations of OSL samples are indicated, with sample depths (m below ground surface) in brackets.

Luminescence samples were opened in the OSL lab at the University of the Witwatersrand under safe red-light conditions. Quartz grains were extracted and prepared following standard OSL preparation procedures (Aitken, 1998; Galbraith et al., 1999). The 180–212 μm size fraction was treated with ~20% solutions of HCl and H_2O_2 to remove carbonates and organic matter, respectively. Quartz grains were separated from heavy minerals using a density separation with sodium polytungstate and then treated with a 42% solution of HF to remove the alpha contribution to the dosimetry and any remaining feldspar contamination.

Measurements of the OSL signal were performed using Risø TL/OSL DA-15 and DA-20 Readers, in which optical stimulation was provided by blue LEDs (478 nm) at $\sim 40 \text{ mW cm}^{-2}$ constrained by a long-pass filter (GG-420). The OSL signal was detected with an EMI 9235QB photomultiplier tube via a single 7.5 mm Hoya U340 detection filter. The measurement procedure follows the Single Aliquot Regenerative dose (SAR) protocol of Murray and Wintle (2003). Dose recovery and preheat plateau tests were used to

determine the ideal temperature at which a known dose could be accurately recovered and to assess whether D_e values remained consistent as temperature changed, respectively (Murray & Wintle, 2003). Prior to these measurements, the quartz grains were exposed to sunlight for 1 h and then bleached with OSL at 205 °C. Preheat temperatures shown to recover a given dose to within 0.9 to 1.1 of unity, and with the most consistent D_e values, were selected for further measurement. The OSL signal was collected from the initial 0.125 s of the decay curve, with the subsequent 0.25 s used for the ‘background’. This channel choice maximizes the dominance of the ‘fast’ OSL component while keeping the signal-to-noise ratio as high as possible, following the low background subtraction of Cunningham and Wallinga (2010). The equivalent dose (D_e) was evaluated using an exponential fit to regenerative dose points.

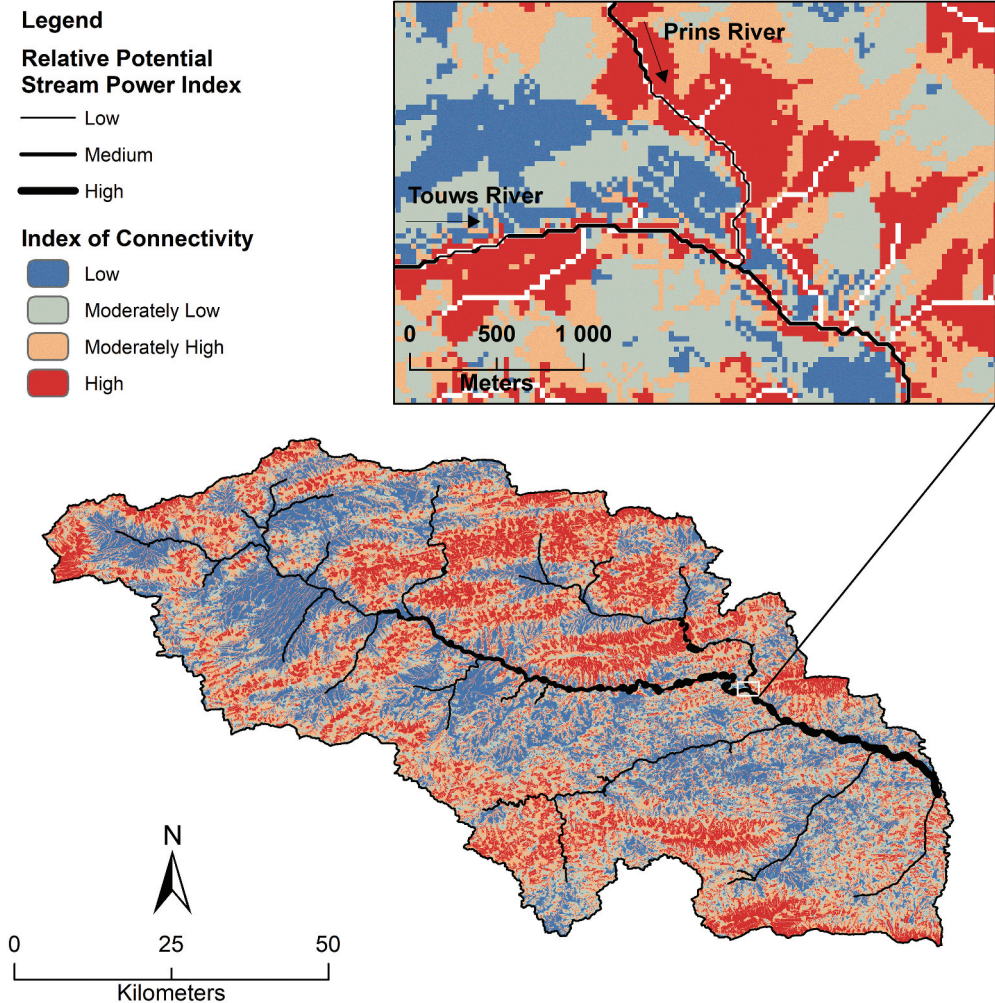
The radionuclide abundances of U, Th and K were measured through ICP-MS (U, Th) and XRF (K) at Earth Lab, University of the Witwatersrand. The environmental dose rates to quartz grains were calculated from the isotopic abundances of the parent radionuclides, using the conversion factors of Guérin et al. (2011). The internal alpha dose rate was corrected according to Vandenberghe et al. (2008). Allowance was made for the attenuation of the beta dose due to grain size (Mejdahl, 1979). Dose attenuation due to water follows Zimmerman (1971), using measured water content values. The cosmic dose rates were determined as a function of altitude, latitude, longitude and depth, according to Prescott and Hutton’s (1994) equations, including the soft component from Madsen et al. (2005). A constant sedimentation rate was assumed for the cosmic component. Detailed luminescence results from individual samples were plotted as abanico plots, following the methodology presented by Dietze et al. (2016) and using the R package described by Burow et al. (2016).

4. Results

4.1. Potential catchment-scale sediment dispersal

Figure 4 shows the spatial variation in potential sediment dispersal across the Touws River catchment (main frame), and at the confluence of the Prins and Touws rivers (inset). The geospatial model of lateral connectivity of catchment surfaces with respect to the drainage network (target feature in the IC analysis) provides a measure of potential sediment supply, while the model of potential stream power provides a measure of potential sediment transfer through the drainage network. Together with lithology (Figure 1), the composite model of supply and transport (Figure 4) yields insight into fundamental geomorphic controls on connectivity across the catchment and at the confluence site. Tributaries of the upper ~50 km of the Touws River draining the west and southwestern catchment margins flow through broad, low connectivity basins with low potential stream power that host sandy Quaternary alluvium. Further downstream and approaching the confluence with the Prins River, the Touws valley has moderately high to high connectivity owing to the steep, well-dissected high ground to the north of the river and is flanked by coarser Tertiary to Quaternary colluvial and alluvial sediments (Figure 1).

The Prins River is largely set within the steep, well-dissected terrain north of the Touws sub-catchment and has higher structural connectivity. There is some Quaternary



sand accumulation in low connectivity, low stream power basins underlain by fine-grained sedimentary rocks, with strong hillslope-channel coupling and medium to high stream power where the river flows through confined, fault-oriented gorges set within quartz and feldspathic arenites of the Weltevrede and Nardouw Subgroups. As the Prins River approaches the confluence with the Touws, lateral connectivity is high due to the confinement within the Prinspoort gorge, but a reduction in longitudinal slope results in a transition from medium to low potential stream power (Figure 4). Such a combination of high hillslope-channel coupling and low stream power would favour local valley alluviation and enhance the potential for tributary valley impoundment by trunk river floodplain development (e.g., K. Fryirs & Gore, 2014; M. C. Grenfell et al., 2008) because

material is being supplied to an area with inherently low transport capacity. The greater spatial variability in MAP and PE within the Prins River catchment (Table 1) may also enhance the potential for impoundment by the trunk river because flow variability results in episodic transport-limitation (Milliman & Farnsworth, 2011).

4.2. Geomorphology and sedimentology of the Prins-touws confluence

The longitudinal profile of the Prins-Touws confluence shows contiguity in the Prinspoort terrace – Touws floodplain surface transition and convexity in the thalweg

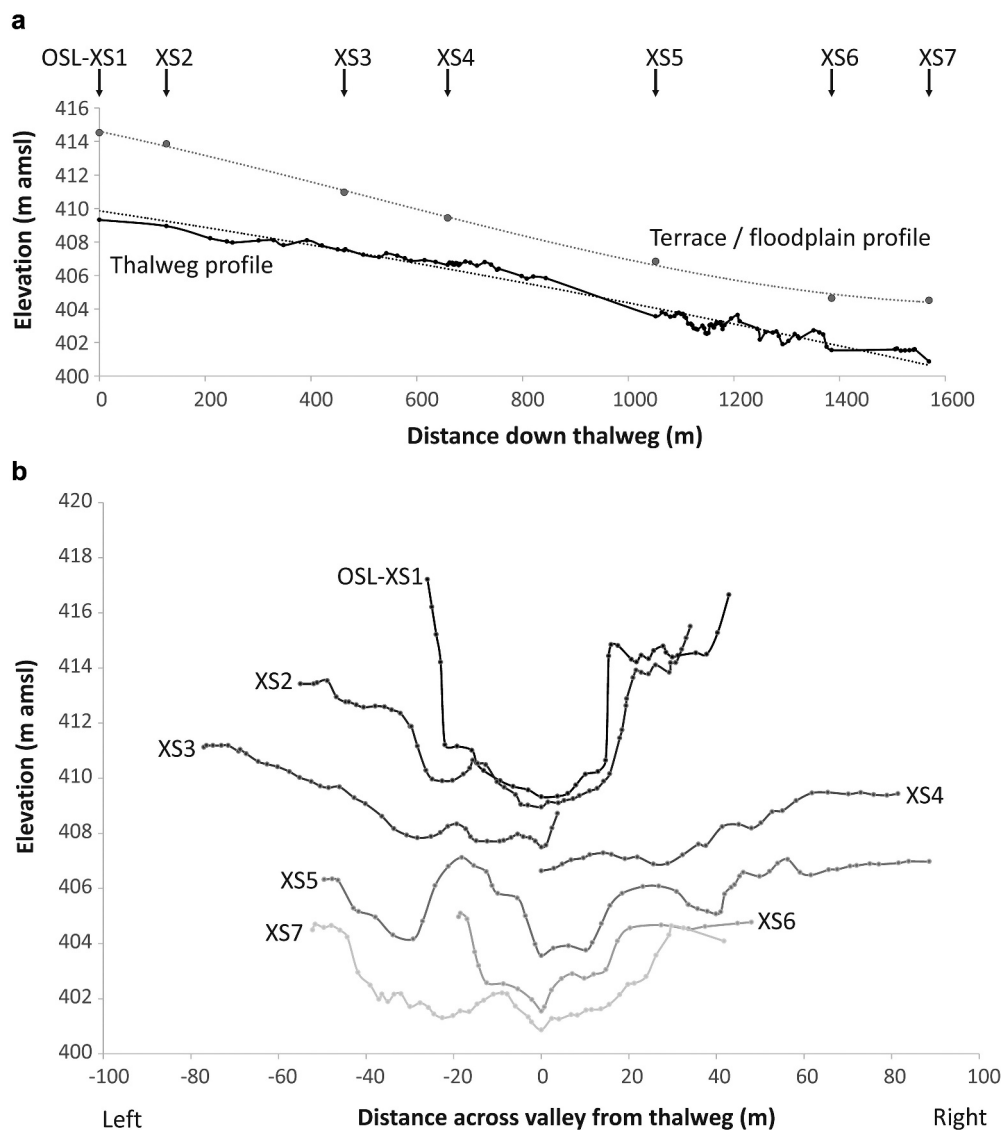


Figure 5. Longitudinal variation in valley thalweg and terrace/floodplain surface elevation (a), and valley cross-sectional form (b), through the Prins-Touws confluence.

Figure 5(a), the latter being indicative of local bed aggradation where the Prins valley begins to widen as it intersects the Touws floodplain XS3, 4, 5; Figures 2 and 5(b). The Prins channel becomes much smaller in cross-section through XS3/4, then develops into a series of small distributaries at XS5 and a single larger re-forming channel at XS6. XS7 crosses the Prins and Touws channels at their confluence, where the Touws River channel is three to four times wider than the Prins. Channels at XS5 and 6 are set within sandy alluvium that could represent tributary deposition through the region of loss of tributary valley confinement or comprise part of the Touws River floodplain (or some combination of these environments). The contiguity of the tributary and trunk surface profiles suggests that dynamic co-alluviation of the trunk floodplain and tributary valley led to quasi-impoundment of the confined valley upstream of the confluence.

Prinspoort terrace comprises a sequence of medium to coarse sand-dominated lenses (targeted for dating) that fine upward with normal grading (particle size distributions of samples are illustrated in Figure 6), suggesting deposition under rapidly waning flood conditions (Miall, 1996) characteristic of this environment. This sequence overlies a bed of massive bedrock and large boulders and is capped at the top by a much coarser unit (cobble and gravel fining upward to sand). Samples with a higher silt and clay fraction have slightly higher organic carbon content (Figure 6), which may suggest the localized development of palustrine wetland conditions in the valley during the time of accumulation or reflect the original catchment soil source. There is evidence of active groundwater discharge draining to the current channel bed from the modern Rietvlei Formation valley margin (a feature of Table Mountain Group rocks in general) and dense stands of *Phragmites australis* at current pool margins adjacent to and immediately downstream of the terrace.

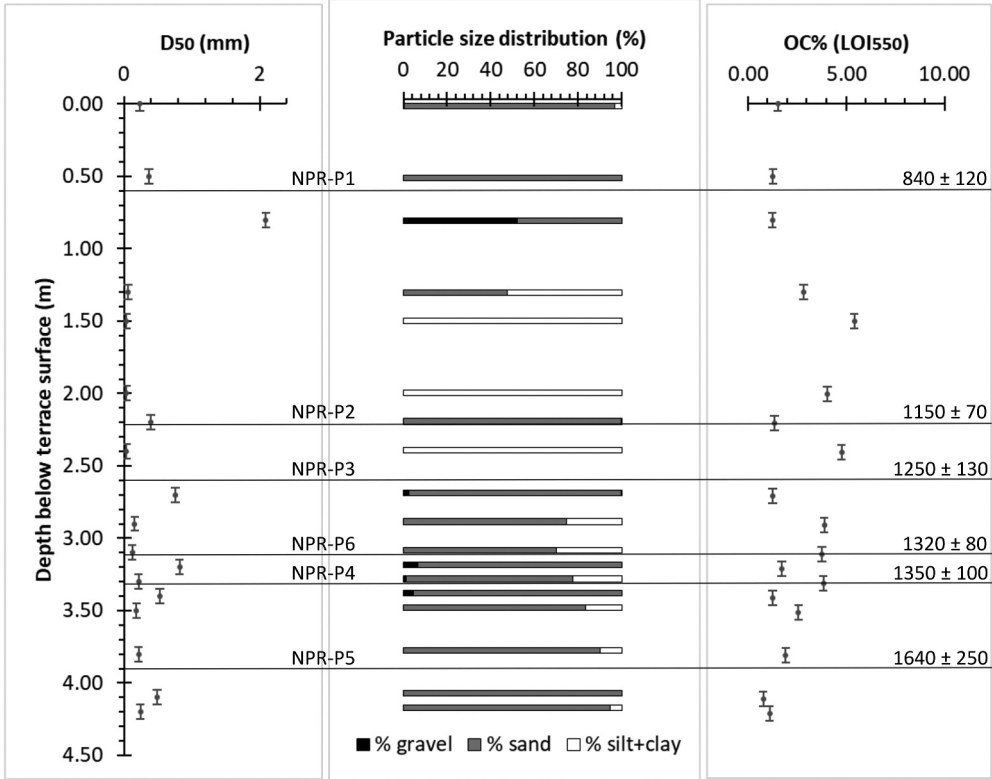
4.3. Luminescence chronology

Luminescence ages reported in Table 2 are based on the central age model (CAM) due to the low signal-to-noise ratios and high overdispersion (e.g., Figure 7) evident in most of the sample set (following Knight & Evans, 2018; Medialdea et al., 2014). All ages are reported in years before the year of collection of the Prinspoort samples (2017), and all are younger than 2000 years. Overdispersion is common in sediments deposited by dryland non-perennial rivers, where although sunlight is usually readily available post-deposition, the episodicity and high concentration of sediment transport events may lead to incomplete and inconsistent resetting of the luminescence signal. This is especially problematic for young deposits, such that the CAM ages represent a maximum possible age for the samples (Knight & Evans, 2018). Sample ages from the terrace follow the expected vertical sequence (older with depth), and the age distribution of floodplain samples is consistent with their distance from the active channel, lending credibility to the dating. However, overlap in ages when accounting for error terms suggests that the floods occurred in relatively quick succession, with an overall average tributary valley vertical aggradation rate of $\sim 2.4 \pm 0.35$ mm/yr during the term of deposition.

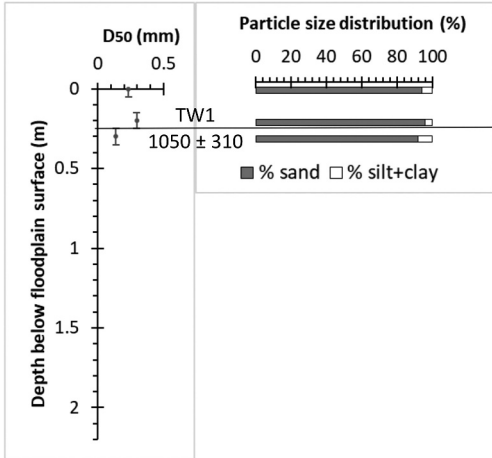
5. Discussion

The stratigraphy of the Prinspoort terrace is described from the base (oldest deposits) to the surface (youngest deposits), with reference to Figure 8. Unit a was immediately above

Prinspoort terrace



Touws River floodplain surface



Touws River modern channel bank

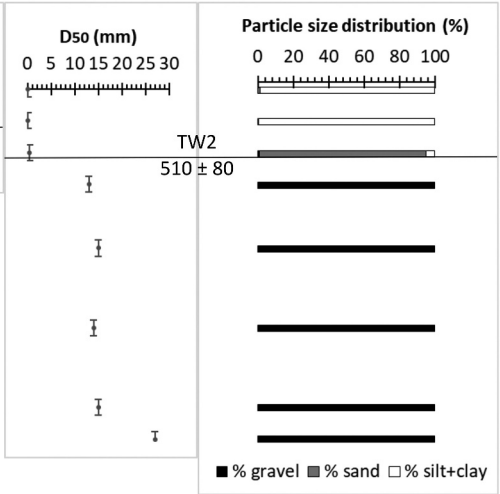


Figure 6. Stratigraphic logs and OSL dates (years BP) for samples from the prinspoort terrace and Touws River floodplain.

Table 2. Luminescence ages of samples calculated using the central age model (CAM). Sample locations are shown in Figures 2 and 3. Measured water contents are considered typical of the long-term average condition of these sites due to the aridity of the local environment and position of sample locations in an open face elevated above observed groundwater discharge.

Sample	Lat (°S)	Long (°E)	Depth (m below surface)	Water Content (%)	Th (ppm)	U (ppm)	K (%)	Total Dr (Gy/ka-1) and error	Overdispersion (%)	Total De (Gy) CAM and De error	CAM Age (years BP) and error
NPR_P1	33.568013	20.847492	0.60	0.17	8.43	3.47	1.843	3.49 ± 0.09	85	2.92 ± 0.41	840 ± 120
NPR_P2	33.568013	20.847492	2.20	0.32	7.17	3.44	2.441	3.93 ± 0.10	40	4.52 ± 0.27	1150 ± 70
NPR_P3	33.568013	20.847492	2.60	0.38	11.48	3.66	2.241	4.09 ± 0.10	76	5.12 ± 0.53	1250 ± 130
NPR_P6	33.568013	20.847492	3.10	0.83	13.28	3.64	2.374	4.31 ± 0.11	28	5.68 ± 0.32	1320 ± 80
NPR_P4	33.568013	20.847492	3.30	0.4	12.56	4.59	2.374	4.51 ± 0.11	42	6.07 ± 0.44	1350 ± 100
NPR_P5	33.568013	20.847492	3.90	0.5	11.26	4.31	2.067	4.05 ± 0.31	69	6.64 ± 0.99	1640 ± 250
TW_1	33.575415	20.850963	0.23	0.72	19.55 ± 0.15	4.80 ± 0.03	2.49 ± 0.02	5.11 ± 0.13	58	5.37 ± 1.59	1050 ± 310
TW_2	33.577213	20.850377	0.45	2.51	18.93 ± 0.15	4.81 ± 0.03	2.46 ± 0.02	5.03 ± 0.13	59	2.55 ± 0.41	510 ± 80

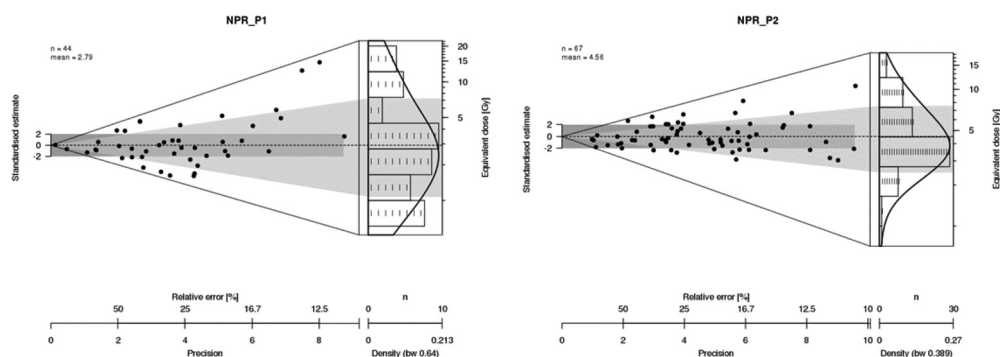


Figure 7. Examples of abanico plots for samples NPR-P1 and NPR-P2 from Prinspoort terrace, showing the De distributions within 2σ .

sheet bedrock, comprised boulders overlain by fine sand low in fines, and is interpreted as a channel lag deposit that represents the onset of tributary valley filling prior to $\sim 1640 \pm 250$ years BP (the time of deposition of unit **b** above). Unit **b** is the earliest in a series of flood deposits archived in the terrace, comprising moderately well-sorted medium sand low in fines (NPR-P5), fining upward to fine sand low in fines. Units **c** and **d** both comprise moderately well-sorted coarse sand low in fines (NPR-P4 and NPR-P6, respectively), fining upward to fine sand low in fines, marking flood events at $\sim 1350 \pm 100$ years BP and $\sim 1320 \pm 80$ years BP, respectively. Unit **e** is a flood sequence from $\sim 1250 \pm 130$ years BP, comprising moderately well-sorted coarse sand low in fines (NPR-P3), fining upward to silty clay. Unit **f** comprises moderately well-sorted medium sand low in fines (NPR-P2), fining upward to silty clay, marking a flood event at $\sim 1150 \pm 70$ years BP.

There is an indistinct transition to unit **g** above, a layer of fine sandy silt that might represent an additional flood event (not dated) that occurred between the time of deposition of units **f** and **h**, or simply general aggradation during the term of impoundment. Unit **h** comprises matrix-supported long-axis imbricated cobbles (in a matrix of poorly sorted fine gravel low in fines), fining upward through moderately well-sorted medium sand low in fines containing fragments of charcoal (NPR-P1) to fine sand low in fines. The terrace surface is characterized by flood-runners, while planar contacts characterize the sand units below. Thus, unit **h** represents a large hyper-concentrated flash flood event at $\sim 840 \pm 120$ ka, after which tributary valley deposition ceased.

In summary, the terrace sequence preserves three main phases of deposition; 1) a high energy cobble and boulder channel bed deposit (unit **a**), 2) a series of sand-dominated layers deposited as sheets within the valley, with a general trend of fining towards the surface indicating a reduction in flood energy over time (unit **b**, **c** and **d** with fining in units **e** and **f**), and 3) a high energy flash flood deposit with charcoal fragments (unit **h**). The overall fining-upward trend observed in the face is interpreted as a temporal reduction in flow velocity associated with decreasing longitudinal slope, which suggests that the Prinspoort tributary base level was gradually being elevated as a result of aggradation downstream, associated with trunk floodplain development, tributary fan development, or co-alluviation at the tributary – trunk confluence. This phase of aggradation was truncated soon after or during the major flood that deposited unit **h**.

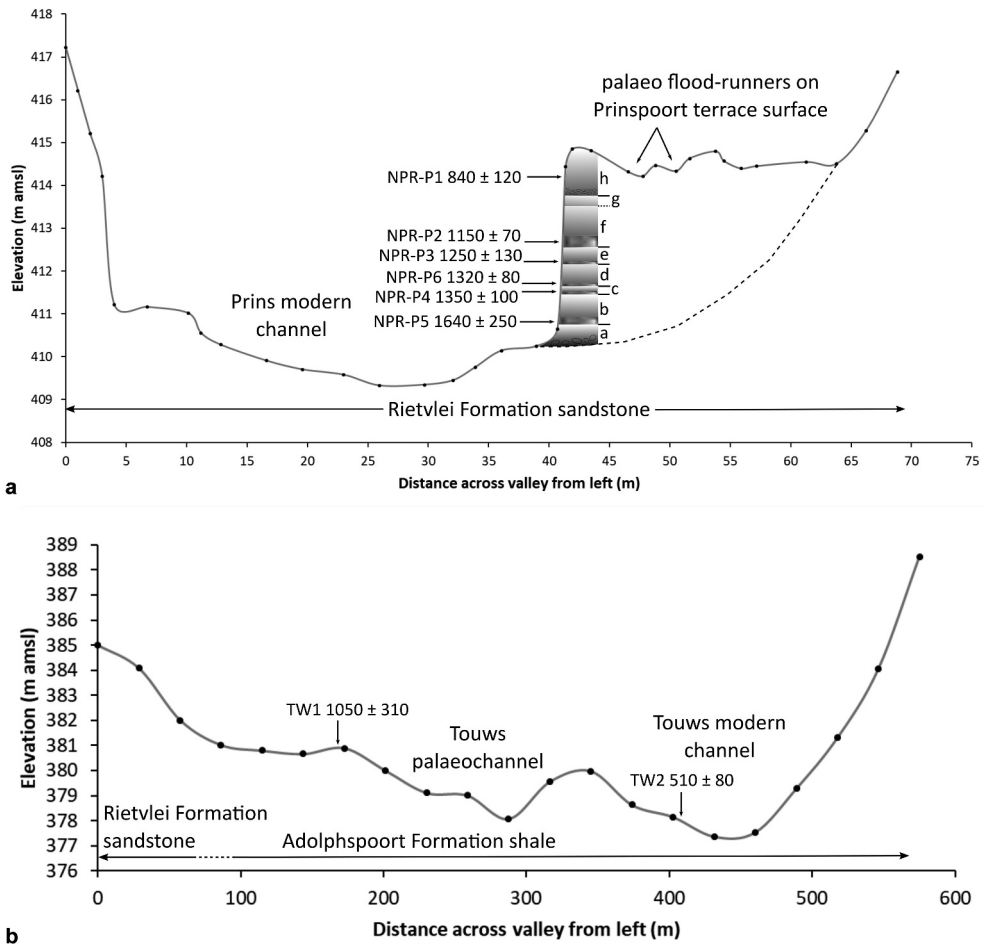


Figure 8. Prinspoort terrace (a) and Touws River floodplain (b) stratigraphy and OSL chronologies in relation to valley cross-sectional form. Descriptions of prinspoort terrace units labelled a to h are provided in the text.

The most recent deposition archived on the floodplain of the Touws River palaeochannel at TW1 (0.23 m below the surface; Figures 3 and 8) comprises a unit of upward-fining medium sand overlying fine sand at 0.25 m below the surface, and dates to $\sim 1050 \pm 310$ years BP. TW2 forms part of the modern Touws River channel bank, comprising a sediment face reflecting relatively recent ($< 510 \pm 80$ years BP, TW2) floodplain accretion by wandering river processes in a mixed gravel/sand bedload-dominated river (S. Grenfell & Grenfell, 2021; M. C. Grenfell et al., 2021). The base of this profile comprises cobble and gravel, fining upward through mainly moderately well-sorted medium sand low in fines (TW2, 0.45 m below the surface) to silty clay.

Overall, the connectivity, geomorphology and sedimentology analyses suggest that floodplain development adjacent to the Touws River palaeochannel, or the combined effects of tributary fan and trunk floodplain co-alluviation, given the proximity to the exit of Prinspoort gorge (Figure 2), impounded the Prins River valley in an area that is

inherently susceptible to transport limitation (Figure 4). The termination of sedimentation on the Touws River floodplain surface $\sim 1050 \pm 310$ years BP may have been associated with realignment of the Touws River to a position favouring tributary – trunk channel coupling (e.g., Brierley & Fryirs, 1999; K. A. Fryirs, Brierley, Preston, & Spencer, 2007; K. Fryirs & Gore, 2014). Ultimately, a large flood event in the Prins River was able to breach the sedimentological local base-level control at the gorge exit, resulting in incision through the tributary valley fill sequence and the formation of Prinspoort terrace. The tributary and trunk floodplain sedimentary archives therefore record a switch (*sensu* K. A. Fryirs, Brierley, Preston, & Kasai, 2007; K. A. Fryirs, 2017; between ca. 1050 ± 310 and 840 ± 120 years BP) from a state of tributary (dis)connectivity to one of connectivity that persists at present. Although beyond the scope of the present paper, we would encourage future research that applies sediment fingerprinting techniques (e.g., Foster et al., 2012) to discriminate spatial and temporal variation in the relative contribution of tributary and trunk sediment sources to geomorphic development at non-perennial river confluence zones, and to test landscape-scale connectivity model predictions of the relative contribution of different sediment source areas.

Two fluvial chronologies are available for comparison with the Prinspoort chronology, avoiding broader comparisons with palaeoenvironmental reconstructions from the Great Karoo which is located at a higher elevation, underlain entirely by Karoo Supergroup lithologies that express a markedly different landscape surface structure and is summer rainfall dominated. The first comparative chronology is a valley-fill sequence in the upper Huis River near Barrydale (Bordy et al., 2018), which is located ~ 45 km south-southwest of Prinspoort in a fynbos-dominated environment within the Langeberg mountain range dividing the Little Karoo from the wetter aseasonal rainfall zone to the south (seaward; Figure 1). The site is similar to Prinspoort in its confined valley form and location upstream of a confluence, although the valley alluviation documented is within the trunk system. Although the authors did not formally investigate the potential influence of the downstream tributary on valley impoundment, other work from similarly well-dissected Cape Fold mountain environments highlights the possibility of tributary alluvial fans impinging upon and impounding the trunk valley (Pulley et al., 2018). However, sediment can accumulate in a confined valley without local base level control if there is a wave of supply that locally or temporarily exceeds the transport capacity. This condition is not typically sustained for a long period of time (Jansen & Brierley, 2004), such that bedrock-confined environments preserve short and sometimes incidental sedimentary records (Bordy et al., 2018; Miall, 2014). A large flood and/or the exceedance by aggradation of a threshold slope (Ellery et al., 2016) will ultimately drive incision and scour.

The Huis River site contains a record of high-magnitude, debris-flow-dominated palaeofloods that are constrained chronologically by a sediment-associated charcoal record covering the period $\sim 2165 \pm 37$ years BP to $\sim 653 \pm 35$ years BP (Bordy et al., 2018). The debris flow facies described are dominated by coarse material mixtures that have more in common with the uppermost portion of the Prinspoort record (unit h; Figure 8(a)) than with the remainder of the sand-dominated units in the terrace. This is likely a function of differences in sediment sources, with the Huis River site being located entirely within a mountain catchment underlain by Table Mountain Group quartz arenites (including a large portion of Nardouw Subgroup

rocks), and the Prins River site being located within a mountain to foothill transition within the Nardouw Subgroup but with a greater diversity of lithologies in the upstream catchment.

As cautioned by Bordy et al. (2018), fire-palaeoflood associations are difficult to constrain because charcoal can be stored within a catchment for some time preceding a large rainfall event. However, the chance of occurrence of large rainfall events immediately following a severe wildfire has the potential to mobilize a large volume of sediment (M. C. Grenfell et al., 2022; Morán-Ordóñez et al., 2020). The Huis and Prins records together support current understanding that fire return intervals are shorter in Fynbos (Huis) than in Succulent Karoo (Prins) environments (Forsyth et al., 2010; Kraaij & van Wilgen, 2014; Van der Merwe et al., 2016), with charcoal present in most of the sediments at the Huis River site, but in only one of the units at Prinspoort (unit h, associated with the large flood that breached valley fill and exposed the terrace). The sand-dominated units at Prinspoort are mostly moderately-well sorted, suggesting accumulation following in-channel fluvial transport, in contrast with the debris flows at Huis River (and Prinspoort unit h), which reflect activation of a range of catchment surface sources.

The second comparative chronology is a set of floodplain fill ages from the Gouritz River catchment (including sites on the lower Touws River and the Groot River, ~30–60 km east-southeast of Prinspoort; Figure 1, Damm & Hagedorn, 2010). These are largely trunk river floodplain environments, some predominantly single-thread, others dual-thread wandering, depending on the degree of confinement (as is the case for the Touws River; M. C. Grenfell et al., 2021). Damm and Hagedorn (2010) identify and describe two phases of floodplain alluviation in this region, the first occurring between $\sim 1215 \pm 25$ years BP and $\sim 875 \pm 25$ years BP comprising a sequence of sand, silt, clay and organic inclusions above basal gravel, and the second between 670 ± 50 years BP and the last century comprising homogenous fine sand. The authors suggest that the first phase of floodplain sedimentation is a response to degradation of the natural vegetation cover due to livestock farming by Khoekhoen herders, who arrived in this region ~1600 years ago with fat-tailed sheep (Deacon, 1995).

We disagree with this suggestion, as there is now broad consensus that the impact of Khoekhoen herders on landscape processes was local and temporary, rather than at the landscape level that would be necessary to support the assumptions of Damm and Hagedorn (2010) (Milton & Dean, 2021; Morris, 2018; Walker et al., 2018). The following passage from Walker et al. (2018, p. 162) articulates this well:

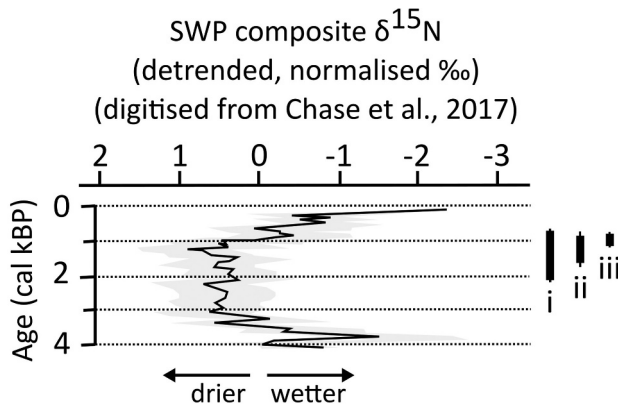
‘While it would be a mistake to think of the precolonial era as a time of stasis, the rate of social and ecological change was slow. Before the colonial period the natural system of the Karoo was exposed to a very low density of usage by herbivores, except at very small spatial scales ... Hunter-gatherer and herder societies were very small in scale and mobile, meeting the vicissitudes of a highly variable climate by moving over vast areas’.

The Karoo is a harsh environment (the name derived from Khoekhoen languages and meaning ‘dry or hard’; Henschel et al., 2018). Predators of fat-tailed sheep are prolific (Walker et al., 2018, including at Prinspoort; Hargey, 2019), surface water is scarce, and river pools become saline due to evaporation during regular and prolonged no-flow periods (M. C. Grenfell et al., 2021) while *potable* groundwater discharge is associated only with Table Mountain Group lithologies (Le Maitre et al., 2009).

In the Prins River catchment, settlement sites would have been dictated by the availability of a stable supply of potable water (such as the discharge of relatively fresh groundwater at Prinspoort), with grazing taking place in the nearest available rangeland. This would have been the riparian zone or the areas of Renosterveld associated with fine-grained nutrient-rich sedimentary rocks of the Bokkeveld Group (Klipbakkop Formation, Ceres Subgroup; [Figure 1](#)). These areas collectively comprise only ~30% of the catchment area (Renosterveld ~25%, Muscadel Riviere ~5%) and have low to moderately low connectivity with the drainage network ([Figure 4](#)) (with the exception of the riparian zone itself, where the dominant and most morphodynamically relevant natural vegetation is *Vachellia karroo* riverine thicket, distributed in association with phreatic potential; M. C. Grenfell & Dube, 2022; M. C. Grenfell et al., 2021). The remainder of the catchment comprises Fynbos (~45%) or Succulent Karoo (~25%) vegetation with low grazing potential. Lithology and associated edaphic factors exert such prominent control on the distribution of vegetation types in this region (Mucina & Rutherford, 2006) that it is unlikely that the climatic variation of the Holocene would have shifted vegetation-type boundaries much beyond the lithological footprint. Based on this crude conceptual ‘palaeoscape model’ (*sensu* Fisher et al., 2010) for Khoekhoen occupation and the understanding that herders would have migrated in response to changes in grazing resource availability, we argue in support of the current consensus that the alluviation at Prinspoort cannot be considered a response to catchment disturbance by Khoekhoen occupation. Neither is it a response to European settlement, as it pre-dates this by several centuries.

The Prinspoort terrace preserves a chronology of tributary valley alluviation covering the period $\sim 1640 \pm 250$ years BP to $\sim 840 \pm 120$ years BP that is regionally synchronous (within a distance of ~ 100 km overall) with valley alluviation in the upper Huis River ($\sim 2165 \pm 37$ BP to $\sim 653 \pm 35$ BP) and floodplain alluviation in the lower Touws and Groot rivers ($\sim 1215 \pm 25$ BP and $\sim 875 \pm 25$ BP), especially with respect to the timing of cessation of alluviation ([Figure 9](#)). Differences in the definition of ‘present’ in the ages BP reported in Bordy et al. (2018), Damm and Hagedorn (2010) and this study are less than 20 years and thus well within the measures of dating error. Having ruled out any impact of Khoekhoen settlement, we now evaluate possible effects of recent changes in rainfall. We restrict our interpretation to the work of Chase et al. (2017) from Seweweekspoort because this site is within the Little Karoo, only ~ 50 km to the east-northeast of Prinspoort ([Figure 1](#)), and yields relatively high-resolution insight into climatic variability over the past 4 ka (the period preceding and including the chronologies discussed above).

The aseasonal rainfall zone is an area of transition affected by circulation types characteristic of both the winter (westerly flow, stratiform) and summer (easterly flow, convective) zones. The relative influence of these circulation types varies over glacial-interglacial cycles, with westerly flow becoming more pronounced during glacial periods, and easterly flow becoming more pronounced during interglacial periods (Braun et al., 2019; Chase & Meadows, 2007). However, insolation-driven warming within glacial periods results in more convective events driven by an increase in tropical-temperate troughs, whereas warming within interglacial periods does not, because these tropical-temperate influences are disrupted by the poleward displacement of westerly flow (Braun et al., 2019; Chase et al., 2017). At 4 ka, after the mid-Holocene thermal optimum,



- i) Upper Huis River valley fill (Bordy et al., 2018)
- ii) Prinspoort valley fill (this study)
- iii) Touws/Groot river floodplains (Damm and Hagedorn, 2010)

Figure 9. Contextualisation of regional chronologies of valley or floodplain alluviation in relation to the seweweekspoort rock hyrax midden $\delta^{15}\text{N}$ record from Chase et al. (2017). Vertical bars (i to iii) show the approximate time of initiation and cessation of sedimentation for each chronology discussed in the text (with line markers indicating approximate error bounds).

conditions at Seweweekspoort became increasingly arid in phase with conditions in the central and southern summer rainfall zone, due to a reduction in the generation and influence of tropical-temperate troughs (Figure 9; Chase et al., 2017). However, this was followed by an abrupt shift to wetter conditions within the last millennium (Figure 9; Chase et al., 2017). Could this shift from relative aridity to relative humidity (and higher-energy rain-bearing circulation types) have initiated widespread re-working of alluvial fills and the breaching of geomorphological buffers (*sensu* K. A. Fryirs, Brierley, Preston, & Kasai, 2007)?

While there is a large amount of evidence supporting a link between increasing aridity (and associated rainfall variability) and trunk channel breakdown (e.g., S. E. Grenfell et al., 2014; Larkin et al., 2020), the implications of increasing aridity for tributary-trunk geomorphic connectivity are poorly constrained. There is no evidence of channel breakdown within the relatively narrow, rapid-rise flood driven wandering reaches of the Touws River at the field site – the river maintains a dual-thread planform where the valley widens sufficiently for bifurcation driven by deposition or excision of the floodplain (M. C. Grenfell et al., 2021). One possible alternative explanation for the tributary impoundment being a product of recent climatic forcing would be that there is an inherent limit on sediment preservation potential in valleys and floodplains of the Little Karoo, or in narrow valley settings more generally. The South African landscape in general is net-degradational over the long term (McCarthy & Hancox, 2000; Tooth et al., 2004), and the Touws system lacks the geological local base levels prevalent in rivers draining Karoo Supergroup lithologies to the northeast and east (M. C. Grenfell et al., 2008; Tooth et al., 2004). As such, there is neither the vertical accommodation space to enable sustained vertical aggradation at the field site, nor the valley width available to

accommodate sustained lateral accretion. A wandering dynamic dominates in the coarser floodplain fill (relative to floodplains from the Karoo Supergroup), characterized by frequent channel realignment and re-occupation of former branch locations. Only local and temporary alluviation is possible. This may be aided by some form of sedimentological impoundment (e.g., the interaction between tributary fan and trunk floodplain development, which itself is temporary), or may simply be a feature of channel realignment and associated floodplain development that defines the natural dynamic and sediment exchange flux pattern of wandering river floodplains (S. Grenfell & Grenfell, 2021).

It is possible that the regional floodplain cycling time in the Little Karoo is in the order of one to two thousand years, and that this has a knock-on effect across the wider catchment drainage network, thereby also influencing the potential residence time of sediments in tributary-trunk confluence zones such as at Prinspoort. Intriguingly, hundreds of kilometres to the east at confined floodplain sites in the central and southern summer rainfall zone (Highveld and Drakensberg Foothills), valley fill cycling times through scour, infill and minor lateral migration are on the order of one to two and a half thousand years (Keen-Zebert et al., 2013). This is an area of much higher rainfall and lower climatic variability (S. E. Grenfell et al., 2014), yet there is an overlap in preservation potential that must reflect the control of valley form (especially width) on fluvial processes (Walsh et al., 2023).

Walsh et al. (2023) use the chronologies of Bordy et al. (2018) and Damm and Hagedorn (2010) as part of an excellent regional synthesis of fluvial responses to late Quaternary climatic variability across western and central southern Africa. However, for insights from the last millennium, sites with relatively long chronologies to the far north and west of the Little Karoo were lumped with these sites from the aseasonal Little Karoo with much shorter records, to infer recent regional changes in flood regime (specifically, a widespread late Holocene increase in flash floods). While the sites with longer chronologies may reflect a recent change to flash flood regimes within an arid climate (Walsh et al., 2023) (with longer duration flows characterizing the wetter climate of the last glacial maximum in the winter rainfall zone; Braun et al., 2019; Chase & Meadows, 2007), the terrain, rain shadow setting and interplay between westerly and easterly circulation types of the Little Karoo (Chase et al., 2017) caution that rapid-rise flood events may always have characterized the hydrology of rivers like the Prins and Touws. To fully discern a shift in fluvial responses to climatic variability in the Little Karoo over the Holocene, we need longer archives, which may require investigations of low connectivity basins in the upper catchment where preservation potential may be higher (Figures 1, 4).

6. Conclusion

This research offers a path towards reducing the number of competing explanations for valley alluviation and exploring the relative effects and possible interactions of different drivers of environmental change. In the previous research at the basin-scale, controls on sediment delivery have been grouped according to geological factors (e.g., lithology, terrain, tectonics), geographic factors (e.g., climate) and human impacts (e.g., accelerated erosion and impoundment), with geological factors accounting for ~60% of the spatial variation in river sediment delivery to the coastal ocean (Syvitski & Milliman, 2007).

Other controls can assume greater importance at smaller spatial scales, but ignoring geological templates in favour of convenient explanations will lead to erroneous assumptions of impact. Geomorphic (dis)connectivity mediates the catchment response to environmental change by integrating the effects of geology, geography and humans on sediment dispersal and facilitates interpretations of co-impact that are required as we enter the Anthropocene (Gong et al., 2023). In this respect, connectivity provides a critical context (Wohl, 2017, 2018) for interpretations of the fluvial response to environmental change, and the integration of connectivity modelling with geochronology has a lot to offer the field.

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Disclosure statement

No potential conflict of interest was reported by the author(s).

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Author contributions

MG and SE conceived and designed the study. MG collected and analysed the field data, drafted Figures 1–6, Figures 8 and 9 and wrote the first draft of the paper. ME provided guidance on OSL sampling, completed the OSL dating, produced Figure 7 and provided feedback on the first draft of the paper. SE also provided feedback on the first draft of the paper.

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