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Initiation and recession of the fluvial knickpoints of the Island of Tahiti (French Polynesia)

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ABSTRACT

In this paper, we study the origin and evolution of the 42 knickpoints spanning the Island of Tahiti, a large extinct shield volcano in the South Pacific Ocean (French Polynesia), by combining DEM analysis and numerical modeling. These knickpoints are located along rivers (107 in total) with a total length exceeding 6 km and with a total drainage of >3 km^2. The knickpoint locations do not correspond to lithology, tributary confluence, or uplift. We argue that these knickpoints have been initiated by a sudden sea level drop of 135 m 20 ky ago, and that the littoral cliffs circling two-thirds of Tahiti are the result of marine erosion that took place 7 ky ago from a stand level that was 5 m higher than now. The head-to-toe height of the knickpoints increases with respect to the knickpoints' distance from the ocean. The major process controlling the knickpoints is plunge-pool incision and the $n=2$ stream-power model works well for modeling the profile form. The mean retreat rate of the knickpoints corresponds very well with a drainage-area dependant model with velocities ranging from 0.17 to 1.2 m/y.

1. Introduction

Initiation and recession of the knickpoints are significant boundary conditions for fluvial system processes. Knickpoints in bedrock rivers are commonly interpreted as the result of a base-level change (uplift of the superior block, or collapse of the inferior block), sea level fall, along-stream differences in lithology, sudden landslides or differential erosion between a main stream and its affluent (Howard and Kerby, 1983; Whipple and Tucker, 1999; Bishop et al., 2005; Lamb et al., 2006; Wobus et al., 2006; Loget and Van Den Driessche, 2009; Haviv et al., 2010). The presence and the height of a knickpoint in a mountain river can potentially provide a way to identify and quantify a renewed uplift or a base-level fall (e.g., Whipple and Tucker, 2002; Farías et al., 2008; Wobus et al., 2008). The migration rates and style of knickpoints have been investigated using field observations (Weissel and Seidl, 1998; Hayakawa and Matsukura, 2003), physical modeling (Gardner, 1983), and simple numerical modeling for the longitudinal river profile perturbations (Seidl and Dietrich, 1992; Whipple and Tucker, 1999). The style of knickpoint migration could be parallel retreat (the knickpoint keeps its form) or replacement retreat (the knickpoint dies out by backwards rotation; Gardner, 1983). The migration rate of knickpoints is a first approximation of the rate at which a network can return to equilibrium after the triggering event (Loget and Van Den Driessche, 2009). Some field and experimental studies have shown that a knickpoint migration rate in the fluvial regime is related to its drainage area, which can be considered as a water discharge proxy (Rosenbloom and Anderson, 1994; Bishop et al., 2005; Crosby and Whipple, 2006). Others have focused on the sediment transport and rock resistance (Sklar and Dietrich, 1998; Hayakawa and Matsukura, 2003; Snyder et al., 2003). Through numerical modeling studies, we can have a better understanding on how the riverbed adjusts to the changes of boundary conditions, such as the efficiency of erosion, the efficiency of sediment transport, uplift or sudden sea level drop (Seidl and Dietrich, 1992; Whipple and Tucker, 1999). However, uncertainties remain on the processes able to generate a knickpoint and on the model(s) that predict(s) its propagation velocity and shape evolution (Gasparini and Brandon, 2011). In this paper, we first describe the geology and hydrology of Tahiti. Then we explain the generation and evolution of its knickpoints, using geomorphological analysis and a numerical modeling of the time evolution of the riverbed profiles based on the aforementioned works of Seidl and Dietrich, and Whipple and Tucker (stream-power law).

2. Field area

2.1. Geological setting

The Tahiti Island lies in the tropical South Pacific Ocean (17°30′S–17°50′S, 149°10′W–149°40′W), and was created around 1.4 My ago by a basaltic outpouring above an intraplate hotspot (Le Roy, 1994; Hildenbrand and Gillot, 2006). It is divided into two geological units connected by an isthmus: the main island Tahiti-Nui to the northwest and the subsidiary island Tahiti-Iti to the southeast. Tahiti-Nui is cut

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near its center by a N110 rift zone. The southern and northern flanks of Tahiti-Nui slipped between 0.87 and 0.85 My ago (Clouard et al., 2001; Hildenbrand et al., 2004). These giant, lateral collapse-type landslides, caused by listric faults, occurred on the northern and southern flanks, with headwalls located near the summit of the first shield volcano (Fig. 1). Thereafter, a secondary shield filled in the U-shaped depression (end of eruption at 0.45 My) (Clouard et al., 2001; Hildenbrand and Gillot, 2006) with basalt alkaline series lavas (Hildenbrand et al., 2004). The principal lava families are basalt and basanite. The lava flows uncovered in the valley walls are generally thin (several meters) with a slope of 8°, with no significant erosional unconformities between flows (Davis, 1918). This continuity shows that the edifice was accumulated rapidly by uniform basaltic deposition. The second shield volcano had a maximum elevation of 2900 m and a 20 km radius with an 8° mean slope (Hildenbrand et al., 2004). The radius of Tahiti-Nui (Fig. 1) is now around 15 km because of the viscoelastic subsidence of the volcanic edifice into the Earth's mantle. The highest elevation (Orohena Peak) is 2241 m. The mean shield volcano slope of 8° is still well preserved today.

2.2. Climate and hydrological setting

The tropical climate of the Tahiti Island is characterized by a warm and rainy period when northeast trade winds are predominant (from December to March), and by a fresh and dry period when southeast winds prevail (from June to September). The rainfall rate varies from 1.5 m/y for the west coast to 3.5 m for the east coast exposed to trade winds, and can be up to 8 m/y on the summit of Tahiti Nui. The normal precipitation average is 1.69 m/y (Laurent et al., 2004).

A hydrographic network of 107 rivers radiates from the Tahiti-Nui main caldera (Fig. 1), and can be split into two families: 1) catchments with knickpoints higher than 200 m and river lengths longer than 6 km; 2) catchments without knickpoints, with river lengths less than 6 km and a river gradient of at least 0.2 m/m (Fig. 2). The basin formation over the first (main) shield occurred earlier than over the second shield. The basin sizes range from less than 1 km^2 to almost 100 km^2 and are more elongated in the depression zone than those in the main shield. In addition, the rivers incised the depression zone more deeply than the main shield. The largest catchment is the Papenoo basin (river 17, 90 km^2), followed by the Punaruu basin (river 96, trade wind rain shadow) and the Papeihia basin (river 46, rainy sectors), respectively. The last two are located in the rift zone. The other minor basins are dispersed all over the island, with sizes uncorrelated with the local climate (Hildenbrand et al., 2008). Almost all of Tahiti's main valleys are spectacular amphitheater-headed valleys, often abruptly terminated by knickpoints (Servant, 1974). The knickpoint locations in Tahiti do not appear to be related to lithology, tributary confluence or uplift (Servant, 1974).

The formation of steep valley headwalls has been classically explained by seepage erosion rather than by surface runoff (Wentworth, 1928; Kochel and Piper, 1986; Hovius et al., 1997), until the waterfall erosion origin was proposed by Lamb et al. (2007). Waterfall erosion is driven by runoff erosion (incision, abrasion or attrition). Seepage erosion operates at locations where ground water exists, the weathering process operating through salt precipitation, chemical dissolution or frost growth. However, in Tahiti, the natural spring locations do not match the valley headwall locations (Hildenbrand et al., 2005). Lamb et al. (2006) note that the transport capacity and incision capacity of spring water acting on bedrock is still uncertain. Some small-scale experiments (Howard and McLane, 1988) show that the sand and fine sediments can be removed by seepage, but that only high runoff discharge could remove all sizes of sediments (Montgomery and Buffington, 1997).

2.3. Base sea level history

The sea level was not constant after the end of island construction 200 ky ago and has undergone large variations thereafter. At the end of island construction, the sea level was at about the same height as today. It dropped 135 m 140 ky ago, and rose again by the same amount 125 ky ago. Then it fell by 135 m at the climax of the last ice age 20 ky ago. It was 5 m higher than today 7 ky ago (Pirazzoli and Montaggioni, 1989; Bonvallot et al., 1993; Fletcher et al., 1995; Lambeck and Chappell, 2001; Peltier, 2002; Heindel et al., 2008; Thomas et al., 2011). Accordingly, Servant (1974) argues that all of the valleys in Tahiti were carved around 30 ky ago, about 100 m below their present-day elevations with respect to sea level. In addition to the sea level changes, the island is slowly subsiding into the mantle at a rate of 0.5 mm/y (Hildenbrand et al., 2004; Fadil et al., 2011).

Two-thirds of Tahiti is circled by littoral plains with a width of less than 2 km (Figs. 1 and 2), covered with sediment deposits and delineated by littoral cliffs, mostly along the western and southern sides (Davis, 1918). We observe littoral cliffs near all of the river mouths. Lava dating (Hildenbrand et al., 2004) indicates that the western and eastern sectors (1.37–0.87 My) are older than the northern and southern sectors (0.85–0.46 My). Fig. 2 shows that the basin edges were cut by littoral cliffs near the ocean side, and Fig. 7 plots the elevations of the littoral cliffs, which range from 60 m to 600 m.

The soil composition and the presence of relict sea caves seem to indicate that these littoral cliffs (e.g. the Maraa Cave which is 60–70 m long and as wide, 5 m above sea level, and which is located at 30 km from Papeete; Davis, 1918) have been created by sea erosion during an episode in which the sea level was 5 m higher than now (Pirazzoli and Montaggioni, 1989; Fletcher et al., 1995; Lambeck and Chappell, 2001; Peltier, 2002; Thomas et al., 2011). On the southwestern and southern sides (rain shadow), the littoral cliffs (200–600 m) are higher than on the other sides (60–150 m). This fact could be explained by the Tahiti subsidence bascule theory (Servant, 1974) which states that the northeastern and eastern sides of Tahiti ramp down faster than the other sides (due to the absence of coral reefs on these sides), or by a regional landslide along the southwest coast near the outlet of river 90 (evidence from bathymetric data; Clouard et al., 2001; Hildenbrand et al., 2004).

3. Observed knickpoint distribution

We extracted the river and ridge longitudinal profiles from a 5 m resolution DEM, provided by the “Service de l'Urbanisme de Tahiti” and map no. 513 of the “Institut Géographique National” (1988). We identified 107 rivers, 30 amphitheater-headed valleys (100–1400 m high knickpoints), 12 intermediate valleys (knickpoints at mid-river paths), 42 knickpoints (river gradient above 0.6 m/m), and contoured the littoral cliffs and littoral plains (northeastern and southern Tahiti, see Fig. 1). Because direct river profiling is difficult in Tahiti (rugged terrain, steep slopes), only 5 knickpoints were visited.

The knickpoints (Fig. 1) are located along rivers with a total length (X_{total} distance in Fig. 3) of more than 6 km. Typical river longitudinal profiles with their edge profiles are shown in Fig. 2. Rivers 3, 10, 17, 21, 22, 29 and 30 are on the northern side; rivers 38 and 47 are on the eastern side; rivers 61, 67, 73 and 74 are on the southern side; and rivers 82, 96 and 99 are located on the western side. Rivers 3, 10, 17, 30, 61, 67, 73, 74 and 96 present concave profiles and their knickpoints show a steep valley headwall near the divide. Rivers 22, 38, 47, 82 and 99 have knickpoints at mid-river paths and exhibit an intermediate morphology with smaller basin sizes (e.g. the Vaipuu (22), Mahatearo (38), Utufai (47), Tereia (82), and Matatia (99) valleys; see Fig. 1). Rivers 21, 29 and 94 present small drainage areas without knickpoints or a small knickpoint (height less than 100 m), but with a slope of around 0.2 m/m over the entire river length. A good correlation