5 Infiltration losses from the Nylsvlei floodplain

5.1 Introduction

Infiltration of floodwaters into the floodplain soils is potentially another significant loss to floodwaters inundating the Nylsvlei floodplain. The other potentially significant loss to floodwaters is to evapotranspiration, and this is dealt with in Chapter 4. Like evapotranspiration, infiltration could influence the hydraulic behaviour of the floodplain and therefore needs to be included in the model. This chapter is about investigating and attempting to quantify the losses of floodwaters to infiltration on the Nylsvlei floodplain.

Infiltration is the term applied to the process of water entry into soil, generally by downward flow of water through all or part of the soil surface (Hillel, 1980). The infiltration rate is defined as the volume flux of water flowing into the soil profile per unit of soil surface area (Hillel, 1980) and is generally expressed as a depth of water per unit time, such as mm/hour. The infiltration rate into the soil depends on the rate that water is applied to the soil surface. If this rate is less than the infiltration capacity of the soil, then water will enter the soil at the rate it is applied and no runoff will occur. On the other hand, if the rate of application is greater than the infiltration capacity of the soil, then the water will enter the soil at the infiltration capacity and the balance will run off. In most cases of interest, water is applied to the soil surface as rain but on the Nylsvlei floodplain, it is applied through flooding. When the floodplain is inundated, water infiltrates at a rate equal to the infiltration capacity of the soil. The infiltration capacity of a soil tends to decrease with time, from the time of first application of water to the soil surface, tending to a constant rate called the final infiltration capacity or the steady-state infiltrability (Hillel, 1980). The decrease in infiltrability is due to many reasons (Hillel, 1980) such as a gradual deterioration in the soil structure, migration of pore blocking particles, swelling of clay, entrapment of air bubbles or the bulk compression of soil air if it is prevented from escaping as it is
displaced by the infiltrating water. The main reason is due to a decrease in the matric suction gradient, which occurs as infiltration proceeds.

The soil type of an area affects the infiltration rate. The soils at the Nylsvley Reserve were mapped by Harmse (1977) (Figure 5.7) using the Binomial Classification System, which consists of soil forms and series. The soil form is an organised combination of soil horizons with the grouping of specific kinds of diagnostic horizons in specific sequences. The soil series constitutes a subdivision of the border soil form to enable a more specific description of soils according to criteria that influence its physical responses, such as texture and degree of leaching (Schulze, 1997). In the Nylsvley Nature Reserve, there are 17 different soil forms classified according to the South African National Soil Classification System. Within this are 34 different soil series differentiated in terms of colour, clay content, base status and degree of calcareousness. Frost (1987) gave a simplified soil map of the Nylsvley Reserve shown in Figure 5.1, based on the map drawn by Harmse (1977) shown in Figure 5.7.

On the Nylsvlei floodplain, the soils are generally stratified alluvium deposited during regular floods, overlying bedrock of various sedimentary and igneous rocks (Figure 5.2) (Morgan, 1996). These alluvial sediment strata (Figure 5.1 and Figure 5.7) comprise alternating bands of clay-rich and sandy material, generally not continuous in the horizontal and vertical directions, indicating different types of sediment carried by sheet flow of floodwater (Morgan, 1996) that probably reflect long-term shifts in climate (Frost, 1987). The soils of the Arcadia, Bonheim and Inhoek forms are found here (Harmse, 1977).

Eluvial horizons (soils deposited by wind) (Figure 5.1) are present in some of the clay-rich soils of the bottomlands, such as at the edges of depressions and drainage channels alongside the Nyl River (Frost, 1987). In these soils, vertical drainage is impeded by the presence of a cutanic B-horizon formed by the illuviation of clay particles from the upper profile and their deposition on the surfaces of the underlying material.
Figure 5.1: Simplified soil map of the Nylsvley Nature Reserve (after Frost, 1987)

Illuviation (soils deposited on a lower soil layer from an upper soil layer) (Figure 5.1) is the main process involved in the consolidation of sediments occurring in
the Nyl River valley (Frost, 1987). These sediments are derived partly from colluvium (soils deposited at the bottom of a slope by gravity) and partly from alluvium deposited during periodic floods. Scouring, sorting and the redistribution of this material during subsequent floods has created a mosaic of different-textured substrates on these bottomlands, with the coarser sediments generally being found along the river banks and the finer fractions occurring in ponded zones away from the river.

Various researchers such as Morgan (1996) and Tooth et al (2002) have held the opinion that the clay alluvium effectively seals the floodplain surface from the aquifer underneath and groundwater supply appears to play only a minor role in inundation of the floodplain. Tooth et al (2002) theorised that groundwater recharge to the aquifers underlying the floodplain occurs via water infiltrating through the gravely sand riverbeds upstream of the floodplain and through bedrock fractures in the headwaters. This would account for the general decrease in size of river channels that occurs as they approach the floodplain margin. The decrease in channel size however, could also be explained by water being stored in off channel areas and being lost to infiltration and evapotranspiration.

Morgan (1996) conducted an in-depth literature review of previous studies into the geohydrology of the Nylsvlei region. He found from previous studies (Porszasz, 1978; Higgins & Rogers, 1993; Foster, 1983), that there are generally two separate aquifers underlying the floodplain: a perched aquifer in the pore spaces of the alluvium and a rock aquifer in the joints, faults and cavities of the rocks underlying the alluvial sediments.

A perched water table separated by clay lenses was observed by Higgins and Rogers (1993) in areas overlying the Waterberg sandstones (the Middelfontein Reach of the study area, shown in Figure 5.2) in the southern area of the floodplain. Here the alluvium can be up to 24m deep and groundwater in the alluvium was found to respond directly to flooding (Morgan, 1996). In the southern region of the Nylsvley Reserve a deep, black, clay soil called Turfveld
occurs (soil map unit Ar1/2 in Figure 5.7 and “vertisols and mollisols” in Figure 5.1), consisting of the erosion products of Karoo basalts located 20km to the southeast. The Turfveld is dominated by smectite and exhibits the classical ‘melon-hole’ *gigai microrelief* typical of vertisols that are highly expansive when wet and contractile when dry (Scholes & Walker, 1993). This region inundates less often and appears to inundate through a combination of back flooding from the main floodplain to the north and water entering from the Eersbewoondspruit (Birkhead *et al*, 2004) as the Blindefontein).

![Figure 5.2: Geology of the Nylosvei floodplain region (after Morgan, 1996)](image-url)
Two chemically distinguishable aquifers were observed by Walton (1989) within the alluvium and bedrock in areas overlying the Rooiberg volcanics and Karoo sandstones (parts of the Nylsvley Reserve and Vogelfontein - Mosdene reaches of the study area, shown in Figure 5.2). Isotope studies suggested that the alluvial aquifer is recharged quicker and more often than the underlying bedrock. However, Walton (1989) speculated that surface water needed to inundate these soils for long periods to adequately recharge the aquifer, due to their low permeability. Groundwater levels at the Nylsvley Reserve remain relatively constant despite inundation and rainfall influences (Morgan, 1996), and Walton (1989) concluded that the confined alluvial aquifer in this region is contained in a nearly static and saturated state. Artesian conditions have been observed within the alluvial aquifer in the northern area of the Nylsvley Reserve, but these conditions are rare and are not reported to have occurred elsewhere on the floodplain (Morgan, 1996). Blight measured a capillary upflow of 289mm in five months on the floodplain in the Nylsvley Nature Reserve, during 2000 and 2001 leading him to think that groundwater flow to the surface could be significant.

The northern section of the Vogelfontein-Mosdene reach overlies Karoo basalts, shown in Figure 5.2. Alluvial deposits are up to 30m deep and generally graded from coarse sand and gravel at the bottom to fine sand and silt near the top (Morgan, 1996) with sediments also becoming progressively more clayey as one approaches the river channel. Groundwater was found to occur in the alluvium and the decomposed basalt at a depth of 10 to 20m in this region (Morgan, 1996). Various authors such as Porszasz and Bredenkamp (1973), Scott and Wijers (1992) and Morgan (1996) concluded that very little recharge takes place in this region due to the presence of montmorillonitic clays and that recharge would come from heavy downpours, taking place through expansion and contraction cracks in the dry soil. The cracks would then rapidly close as the clay swelled, preventing floodwaters from infiltrating easily. Massive cracks of the order of 20-50mm wide and possibly several metres deep were observed by Morgan (1996) in the dry winter months of 1993 to 1995 for example. Blight (1997a) observed that in this sort of soil (but this observation was not specifically for Nylsvlei),
uninterrupted open vertical cracks usually persist only to about one metre deep and the greatest depth to which open vertical cracks can be discerned is about 1.5m. These cracks are often not continuous but terminate in an inclined fissure at one end or the other; no inclined fissures exist at a depth of less than about 0.5m. The southern floodplain areas were not observed to have the same crack phenomenon by Morgan (1996) and this was thought to be due to “the southern floodplain areas receiving more frequent surges of floodwater which appear to maintain the moisture state of these soils at a fairly consistent level”.

Morgan (1996) conducted a study of borehole records on the floodplain between 1970 and 1994, and found that only one borehole had records of groundwater rising to the surface: the southern-most borehole on the farm Zandfontein in the Middelfontein reach of the study area. The water table here only rose to the surface during the extremely large flood events of the 1970s. Nevertheless, groundwater levels in boreholes in these southern regions of the floodplain (as far north as Vogelfontein) remained less than five metres below the surface throughout the study period. Morgan (1996) therefore deduced that complete recharge of the alluvial aquifer is not a necessary precursor to flooding. However, water levels in these boreholes did respond to flooding showing that the alluvial aquifer is recharged during a flood.

5.2 Methods of measuring infiltration

5.2.1 Ponded infiltrometers

The ponded infiltrometer can be used to simulate infiltration under ponded conditions and can vary in size from 0.1m² to several square meters (Scholes and Savage, 1989). The water level in the pond is either maintained at a constant level (constant head type) or permitted to drop (falling head type). Double ring infiltrometers (which were used on the Nylsvlei floodplain) consist of two rings one inside the other, and water is maintained at the same level in both rings. The infiltration rate is determined from the rate of drop in the inside ring, the outside ring is used to prevent lateral flow from the inside ring so that the vertical infiltration rate can be measured. Single ring infiltrometers (which were also used
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on the Nylsvlei floodplain) are also available but a correction is required for lateral flow (Scholes and Savage, 1989).

Infiltration tests conducted \textit{in-situ} can give misleading results as the limiting infiltration rate depends on not only the slope and state of compaction, but also moisture content of the material underlying the surface (Blight, 1997a). Blight (1997a) conducted double-ring infiltrometer tests on initially dry ground and on ground that had been thoroughly wet by sprinklers before the test began, both on exactly the same soil. The limiting infiltration rates still differed markedly after two weeks (10mm/h for the wet surface versus 30mm/h for the dry surface) due to augmentation of the infiltration rate from significant suction gradients for the initially dry soil (suction gradients can be significant as these tests affect only a small volume of soil). He also attributed the discrepancy to possible closing of cracks and fissures in the soil during the test (and before the test for the initially wet test area).

\textbf{5.2.2 The Guelph Permeameter}

The Guelph Permeameter (pictured in Figure 5.3) produced by Soilmoisture Equipment Corp. (http://www.soilmoisture.com; Soilmoisture Equipment Corp., 1987), is a constant-head device based on steady water infiltration into a 6cm diameter augered well. It is a simple, relatively rapid and field-based method and can easily measure the field-saturated hydraulic conductivity expressed in cm/sec ($K_{fs}$), at various depths within the root zone (Bagarello et al, 1999). Hydraulic conductivity is the measure of the ability of a soil to conduct water under a unit hydraulic potential gradient. $K_{fs}$, or field saturated hydraulic conductivity refers to the saturated hydraulic conductivity of soil containing trapped air (Soilmoisture Equipment Corp., 1987). $K_{fs}$ is more appropriate than the truly saturated hydraulic conductivity for vadose (unsaturated) zone investigations because by definition, positive pressure heads do not persist in unsaturated conditions long enough for entrapped air to dissolve (Soilmoisture Equipment Corp., 1987). At Nylsvlei, the floodplain can be flooded for long periods (of the order of months) so entrapped air in the soil could become dissolved. Hydraulic conductivity decreases as the
soil water suction increases; this relationship is called the conductivity-pressure head relationship.

Figure 5.3: The Guelph Permeameter on the floodplain alluvium in the Nylsvley Nature Reserve, note the Nyl River channel behind

The Guelph Permeameter can also measure the Matric Flux Potential ($\Phi_m$), the measure of a soil’s ability to pull water by capillary force through a unit cross-sectional area in a unit time. Sorptivity ($S$) is also measurable using the Guelph Permeameter; this is a measure of the ability of a soil to absorb a wetting liquid. In general, the greater the value of $S$, the greater the volume of a wetting liquid that can be absorbed and the more rapidly the liquid is absorbed (Soilmoisture Equipment Corp., 1987).
When a constant well height of water is established in an augered hole in the soil using the Guelph Permeameter, a ‘bulb’ of saturated soil with specific dimensions is quickly established (Soilmoisture Equipment Corp., 1987). This ‘bulb’ is very stable and its shape depends on the type of soil, the radius of the well and the head of water in the well. The shape of the ‘bulb’ is numerically described by a factor (\(C\)) used in the calculations. Once the unique ‘bulb’ shape is established, the outflow of water from the well reaches a steady state flow rate, which can be measured. The rate of constant outflow of water together with the diameter of the well and height of water in the well are used to determine the field saturated conductivity, matric flux potential and sorptivity of the soil.

Disadvantages of the Guelph Permeameter include that it can only measure \(K_{fs}\) to a minimum rate of 0.036 mm/hour (some soils had lower hydraulic conductivities than this at the Nylsvley Reserve). It may sometimes measure the horizontal permeability of the soil, which can differ from the surface infiltration rate that occurs in a vertical direction under ponded conditions (Blight et al., 2001). The Guelph Permeameter appears to be susceptible to temperature changes, which becomes significant when measuring infiltration in low permeability soils. The Guelph Permeameter behaves like a thermometer, water moves up or down in the reservoir with decreases or increases in temperature. This can be particularly serious in the early morning, late afternoon and in partly cloudy conditions. It is therefore necessary to shade the Guelph Permeameter (in this study a beach umbrella was used) to minimise these effects. Despite this, problems were still experienced on partly cloudy, hot and windy days and in the early morning and late afternoon.

A cylindrical uncased hole is augered by hand, with a standard diameter of 6cm. The Guelph Permeameter is mounted over the hole and two infiltration measurements are taken; first with a constant water depth of 5cm maintained in the hole and then with a constant water depth of 10cm maintained in the hole. The level drop (due to infiltration) in the Guelph Permeameter reservoir is monitored for each depth and recorded, until the rate of level drop reaches a steady state.
These data at the two steady states are then used in an equation to calculate $K_{fs}$. The time required to reach steady state depends on the hole diameter, the head of water in the hole, the soil hydraulic and physical properties and the antecedent soil water content at the time of measurement (Reynolds et al., 1985).

Three equations were used to calculate $K_{fs}$: the equation given by Soilmoisture Equipment Corp. (1987), one given by Murdoch (undated) and one given by Blight (1997b) for an uncased hole. The latter two were used when the Soilmoisture Equipment Corp. (1987) method returned negative values for $K_{fs}$. According to Soilmoisture Equipment Corp. (1987), negative values are returned when there are fissures or cracks in the soil or when the soil is stratified. These are termed hydrologic discontinuities.

The equation given by Soilmoisture Equipment Corp. (1987) is an empirical, difference equation relying on the steady-state rate of fall for two depths of ponded water within the augered hole. The standard depths are 5cm and 10cm and a standard well diameter of 6cm is used for this equation. The equation used to find the $K_{fs}$ in cm/s is given below:

$$K_{fs} = (0.0041 \times (X) \times (R_2)) - (0.0054 \times (X) \times (R_1))$$  \hspace{1cm} (5.1)

where:

- $X$ is the reservoir constant given on the Guelph Permeameter, in cm$^2$
- $R$ is the steady state rate of fall of water in the reservoir in cm/s. $R_1$ is the steady state rate of fall when a ponded depth of 5cm is used in the augered hole, and $R_2$ is the steady state rate of fall when a ponded depth of 10cm is used in the augered hole.

The empirical equation given by Murdoch (undated) finds $K_{fs}$ in cm/s using either the 5cm or 10cm ponded depth steady state rate of fall and the two results can then be compared:
\[ K_{fs} = \frac{CAR}{2\pi H^2 + C\pi r^2 + \frac{2\pi H}{\alpha^*}} \]  

(5.2)

where:

- \( C \) is a dimensionless shape factor from the graph of \( C \) versus \( H/a \) given by Soilmoisture Equipment Corp. (1987)
- \( A \) is the reservoir constant (cm\(^2\))
- \( R \) is the steady state rate of fall of water in the Guelph Permeameter reservoir (cm/minute)
- \( H \) is the steady depth of water in the augered hole (cm)
- \( r \) is the radius of the well (cm)
- \( \alpha^* \) is:
  - 0.01 for compacted structureless, clayey materials – landfill cap, lacustrine and marine clays
  - 0.04 for fine grained, unstructured clay
  - 0.12 for structured soil, clays to loams and unstructured medium to fine sand
  - 0.36 for coarse-grained sand and gravel and highly structured soil with large cracks

The equation given by Blight (1997b) for an uncased hole is physically derived and he maintains that this is the best method to use (Blight, pers.comm.). The results of this method generally agreed with the Soilmoisture Equipment Corp. (1987) method (Table 5.3). The equation for an uncased hole in isotropic soil is (Figure 5.4):

\[ K = \frac{Q_c}{F + H_c} \]  

(5.3)
where:

- \( Q_c \) is the steady-state flow rate of water into the soil, calculated from the steady-state rate of fall of water in the Guelph Permeameter reservoir and the cross sectional area of the reservoirs (the reservoir constants)
- \( H_c \) is the constant depth of water in the hole above the water table, taken here as the constant depth of water in the hole from the bottom of the hole
- \( F \) is a factor found by the following equation:

\[
F = \frac{2\pi L}{\ln \left( \frac{L}{D} + \sqrt{1 + \left( \frac{L}{D} \right)^2} \right)}
\]

(5.4)

where:

- \( L \) is the depth of water in the hole
- \( D \) is the diameter of the hole

Figure 5.4: Method to determine infiltration into a soil from an open hole in uniform soil (after Blight, 1997b)

Bagarello et al (1999) investigated the use of a Guelph Permeameter in cracked clay soils. When cracked clay soils are dry, the network of shrinkage cracks can
cause the soil’s hydraulic conductivity to be similar to that of a sandy soil. The hydraulic conductivity of these types of soils has been known to change by several orders of magnitude within hours. Detecting how fast these cracks close and the initial $K_{fs}$ can be difficult. Unrepresentatively low values of $K_{fs}$ can be found in cracked clay soils as a result of smearing and/or compaction of the soil during the augering process that may destroy the soil structure along the well wall; inaccurate detection of when steady flow out of the Guelph Permeameter has been reached; and ‘artifact’ (measurement induced) swelling of the soil that may cause cracks to close. Thus the use of a Guelph Permeameter in cracked soil to find the initial $K_{fs}$ requires that an initial or ‘quasi’ steady flow be attained before crack closure can exert an appreciable influence on both the time to steady flow and the antecedent $K_{fs}$ of the soil. Cracking clay soils have been reported to occur on the Nylsvlei floodplain and only appear to allow water to infiltrate to any appreciable degree for a very short time initially (discussed in 5.1).

5.2.3 Other study methods

Bradley (1996), as part of an investigation into the hydrology of a wetland in the United Kingdom, constructed an 8m$^2$ impounded area into which was pumped water maintained at a constant head. A network of dip-wells was then installed to a depth of 0.5m or 1m below the surface and changes in subsurface pressure distribution were monitored. The hydraulic gradient across the two layers of dip-wells was proportional to the vertical hydraulic conductivity of the soil. The Kirkham formula for the seepage tube method was then used to find the hydraulic conductivity of the soil.

5.3 Previous studies of infiltration at Nylsvlei

Blight (2001) and Blight et al (2001) conducted permeability tests using a Guelph Permeameter, a double ring infiltrometer and a single ring infiltrometer at the same place on the floodplain in the Nylsvley Nature Reserve as the evapotranspiration measurements, where the floodplain is relatively narrow – only 100m wide (site 31 in Figure 5.7). The coordinates of the site are S 24.64916, E 28.69042. The area consists of hydromorphic grassland with a relatively narrow zone - 20 to 30m in width - of backflooded grassland (Blight et al, 2001).
At first, attempts were made to measure infiltration using double ring infiltrometers with an inside ring of diameter 0.6m and outside ring of diameter 1.0m (Blight et al., 2001) during December 1997, January 1998 and May 1998. Problems were experienced with leakage from the outer ring and an attempt was made to seal the rings using plaster of paris (Blight et al., 2001). The results varied from 27.3mm/hour to 0.28mm/hour and hence another method of measuring infiltration was sought.

A 0.5m diameter single ring infiltrometer was also used. This worked well sealed into the soil with plaster of paris, but problems were expected by Blight et al. (2001) in dry, cracked soil surfaces.

The Guelph Permeameter was used by Blight et al. (2001) to measure permeabilities along a line traverse of some 80m stretching from the floodplain margin to the channel, passing through the site where evaporation measurements were taken (site 31 in Figure 5.7). The first set of data collected during December 2000 (in 50mm deep augered holes with a constant water depth of 25mm) ranged from 190mm/h at the floodplain margin to an average of about 60mm/h on the floodplain. Further measurements were conducted in 2001 in 200mm deep holes with water depths of 50mm and 100mm; the results are shown in Figure 5.5. The results at chainage 100m fell within the channel and probably represent the permeability at a depth of about 150mm (Blight et al., 2001).

![Figure 5.5: Infiltration measurements along the traverse conducted in the Nylsvley Nature Reserve in 2001 (after Blight et al., 2001)](image-url)
A profile hole was also augered and infiltration measurements were conducted at increasing depths at chainage 50m, as shown in Figure 5.6. These are compared with measurements at chainage 100m (in the channel) and at the bridge at the upstream boundary of the Nyalsvley Reserve (transect 2 in Figure 5.7).

**Figure 5.6:** Depth profile of infiltration measured at the traverse compared with measurements in the channel and at the bridge (transect 2 in Figure 5.7) at the Nyalsvley Nature Reserve (after Blight et al, 2001)

In June 2001, Blight (2002a) conducted a soil transect along the traverse, the soil profile to 700mm below ground level was investigated. At the edges of the floodplain (roughly) there is a silty-sand (~50mm deep) overlying a silty-clay and a clayey silt down to about 250mm depth in turn overlying a silty-sand. Within the floodplain itself, there is a soft, silty clay of about 100mm depth overlying a stiff, silty clay to about a depth of 400mm overlying a silty-sand. There was a small artesian head in the aquifer in the sand colluvium but by September 2001 this artesian head had disappeared. Clay content increased with proximity to the channel and sand content decreased, thus it could be expected that the soil in the channel is less permeable than that further away. Permeability decreased steeply with increasing clay content and could be very small at clay contents exceeding 50% (Blight et al, 2001). Most of the samples taken exhibited clay contents exceeding 25%. Blight et al (2001) also mention that for these clay contents, permeabilities (as indicated in Figures 5.5 and 5.6) should be of the order of 0.01 to 0.001mm/h rather than 1 to 100mm/h.
Blight (pers. comm.) attributed these high infiltration rates to cracks in the clay soil. Blight *et al* (2001) attributed the decrease in permeability of the soil with depth at chainage 50m to overconsolidation of the clay by desiccation of the soil in the dry season, or the possibility of deeper soil layers having higher clay contents than the surface soil layer due to earlier deposition of more clayey sediment.

Blight (2002a) noted that during the time that he conducted this study, from 24 August 2000 to 26 September 2001, the floodplain was visibly wet for most of the time but never flooded, even to a depth of a few millimetres. From the water balance, he showed that in a drier than average year like 2000/2001 (84% of the average rainfall - 555mm), upward recharge does take place by capillary flow from a shallow groundwater table - a capillary upflow of 289mm was found to occur in five months.

Fourie and James (James pers. comm.) also conducted some Guelph Permeameter tests at the traverse site (site 31 in Figure 5.7) and at a site on Transect 2 (Figure 5.7), which are tabulated in Table 5.1.

### Table 5.1: Results of Fourie and James’s (pers. comm.) study using the Guelph Permeameter in June 2000 at the ring infiltrometer sites (site 31 in Figure 5.7) and at the Nylsvley Bridge (Transect 2 in Figure 5.7)

<table>
<thead>
<tr>
<th>Site</th>
<th>Water depth (mm)</th>
<th>Infiltration rate (mm/hour) (Equation 5.3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ring sites</td>
<td>50</td>
<td>5.6</td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>0.8</td>
</tr>
<tr>
<td>Nylsvley bridge (at last culvert opposite Nylstroom turnoff)</td>
<td>50</td>
<td>3.3</td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>1.4</td>
</tr>
<tr>
<td>Nylsvley bridge (nine fence poles from last culvert to south)</td>
<td>50</td>
<td>12.2</td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>14.4</td>
</tr>
</tbody>
</table>

### 5.4 Fieldwork at the Nylsvley Nature Reserve during 2002

Infiltration tests were conducted at the Nylsvley Nature Reserve during October and November 2002. Being at the beginning of the rainy season, these
measurements probably represented the same conditions as those just prior to a flood. The reserve was visited three times. The first visit in early October 2002 coincided with a rainy day and the following days, although sunny, had damp ground. The second visit was about two weeks later towards the end of October 2002 and no rain had fallen since before the first trip. The ground was dry and cracked in places and moisture was only evident in some holes at a depth greater than about 10cm. The third visit took place in mid November and no rain had fallen recently on the floodplain again, the only significant rain had fallen about two weeks before the visit. The soil was again dry and cracked but from a depth of about 50mm to 100mm, the soil was damp and the cracks appeared to not extend beyond these depths. The tests were conducted in a below average rainfall season and the floodplain was dry, save for some shallow water in the channel mainly in the form of stagnant pools with aquatic plant vegetation.

Tests were planned in all of the soil map units (‘karteringseenheid’) (Table 5.2 and Figure 5.7) that are prone to inundation. The soil map given in Harmse (1977) of the reserve was used to plan the test positions, as shown in Figure 5.7 with positions of the test holes (results of each test hole are given in Appendix A). Fanie Baloyi who was the resident CWE employee at the Nylsvley Reserve pointed out areas known to flood, and soil map units in these areas were tested. Tests were planned for five surveyed transects that have been previously studied in the reserve but due to the long time required to conduct each test, a full traverse was only conducted on four of the five transects with the fifth transect having only one test hole. Transects that were traversed were numbers two (at the Nylsvley Bridge), three, four, five and six, shown in Figure 5.7. The sixth transect (transect seven) at the Vogelfontein road was excluded as the soil map did not include this area and due to time constraints. Data were also collected from the southern branch of the floodplain, south of Stemmerskop where inundation occasionally occurs and in other areas on the floodplain with different soil map units prone to inundation.
A hole depth of 200mm was used for most of the sites and profile holes deeper than 200mm (to a maximum depth of 800mm - the greatest depth that could be measured) were tested in November 2002 and are shown in Table 5.3 and Appendix A.

Table 5.2: The dominant soils that make up the soil map units as shown in Figure 5.7 and tested for permeability

<table>
<thead>
<tr>
<th>Soil map unit</th>
<th>Dominant soil</th>
<th>Sub-dominant soil</th>
<th>Soil origin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alluvium</td>
<td>Swamp ('moeras' - Harmse, 1977)</td>
<td></td>
<td>Alluvial</td>
</tr>
<tr>
<td>Ar1/2</td>
<td>Floodplain ('gelykvlakte' - Harmse, 1977)</td>
<td></td>
<td>Vertisols &amp; Mollisols</td>
</tr>
<tr>
<td>C1</td>
<td>Limpopo</td>
<td>Lindley</td>
<td>Illuvial</td>
</tr>
<tr>
<td>C4</td>
<td>Dundee (clayey)</td>
<td>Cromley/Arniston</td>
<td>Illuvial</td>
</tr>
<tr>
<td>C7</td>
<td>Escourt</td>
<td>Sterkspruit</td>
<td>Eluvial</td>
</tr>
<tr>
<td>D3</td>
<td>Albany</td>
<td>Limpopo/Lindley</td>
<td>Eluvial</td>
</tr>
<tr>
<td>Ok7/4</td>
<td>Limpopo</td>
<td></td>
<td>Illuvial</td>
</tr>
<tr>
<td>Va2/3</td>
<td>Arniston</td>
<td></td>
<td>Illuvial</td>
</tr>
<tr>
<td>Va2/4</td>
<td>Lindley</td>
<td></td>
<td>Illuvial</td>
</tr>
</tbody>
</table>

There was evidence of clay layers that are less permeable interspersed with more sandy layers of higher permeability, as noted by various other authors (Blight et al., 2001; Morgan, 1996, Tooth et al., 2002). In the deeper profile holes, the soil generally became more clayey with depth. The clayey soils were usually stiff and black and the sandy soils were lighter in colour, although there were exceptions.

The results of the Guelph Permeameter tests (presented in Table 5.3 and Appendix A) reveal differences of as much as two orders of magnitude for the same sites using the three different calculation methods (equations 5.1 to 5.4), but usually the difference was an order of magnitude or less. There were similar variations in results for different holes in the same soil map unit, and between soil map units (Table 5.2 and Table 5.3). The permeabilities calculated using equations 5.3 and 5.4 (Blight, 1997b) often show a lower value for the 10cm water depth than for the 5cm water depth due to differences in soil saturation resulting from the first test. Fourie (pers. comm.) maintains that the Guelph
Permeameter is unlikely to be able to measure permeabilities to accuracies better than an order of magnitude.

The permeability of the floodplain alluvium (populated by wild rice and labelled as ‘moeras’ (swamp) in Figure 5.7) ranged from infiltration rates too low to measure with the particular Guelph Permeameter used (referred to as ‘Very Low’ in Appendix A and Table 5.3) to 70.2 mm/hour (Table 5.3 and Appendix A). However, 75% of the permeabilities calculated using all the methods for all the holes as given in Appendix A (excluding negative values or holes where there were burrows) had values less than 1.0mm/hour. The infiltration rates measured by Blight et al (2001) using the Guelph Permeameter (mostly at depths shallower than 200mm) and ring infiltrometers (on the surface) fall within the same range as the infiltration rates measured at depths of 200mm. Permeability in the deeper profile holes in the alluvium decreased with depth and generally became too low to measure at a depth of 400mm to 800mm, with the particular Guelph Permeameter used. Reasons for this include low permeability clay layers observed at greater depths, the reduced influence of roots (which enhance soil permeability) with depth, the existence of animal burrows and cracks in the soil at shallow depths. Grass roots were observed in most auger holes to depths of approximately 200mm (high permeabilities were observed at depths shallower than this) and burrows were observed in two auger holes at depths shallower than 200mm. Times to reach steady state for shallower depths were generally shorter; Bagarello et al (1999) state that steady-ponded infiltration rates are obtained quickly when flow is primarily controlled by macropores. Blight (2001b) commented that some of these high measured permeabilities may be due to the Guelph Permeameter measuring the horizontal permeability of the soil.

The permeabilities measured at 200mm depth in the alluvium expressed as a daily rate (ranging from 1.2mm/day to 1685mm/day) are higher than the evapotranspiration rates derived from energy balance measurements on the Nylsvlei floodplain (0.4mm/day to 4.7mm/day depending on the month of the
year - Table 4.5) while the infiltration rates at depths below 400mm were often lower than this.

Anecdotal evidence of low soil permeability in the floodplain alluvium also exists: during the study conducted in October and November 2002, it was noted that no puddles remained on the floodplain but in certain auger holes where water was left overnight, water was still present in the morning at a similar depth as the night before. Marneweck (pers. comm.) mentioned that when he excavated holes in the floodplain with an auger during a wet year when the water table was high, the holes took at least 12 hours to fill.

Two other soil map units (other than alluvium), C4 and Va2/3, (as recorded by Harmse (1977)) occur on the floodplain within the Nylsvley Reserve (Table 5.3 and Figure 5.7). Soil map unit C4 had one test hole, on an island where a group of acacias grow (hole 15 in Figure 5.7), and was only measured at 200mm depth. Va2/3 had permeabilities that varied more than the alluvium; the Va2/3 soils were observed to vary from silt to clay with increasing proximity to the channel. Permeabilities at depths of 400mm and 800mm were low except in hole 23 (shown in Figure 5.7) at 800mm, which had a high infiltration rate due to a layer of relatively permeable yellow silt at this depth.

Very low permeabilities were measured on the southern branch of the floodplain in the Nylsvley Reserve, near the farm Blindefontein, at depths of 200mm. This area consists almost entirely of soil map unit Ar1/2 (Harmse, 1977) (Table 5.3 and Figure 5.7) and exhibits the hummocky appearance (the so-called gilgai microrelief) of a highly expansive and contractile clay (Morgan, 1996). This area is not inundated in every flood event, and is possibly backflooded from the Nyl River (Morgan, 1996) in addition to flooding from the Eersbewoondspruit.
Figure 5.7: Soil map of the Nylovley Nature Reserve (after Harmse, 1977), with numbered sites of infiltration measurements using the Guelph Permeameter and ring infiltrometers, and numbered transects. Refer to Appendix A for the soil map unit at each site and Table 5.2 for corresponding soil type and origin.
Table 5.3: Summary of infiltration rates (mm/hour) in different soil types measured at the Nylsvley Nature Reserve, during October and November 2002 with a Guelph Permeameter

<table>
<thead>
<tr>
<th>Soil map unit</th>
<th>Hole depth</th>
<th>Method</th>
<th>Water depth</th>
<th>Number of tests</th>
<th>Minimum</th>
<th>Maximum</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>200mm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alluvium</td>
<td>200mm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>400mm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>600mm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ar1/2</td>
<td>200mm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>800mm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>1400mm</td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

* ‘Negative’ means that the Soilmoisture Equipment Corp. (1987) equation (equation 5.1) returned a negative value for the Kfs of the soil

** ‘V. Low’ denotes infiltration rates too low for the Guelph Permeameter to measure
Low permeabilities were measured in the sodic areas (which lie in soil map units C1, D3 and Va2/4 in Appendix A, Table 5.3 and Figure 5.7), generally situated at higher elevations where only large floods would reach. Each test hole fell within a different soil type (holes 11, 16 and 21 in Figure 5.7) despite the similar appearance of the soils – a yellow sandy-clay. Infiltration rates measured in these soils at 200mm depth varied from 0.14 mm/hour to 3.96 mm/hour depending on the method used to calculate the infiltration rate. Anecdotal evidence in these sodic areas suggests that infiltration rates are very low: some rainwater puddles stood on the surface of sodic areas for more than five days during the study in October 2002. McCarthy (pers. comm.) noted that a borehole drilled to a depth of approximately seven metres on a sodic island in the Nylsvlei floodplain remained dry despite the floodplain surrounding the island being inundated. The surface layer of the sodic soils appeared to have a low permeability and was very hard: the surface layer (5cm to 10cm thick) was very difficult to auger through. Some higher permeabilities were measured in soil map unit Va2/4 in areas that appeared to not be sodic (holes 5 and 8 in Figure 5.7).

Certain areas that are inundated less often due to their higher elevations (soil map units C7, Ok7/4 and Va2/4 Appendix A and Table 5.3 and Figure 5.7) had high infiltration rates. These areas may have higher permeabilities due to the presence of roots from trees and bushes that were observed to grow here in addition to the type of soil.

Aardvark holes and holes dug by other animals could provide preferential flowpaths for water to more permeable soil layers at greater depths, when the area is flooded. No aardvark holes were observed however on the floodplain alluvium near the channel - they were only observed at higher elevations and in other soil types. Warthogs may also play a role in affecting infiltration and ponding losses on the floodplain, as they tend to dig up the wild rice for their rhizomes, resulting in shallow depressions a few square metres in area free of plant growth. The depth of these depressions was observed to seldom be greater than 20cm but these
Chapter 5: Infiltration losses from the Nylsvlei floodplain

shallow depressions could form small ponds allowing water to infiltrate and evaporate over time when the floodwaters recede.

5.5 **Recommended infiltration rate for the Nylsvlei floodplain**

The Guelph Permeameter and ring infiltrometer tests conducted on the floodplain by Blight *et al* (2001) and in this study suggest that soil permeability varies with depth and to a lesser extent between soil map units. Permeability usually decreased with depth (due to the influence of roots, animal burrows, desiccation cracks, and soil layers with different permeabilities) but certain test holes revealed relatively high infiltration rates at increasing depths, implying that high and low permeability soil strata lie interlayered under the floodplain. These findings agree with those of other researchers such as Higgins and Rogers (1993), Morgan (1996) and Tooth *et al* (2002). Long-term infiltration losses to floodwaters are therefore governed by the soil layer with the lowest permeability and as various test holes yielded soil permeabilities too low to measure with the Guelph Permeameter (0.036 mm/hour or 0.864 mm/day), these infiltration losses are likely to be low compared with evapotranspiration losses (0.4mm/day to 4.7mm/day depending on the month of the year). Groundwater may contribute to floodplain inundation in certain areas of the floodplain: Blight (2002a) measured a capillary upflow of 289mm in five months at the Nylsvley Reserve. Morgan (1996), however found from an analysis of borehole records that this only occurs in a few isolated areas on the floodplain. This study did not conclusively determine infiltration losses on the Nylsvlei floodplain due to a lack of soil permeability data at depths greater than 800mm and the great variation in the data collected. The infiltration losses were therefore lumped with ponding losses (where receding floodwaters are lost to infiltration and evaporation in storage areas such as dams and depressions) and determined through a water balance using the hydraulic model with observed input data (explained in Chapter 6).
5.6 Summary

The nature of the groundwater-surface water interactions on the Nylsvlei floodplain has been studied by many researchers (Porszasz, 1978; Foster, 1983; Walton, 1989; Higgins & Rogers, 1993; Morgan, 1996; Tooth et al., 2002). The consensus has been that the clay alluvial soils (consisting of layers of relatively high and low permeability) contain an aquifer separated from the surface by these clay layers that reacts to inundation but hardly ever reaches the surface, suggesting that groundwater does not contribute significantly to inundation. Infiltration losses are also thought to be low due to these clay layers, only being high at the beginning of the rainy season when desiccation cracks in the clay provide preferential flow paths to the floodwaters. Blight (2002a) as part of this study however, measured an upflow of water to the surface in the Nylsvley Reserve. Infiltration measurements by Blight et al. (2001) and in this study suggest that soil permeability varies with depth and with soil map unit, being lower than evapotranspiration losses at certain depths. For the purposes of this model, infiltration losses were lumped with ponding losses (where receding floodwaters are lost to infiltration and evaporation in storage areas such as dams and depressions) and determined through a water balance using the hydraulic model with observed input data. The previous three chapters have dealt with the collection of data necessary to model the hydraulics of the Nylsvlei floodplain. The next chapter is about using these data with certain modelling programs to set up the model.